

**EXPLORING THE INTERACTION BETWEEN CLIMATE, HYDROLOGY,
AND IMPACT CRATERING ON MARS**

BY

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PREFACE

The word “geology” comes from the Greek, “*ge*,” meaning “earth”, and “*logia*,” meaning “study of”. As the name implies, geology is the study of the Earth. The field of *planetary geology*, then, suggests solely through etymology (also a term of Greek origin) that the study of planets is itself grounded in the study of the Earth. This is not far from reality. Planetary geology relies heavily on *comparative planetology*, the practice in which information can be derived of fundamental geologic processes by comparing two planets which experience different physical conditions. For example, if we wonder how an impact crater might form differently in a glacial environment on Mars, we may first look to how the impact *process* in non-glacial terrains operates on the Earth. If we wonder how the unique conditions of Mars affects the dynamics of groundwater in the deep subsurface, we would first look to the *fundamental principles* of groundwater hydrology on the Earth. Not coincidentally, these are two topics that will be broached in later chapters of this dissertation. Critically, understanding processes on other planetary bodies better informs our knowledge of the fundamental geologic processes at work, many of which affect our very own Earth, whose early geologic history has been obscured by some of the very geologic processes we seek to understand. Of the rocky terrestrial planets, Mars is considered to be the most “earth-like,” primarily because it is the only planet in our solar system besides the Earth with a hydrological cycle. Unlike the Earth, however, its early geologic history has not been erased by plate tectonics and fluvial erosion. In principle, this allows planetary geologists to probe nearly the entire history of the planet.

Despite potential hydrologic similarities, major differences exist between the Earth

and Mars. Mars has approximately half the radius of the Earth, with $1/10^{\text{th}}$ the mass and approximately $1/3^{\text{rd}}$ the gravity. The topography of Mars is dominated by the crustal dichotomy: a low topography northern lowlands and a high topography southern highlands. Like the Earth, the majority of Mars' surface area is dominated by basaltic crust, albeit in the absence of plate tectonics and seafloor spreading. Mars' geologic history is divided into the earliest Noachian period (~ 4.55 Ga to ~ 3.6 Ga), the Hesperian period (~ 3.6 Ga to ~ 3 Ga), and the Amazonian period (~ 3 Ga to the present). While few clues are left of the terrestrial Hadean eon (~ 4.55 to ~ 4 Ga), Mars preserves a rich geologic history in the parallel Noachian period, the time in which Mars was the most geologically active.

During the earliest parts of the Noachian period, Mars had a dynamo and an internally generated magnetic field. The largest volcanic provinces formed during this time, producing the Tharsis bulge upon which Olympus Mons towers over the martian landscape. Mars also experienced heavy impact bombardment during this period, which led to the formation of the largest impact basins, and possibly even the crustal dichotomy itself. All of the impact craters that formed during this time period have been severely modified by fluvial erosion, and some even once held lakes. Indeed, the now-dry ancient valley networks that branch across the Noachian terrains of Mars are testament to an era of widespread fluvial activity, hinting at perhaps a much different climate than at present. The exact nature of this early martian climate is the subject of much debate, but understanding the state of the atmosphere and the physical conditions under which Mars experienced such fluvial erosion offers the tantalizing prospect of uncovering clues to the Earth's Hadean eon, and indeed, perhaps the conditions which fostered the origin of life

on Earth.

By the Hesperian period (similar in timing and duration to Earth's Archaean eon; ~4 Ga to ~ 2.5 Ga) fluvial erosion, impact bombardment, and volcanic activity had largely waned, with the exception of very early widespread flood volcanism. During this period, the climate of Mars likely settled into its current hyper-arid, hypothermal state. During this time, the largest fluvial channels in the solar system formed. These outflow channels are wide areas of scoured terrain and streamlined islands that hint at fluvial processes on a colossal scale, and perhaps an active hydrologic cycle through the Hesperian. The outflow channels are considered the primary evidence for an ancient global martian groundwater system, although the exact mechanisms related to their formation remains enigmatic.

The Amazonian period is the longest period of Mars' history, and it is also the time in which the observable aspects of the hydrologic cycle shifted from liquid water to, primarily, ice. This period corresponds to the Proterozoic (~2.5 Ga to ~0.5 Ga) and Phanerozoic (~0.5 Ga to present) eons of Earth history. Throughout Mars' history, massive obliquity shifts occurred due to secular spin orbit resonances. In the Amazonian period, the geologic record of such obliquity shifts is preserved in the form of mid-latitude (and some equatorial) ice sheets, when water from the polar ice caps was mobilized as vapor and deposited in colder regions. Most of these thick surface ice sheets have long sublimed away into the thin martian atmosphere, leaving behind legions of debris-covered glacial deposits; a final death mask of countless orbitally-induced glaciations. Also characteristic of Amazonian Mars are the striking "fluidized" appearance of martian impact ejecta, which hints at substantially different target

properties compared with the Moon. These “layered ejecta craters” pockmark the Amazonian terrains of Mars, and suggest the presence of frozen water in the pore-space of the shallow martian crust. The wide variety in the morphologies of these craters points to complex relationships between the impact process and target properties, the nature of which remain under vigorous debate. Resolving this debate may provide important clues into the nature of the martian substrate and the impact cratering process itself. Moreover, understanding which of these impact craters formed during glacial periods and interacted with surface ice sheets offers the enticing prospect of a better understanding the glacial and hydrological cycles of Mars during the Amazonian.

This dissertation is primarily concerned with the study of, and interaction between climate, hydrology, and impact cratering on Mars:

(1) **Climate:** What is the composition and pressure of the early martian atmosphere?

What climate environment during early Mars history is required to produce the highly degraded nature of the impact craters (and valley networks) that formed during the Noachian period? Was early Mars characterized by a warm and wet climate, with erosion dominated by rainfall precipitation and surface runoff? Or were the Noachian terrains of Mars dominated by thick surface ice sheets that underwent intermittent melting and runoff? How are these processes reflected in the hydrology of Mars?

(2) **Hydrology:** How much groundwater and pore-ice is present in the martian crust?

Is the subsurface “earth-like” and saturated with water, or “mars-like,” and arid like the surface? How are these climate regimes and fluvial processes reflected in the geologic record of the deepest basins of Mars? What processes are responsible

for fluvial activity in the wetter Noachian period, and the later arid Hesperian and Amazonian periods? Is any of this fluvial activity related to the impact cratering process?

(3) **Impact cratering:** On Mars, impact cratering fuses climate and hydrology together. How does ejecta emplacement differ for impacts forming in surface ice, and can differing mechanics allow us to pin-point craters that formed in such deposits? Can we use the degradation state of impact craters to assess ancient Mars climate and hydrology? How might impact craters degrade under different climate (e.g., warm vs. cold) scenarios? Critically, what capacity (as a function of time) do impact events into surface ice sheets have to generate fluvial erosion? How does impact cratering shape the landscape to provide locations for water/ice accumulation, and is the hydrologic consequences from specific climate regimes be recorded in Mars' largest basins? Can impact craters be used to assess the target structure and water content of the shallow martian crust? If so, how might these results in turn inform us about ancient martian climate and hydrology?

Many of these questions are fundamental to Mars science, and we seek to address some aspect of each of these problems in this dissertation.

In Chapter 1, we critically evaluate the origins of what is perhaps the most enigmatic crater morphology on Mars: double-layered ejecta (DLE) craters. Discovered in the late 1970's with Viking imagery, the exact formation mechanism for these Amazonian-aged craters has remained elusive. We assess whether the unique ejecta morphology of these craters could be controlled by previously proposed models involving either atmospheric interactions or ice within the target, and specifically seek to provide an end-

to-end model for their formation. We first assess whether atmospheric interactions are responsible by evaluating relationships between ejecta morphology and target topography. Next we examine evidence for ice associated with the target of DLE craters to assess the specific target properties that may have played a role in their formation. We then compare the kinematic and frictional conditions of the ejecta of these craters to terrestrial landslides to evaluate the free-surface flow conditions which produce similar morphologies. This then provides information on the flow conditions of martian impact ejecta, which offers meaningful insight into the target properties and ejecta emplacement mechanics. The results of this study suggest that DLE craters on Mars are produced by impacts into thick surface ice sheets that were emplaced in the mid-latitudes during periods of higher obliquity, and the unique ejecta morphology is related to the unique condition of ejecta sliding on surface ice deposits. This highlights the interaction between orbital changes, which modify the martian climate and produce glacial ice ages, and the impact cratering process.

In Chapter 2, we evaluate whether the degraded Noachian highland craters could have been modified in a climate much different than the warm and wet climate canonically inferred. In such a cold and icy Noachian climate with a denser atmosphere, the southern highlands is predicted to act as a cold trap and accumulate thick surface ice deposits. Thus, we first review Amazonian-aged crater morphologies that are interpreted to form in surface ice sheets in order to inform Noachian crater morphologies that might be expected to form in ancient, thicker regional surface ice deposits. We then evaluate the unique modification processes that might operate in a cold and icy climate to degrade these craters, and consider the variety of characteristics of Noachian highland craters to

assess whether formation and degradation in surface ice sheets is consistent with their morphologic properties. Important degradation processes include contact melting between superposing hot ejecta and surface ice, and basal melting of surface ice sheets from an elevated geotherm due to the insulating effects of the superposing ejecta. Notably, we find that crater formation in thick surface ice sheets in the Noachian and subsequent removal of the ice (perhaps in the early Hesperian period) leads to a size-dependant crater cavity shrinkage, which might readily explain the paucity of small Noachian-aged craters observed in crater-size frequency distributions. We find that the crater morphology and predicted degradation scenarios are plausibly consistent with crater formation and modification in an icy highlands scenario. This emphasizes the importance of assessing the effects of the ancient martian climate on impact crater morphology.

In Chapter 3, we provide an end-to-end assessment of the process of impact ejecta-induced melting to provide quantitative constraints on the history and documentation of impact-related fluvial erosion on Mars. These processes include: (1) Contact melting of surface ice by superposing hot ejecta. This process has been interpreted to be active in the Hesperian and Amazonian periods of Mars, and may be important in Noachian Mars under the cold and icy background climate explored in Chapter 2. Critically, however, the relationship between crater diameter and ejecta temperature is not established, leading previous works to overestimate ejecta temperature due to the range of temperatures documented at terrestrial impact craters. We take an alternative approach, and start at first-principles to establish the relationship between crater diameter and ejecta temperature on Mars. We then use this as an input into thermal models to quantitatively

assess the contact melting process. (2) Basal melting of surface ice due to the insulating effects of superposing ejecta. This can lead to an elevation of the melting isotherm, leading to basal melting of the underlying cryosphere, and eventually the surface ice sheet itself. We evaluate the conditions and timing under which these two processes may operate on Mars, as well as the hydrologic fate of the meltwater, for comparison with geomorphologic observations. We find that contact melting is likely to have operated over the entire martian geologic history, while basal melting may have operated during times of elevated geothermal heat flux in the Noachian period. With this framework in hand, we then highlight fluviomorphologic indicators that can be used to assess whether surface ice was present at the time of crater formation. We document several candidate Noachian-aged impact craters which exhibit evidence for surface ice at the time of their formation, raising the intriguing possibility of a cold and icy background climate in the Noachian. This underscores the effects of climate on the hydrologic outcome of impact cratering on Mars.

In Chapter 4, we assess the state of groundwater and pore-ice in the martian subsurface through time in order to evaluate the broader hydrologic history of Mars. We first estimate the thickness of the ice-cemented portion of the upper martian crust using global excavation depth relationships of martian impact crater populations inferred to form in targets rich in pore-ice. We then compare these estimates with the results of thermal models, varying parameters such as heat flux, obliquity, and atmospheric pressure and surface temperature. In doing so, we can assess the conditions under which the thermal models can well-match the inferred thickness of the ice-cemented cryosphere. This allows us to determine whether groundwater is extant in the martian subsurface, or

alternatively if the martian groundwater system has long since frozen over in the death grips of planetary heat flux decay. The results of this study suggest that the martian groundwater system is supply-limited, with insufficient groundwater to fill the pore-space of the subsurface, and that the global groundwater system froze over in a more ancient period in Mars history. This model of a supply-limited cryosphere and groundwater system is a significant departure from the canonical assumption of abundant groundwater in the past and present, and so we review the major implications for the hydrologic evolution of the planet. This underlines the critical effects of surface climate (obliquity, atmospheric pressure) on the deep subsurface groundwater hydrology, and the interaction with impact craters which act as probes into the deep cryosphere/groundwater system.

In Chapter 5, we evaluate whether the enigmatic “honeycomb terrain,” dated to the Late Noachian and located in the largest impact basin in the southern highlands of Mars, could be formed through diapirism, and review the climatic implications for such an origin. While previous investigators have shown that the morphology of the honeycomb terrain is consistent with a diapiric origin, it remains unclear whether diapirism is physically plausible under these conditions, or if the diapir-forming layer was composed of salt or ice. Determining the nature of this material is critical because they each make different predictions for the early martian climate, with salt favoring a warm climate capable of sustaining an evaporating ocean/lake, while an ice diapirism scenario favors freezing of an ocean/lake or widespread glaciation. In this study, we combine thermal models with the analytic solutions of recent 2D thick plate models to assess the parameter space in which diapirism is predicted to occur for both salt and ice in the Hellas basin. Informed of the requisite salt/ice thicknesses by the models, we then assess the water

inventory required to cycle through Hellas to form the diapir-forming layers. We find that both ice and salt diapirism are physically viable, but the higher thicknesses of salt required to reproduce the observed wavelength of the honeycomb terrain requires a prohibitively large water volume. In tandem with the possibility that an ice diapirism origin may be related to either a frozen ocean/lake and/or glacial inflow, we find that an ice diapirism origin for the honeycomb terrain is the most likely scenario. The results of this study emphasize the interaction between impact cratering (the basin as a sink for the diapir-forming material), the hydrology within the basin, and the climate which leads to such hydrologic conditions.

Finally, in Chapter 6 we synthesize the results of each chapter within this dissertation. We review how the interaction between climate, hydrology, and impact cratering may offer important clues to the geologic history of Mars. Then, informed by our results, we explore avenues for future work. These “next steps” build on this research and may offer even more insight into the geologic processes operating on Mars through time.

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Chapter 1:
**Testing landslide and atmospheric-effects models for the formation
of double-layered ejecta craters on Mars**

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Abstract

Double-layered ejecta (DLE) craters are distinctive among the variety of crater morphologies observed on Mars, but the mechanism by which they form remains under debate. We assess two ejecta emplacement mechanisms: (1) atmospheric effects from ejecta-curtain-induced vortices or a base surge; (2) ballistic emplacement followed by a landslide of ejecta assisted by either surface- or pore-ice. We conduct a morphological analysis of the ejecta facies for three DLE craters which impacted into irregular pre-existing topography. We find that the unique topographic environments affected the formation of grooves and the inner facies, and thus appear to be inconsistent with an atmospheric-effects origin but are supportive of the landslide hypothesis. We distinguish between the two landslide models (lubrication by either surface- or pore-ice) by assessing relationships between DLE crater ejecta and morphologic features indicative of buried ice deposits, including sublimation pits, ring-mold craters, expanded secondary craters, and excess ejecta craters. The association of DLE craters with these features suggests that surface ice was present at the time of the impacts which formed the DLE craters. We also compare the Froude numbers of DLE crater ejecta to landslides, and find that the ejecta of DLE craters are kinematically and frictionally similar to terrestrial landslides that overran glaciers. This suggests that the grooves on DLE craters may plausibly form through the same shear/splitting mechanism as the landslides. In summary, our analysis supports the hypothesis that DLE craters form through meteoroid impacts into decameters-thick surface ice deposits (emplaced during periods of higher obliquity) followed by ejecta sliding on the ice.

1. Introduction

Martian layered-ejecta craters possess unique characteristics compared to the ballistically dominated ejecta of their lunar and mercurian counterparts: their ejecta deposits display relatively distinct boundaries rather than a gradational thickness and appear to have been fluidized upon emplacement (Carr et al., 1977). The unique ejecta morphology associated with layered-ejecta craters is typically attributed to fluidization by subsurface and/or surface volatiles (Carr et al., 1977; Mouginis-Mark, 1981, 1987; Woronow, 1981; Wohletz and Sheridan, 1983; Kuzmin et al., 1988; Costard, 1989; Barlow and Bradley, 1990; Barlow, 1994, 2005; Stewart et al., 2001, 2004; Baratoux, 2002; Barlow and Perez, 20003; Barnouin-Jha et al., 2005; Osinski, 2006; Boyce and Mouginis-Mark, 2006; Komatsu et al., 2007; Senft and Stewart, 2008; Oberbeck, 2009; Weiss and Head, 2013, 2014, 2017; Jones and Osinski, 2015; Jones, 2015) and/or emplacement influenced by atmospheric vortices generated by the advancing ejecta curtain (Schultz and Gault, 1979; Schultz, 1992; Barnouin-Jha and Schultz, 1998; Barnouin-Jha et al., 1999a, 1999b; Komatsu et al., 2007).

Double-layered ejecta (DLE) craters (Fig. 1A) are a particularly unusual subclass of the wide variety of layered-ejecta craters on Mars (Fig. 2) that include single-layered ejecta (SLE), multiple-layered ejecta (MLE), low-aspect-ratio layered ejecta (LARLE), and pedestal (Pd) craters. DLE craters (Fig. 1A) range from ~1 to ~35 km in rim-crest diameter (7.8 km on average) and exhibit two ejecta facies. The inner facies is characterized by a distinctive radial texture of longitudinal ridges and grooves (Fig. 3A), transverse fissures, an annular depression at the base of the rim (a “moat”), and enhanced annular topography at the distal edge (Fig. 4A) (Carr et al., 1977; Schultz and Gault,

1979; Boyce and Mouginis-Mark, 2006; Weiss and Head, 2013). The outer facies exhibits a higher runout distance, shorter wavelength grooves (Fig. 3B), and a more sinuous terminus characterized by a subdued rampart (Fig. 4A) (Barlow, 1994, 2006; Boyce and Mouginis-Mark, 2006; Mouginis-Mark and Baloga, 2006). It is important to note that DLE craters consist of two observable *facies* of ejecta, rather than stratigraphic layers. Barlow (2015) found that only 27% of DLE craters in the middle to high latitudes exhibit observable grooves, which was attributed to the obscuring of grooves by climate-driven processes (e.g., icy mantle deposition). DLE craters typically lack secondary craters (Boyce and Mouginis-Mark, 2006) and are located in the middle to high latitudes in both hemispheres (Fig. 2) (Barlow and Perez, 2003; Boyce and Mouginis-Mark, 2006; Weiss and Head, 2014; Li et al., 2015). Ejecta mobility (EM, the ratio of ejecta-runout distance from the rim crest over the crater radius) has been used to characterize the layered-ejecta craters (Carr et al., 1977; Mouginis-Mark, 1979; Costard, 1989; Barlow and Pollak, 2002; Barlow, 2005; Boyce et al., 2010; Robbins and Hynek, 2012; Weiss and Head, 2014; Jones and Osinski, 2015; Li et al., 2015), which typically have EM values of ~1-2 (e.g., Barlow, 2005; Li et al., 2015). DLE craters exhibit anomalously high EM values compared with other martian layered-ejecta morphologies, displaying an average EM of ~3 for the outer ejecta facies (Barlow, 2005; Weiss and Head, 2014; Li et al., 2015), and ~1.5 for the inner facies (Barlow, 2005; Li et al., 2015). DLE craters have been hypothesized to form through:

- (1) Interaction with atmospheric vortices, which scour the grooves on the inner facies and emplace the outer facies as a turbidity flow (Schultz and Gault, 1979; Schultz, 1992);
- (2) The incorporation of volatiles from within the target materials (Mouginis-Mark,

1981; Barlow and Bradley, 1990; Barlow; 1994, 2005; Barlow and Perez, 2003; Osinski, 2006; Boyce and Mouginis-Mark, 2006; Jones and Osinski, 2015; Jones, 2015) and outward gravity-driven flow of the near rim-crest ejecta to form the inner facies (Komatsu et al., 2007);

(3) The collapse and base surge of an explosion column, which scours the grooves on the inner facies. In this latter model, the outer facies is either emplaced as the eroded material (Mouginis-Mark, 1981; Boyce and Mouginis-Mark, 2006) or emplaced ballistically (Harrison et al., 2013);

(4) Some combination of the above factors (Barlow, 2005; Boyce and Mouginis-Mark, 2006; Komatsu et al., 2007);

(5) Inner ejecta facies formation through impact melt overtopping the crater rim (Osinski, 2006; Osinski et al., 2011);

(6) Variations in excavation angle resulting from impact into a buried icy layer (Senft and Stewart, 2008);

(7) Impact into a volatile-rich substrate followed by a pore-ice lubricated landslide of the near-rim crest ejecta (Wulf and Kenkmann, 2015);

(8) Impact and excavation through a surface snow and ice layer overlying a volatile-rich substrate, followed by ballistic ejecta emplacement and a surface ice-lubricated landslide off the slope (Fig. 4C) of the structurally uplifted rim-crest to form the inner facies (Weiss and Head, 2013).

In the latter two landslide scenarios for the formation of DLE inner ejecta facies (7 and 8 above), the transverse fissures (white/black arrow; Fig. 1A) are analogous to those observed on the surfaces of landslides (white/black arrows; Fig. 1B and C) (Carr et al.,

1977; Weiss and Head, 2013). Those features have long been attributed to tension in areas of steep slopes (Shreve, 1966; Stumpf et al., 2013), and to compression in the landslide front within the gentler-sloped accumulation zones, which is accommodated by lateral extension upslope (Stumpf et al., 2013, and references therein). Transverse fissures may also reflect interaction with the bedrock during sliding (Niethammer et al., 2012). Furthermore, in the two landslide scenarios (7 and 8 above), the grooves on the DLE inner facies are analogous to longitudinal grooves (also referred to as “flowbands”) formed on the surfaces of terrestrial landslides (Dufresne and Davies, 2009), particularly those that slide on snow and ice (Fig. 1B and C) (Shreve, 1966; Dufresne and Davies, 2009; Shugar and Clague, 2011; De Blasio, 2014).

In this contribution, we test whether the morphology of DLE craters is consistent with either the atmospheric-effects models (Fig. 5A and B) or landslide models (Fig. 5C and D) for their formation. We further assess the association of DLE craters with ice-related features to determine which of the two landslide models (i.e., landslide assisted by lubricating surface-ice; Weiss and Head, 2013, or a landslide assisted by melted pore-ice; Wulf and Kenkmann, 2015) is more likely for the formation of the inner ejecta facies of DLE craters (Fig. 5C and D). We first test the landslide and atmospheric-effects models by conducting a morphological analysis of a sample of DLE craters to assess ejecta/groove relationships to preexisting topography (Section 2). We then examine recent independent evidence for surface ice associated with DLE craters (Section 3), and assess groove and rampart formation by comparing the ejecta of DLE craters to terrestrial landslide deposits (Section 4). Finally, we review the ejecta-deposition chronology proposed for DLE craters (Section 5).

2. Morphological tests: Landslide versus atmospheric-effects

The radial grooves (Fig. 3) present on the inner ejecta facies of double-layered ejecta (DLE) craters (Fig. 1A and 3A) have been the subject of debate (Boyce and Mouginis-Mark, 2006; Weiss and Head, 2013; Harrison et al., 2013; Boyce et al., 2014; Boyce and Mouginis-Mark, 2014; Wulf and Kenkmann, 2015; Pietrek et al., 2016, 2017). Previous investigators have attributed them to either scouring of the primary ejecta facies by atmospheric vortices (Schultz and Gault, 1979; Schultz, 1992; Komatsu et al., 2007) or base surge (Boyce and Mouginis-Mark, 2006; Komatsu et al., 2007; Harrison et al., 2013), or as primary landslide features (Weiss and Head, 2013; Wulf and Kenkmann, 2015) (Fig. 5). We conducted a global survey of DLE craters using THEMIS and CTX images to find crater ejecta interacting with pre-existing topography. Out of the 752 DLE craters examined, we present three illustrative examples which may provide insight into the formation of the inner ejecta facies. We selected these craters due to their relatively well-preserved grooved texture and morphology, and unique target characteristics: they each formed in spatially heterogeneous pre-existing topography. Because the landslide models suggest that the inner ejecta facies was ballistically emplaced and transitioned into a gravity-driven flow off of the uplifted rim, the relationships between the ejecta morphology and pre-existing topography of these craters may provide insight into the ejecta emplacement processes that were operating. We use the relationship of the inner facies to pre-existing topography in order to test the different model predictions outlined below:

Groove formation through scouring (Fig. 5A and B): In this hypothesis, the inner

ejecta facies is ballistically emplaced and flows outward. Scouring from either atmospheric vortices or a base surge erodes the grooves and the depressed annulus, and the suspended material is deposited as the outer facies. In this scenario, the outer facies should only be present if the grooved inner facies is present. This reasoning applies only to relatively well-preserved craters where the grooves and/or outer facies are unlikely to have been eroded away.

Groove formation through a landslide (Fig. 5C and D): In this scenario, the ejecta are ballistically emplaced onto the target. The near-rim ejecta landslides off the uplifted rim immediately upon emplacement, forming the inner facies. Grooves form on the inner facies through shear/splitting in the landslide. The outer facies is composed of ejecta ballistically emplaced beyond the structurally uplifted rim which then slides/flows outwards. In this case, grooves would be expected to be present near the rim only where the inner ejecta facies is fully developed. The outer facies (uniquely characterized by high EM, high sinuosity, and thin, shorter-wavelength grooves) may be present in cases where the inner facies (characterized by low EM, low sinuosity, and longer-wavelength grooves) is not present.

2.1 Example 1: Impact onto a crater wall

A 9 km diameter DLE crater that formed just inside the walls of two preexisting craters (Fig. 6A) illustrates an example where the target topography to the northeast (the pre-existing crater rim) prevented the formation of the inner ejecta facies in the upslope direction. The northeast zone near the rim crest (shown in yellow in Fig. 6B) has thick ejecta of gradually decaying thickness (Fig. 6C) and appears subdued in map-view

compared with the sharp rim crest adjacent to the inner ejecta facies in the southeastern half (Fig. 6D). The northeast ejecta deposit near the rim crest, upslope from the crater, smoothly transitions into the outer ejecta facies (Fig. 6D) and does not display any observable grooves. This is in contrast to the prominent southwest section of the inner ejecta facies (Fig. 6B), which displays radial grooves originating from the sharp rim crest (Fig. 6D).

2.2 Example 2: Oblique impact on a slope

Another example of a DLE crater where the local topography has influenced emplacement of the inner ejecta facies is shown in Fig. 7A. The impact occurred northwest of Elysium Mons, on a regional slope of 1.7° trending east-west such that the downslope surrounding terrain to the west of the crater is ~ 100 m lower elevation than the upslope surrounding terrain to the east of the crater (Fig. S2 in supporting online material). The lack of ejecta in the western section (Fig. 7B) suggests that the impact which formed this crater was oblique, and the impact direction was broadly west to east. This likely caused bulking of ejecta to the east, in the upslope direction. The crater exhibits a defined inner ejecta facies and radial grooves in the southwestern (downslope) half (Fig. 7A and B). Much like the previous example (Fig. 6), the ejecta appears thicker near the rim crest in the northeastern (upslope) quadrant and exhibits a smooth gradation into the outer facies (Fig. 7C). No grooves are observed on the upslope northeastern ejecta deposit near the rim crest, which also lacks an inner facies (Fig. 7D), in contrast to the downslope southwestern ejecta which exhibits a defined inner facies with grooves (Fig. 7E).

2.3 Example 3: Impact onto a mound

An example of a DLE crater that formed adjacent to a pre-existing, high-topography mound on the southern rim-crest is shown in Fig. 8A-C. No grooves are observed on the top of the mound (Fig. 8D). The inner ejecta facies directly downslope from the mound has a longer runout distance by ~5 km (the width of the mound) relative to adjacent ejecta within the inner facies (green area in Fig. 8B). The presence of the mound appears to have offset the ejecta downrange from the crater without interfering with its emplacement.

2.4 Interpretation

If the inner ejecta facies were the primary ejecta facies (i.e., in the atmospheric vortex and base-surge models (Fig. 5A and B) and the grooves and outer ejecta facies formed through scouring and subsequent deposition of the eroded material, the inner ejecta facies would be expected to (1) always be present, and (2) display radial grooves if there were an outer ejecta facies. According to most scouring hypotheses (Schultz and Gault, 1979; Schultz, 1992; Boyce and Mouginis-Mark, 2006; Komatsu et al., 2007), the larger runout and more sinuous ejecta facies, which corresponds to the outer facies in typical DLE craters, should also include an inner ejecta facies. This is not the case in the examples discussed above (Figs. 6-8). The presence of grooves only in areas that display a fully developed inner ejecta facies (Figs. 6-8) suggests that these two features are genetically related. If grooves were to form through a scouring process, their presence would not be expected to be strongly influenced by topography, and they would be expected on the

mound in Fig. 8D. The mound may have avoided scouring if it were composed of significantly stronger material, but we consider this unlikely given that scouring by impact blast winds has been observed on Mars in wrinkle-ridged plains several tens of kilometers away from the parent crater (Fig. 1 in Quintana and Schultz, 2014). These examples show that radial grooves are only present where the formation of an inner ejecta facies is not prevented by target topography (Figs. 6-8), which is inconsistent with the suggestion that the inner ejecta facies is the primary ejecta, and the grooves and outer facies form through atmospheric effects.

Can these unique morphologies (Figs. 6-8) alternatively be explained by the landslide models? Formation of the rim-crest in a topographic low adjacent to a pre-existing crater wall could have reduced the structural uplift slopes to the northeast for the DLE crater shown in Fig. 6, preventing a landslide and resulting in thickened ejecta to the northeast. Impact into a local/regional slope for the DLE crater in Fig. 7 is predicted to increase the structural uplift angles to the west (where the topography slopes downhill), and decrease the structural uplift angle in the east (Fig. S2 in supporting online material), which would favor a landslide (and inner facies formation) to the west, and a thickened rim to the east. The effects of regional slope in this case were likely to have been enhanced by the oblique nature of the impact, which may have resulted in bulking of ejecta in the upslope direction (where it would be less likely to landslide off the rim). It appears that the mound on the southern rim-crest of the DLE crater in Fig. 8 has offset the inner facies downrange from the rim-crest by ~5 km, consistent with the mound offsetting the ejecta sliding-plane (ejecta was still able to slide off the steep-sided mound; Fig. 8C), as expected if the inner facies formed as a landslide.

These examples discussed above (Figs. 6-8) show that the morphology of the inner ejecta facies of DLE craters is consistent with a landslide mode of formation; however, the nature of the target substrate remains in question. For example, did the landslide occur on a surface composed of snow/ice as proposed in the glacial-substrate model (Weiss and Head, 2013)? Alternatively, did the landslide result from volatile-rich near-rim ejecta sliding on a rocky surface as proposed by Wulf and Kenkmann (2015)? Next, we evaluate the evidence for ice associated with DLE crater ejecta in order to distinguish between the two landslide models.

3. Ejecta association with ice-related features

Did the landslide occur on snow/ice (Weiss and Head, 2013) or rock (Wulf and Kenkmann, 2015)? In order to address this question, we evaluate the associations of DLE crater ejecta with ice-related features and assess whether these are consistent with ice sheets present at the surface at the formation time of any given DLE crater (Weiss and Head, 2013), or if instead they are consistent only with the presence of pore-ice within the ejecta (Wulf and Kenkmann, 2015). We first examine sublimation pits associated with pedestal craters and some DLE craters (Section 3.1), then present two examples of ring-mold craters associated with DLE crater ejecta (Section 3.2). We discuss recent work on expanded secondary craters and their relationship to ice associated with DLE craters (Section 3.3), and then review the significance of the excess ejecta crater population for substrate characteristics of DLE craters (Section 3.4). Finally, we summarize the implications of the ice-related features on the formation mechanism of DLE craters (Section 3.5).

3.1. Pedestal craters and marginal sublimation pits

Pedestal craters (~2 km diameter on average) are characterized by a crater bowl perched atop a plateau (Kadish et al., 2008). Pedestal craters are present poleward of ~33°N and ~40°S in the mid-high latitudes (Kadish et al., 2009), are elevated above the surrounding terrain by ~20-200 m (~50 m on average), and are hypothesized to result from impact into a surface snow and ice layer (Wrobel et al., 2006; Kadish et al., 2008; 2009; 2010; Kadish and Head, 2011; Barlow et al., 2014). Weiss and Head (2013) interpreted pedestal and DLE craters as a size continuum of craters that formed in surface ice (e.g., Kadish and Head, 2011): smaller impactors form pedestal craters, while larger impactors excavate ejecta from below the surface ice (e.g., Weiss and Head, 2015) and create DLE craters. Kadish et al. (2008) identified sublimation pits marginal to some of the pedestals of these craters located in Utopia Planitia and Malea Planum, and interpreted their presence to indicate the existence of an icy layer composing the pedestal. Subsequent SHARAD radar analysis provided support for the icy composition of pedestals (Nunes et al., 2011). Our examination of craters within Utopia Planitia finds a number of DLE craters with marginal pits; two examples of DLE craters with marginal pits are shown in Fig. 9A and G (also see white arrows in Fig. 16A). The pits are topographic lows which border the outer ejecta facies (Fig. 9C-F). Much like the sublimation pits found marginal to pedestal craters (Kadish et al., 2008), we do not find the pits to be preferentially pole-facing or equator-facing. Due to their morphologic similarity with the sublimation pits observed around pedestal craters (nearby pedestal craters depicted as red arrows in Fig. 9G) (Kadish et al., 2008), we interpret the pits

marginal to the outer ejecta facies of DLE craters as sublimation pits. If the interpretation that sublimation pits around pedestal craters are the result of the underlying glacial substrate (Kadish et al., 2008) is correct, the pits observed around the DLE craters in Fig. 9 also point to the presence of an icy layer underlying the ejecta.

3.2. Ring-mold craters

Ring-mold craters (RMCs) are a unique morphologic crater class characterized by an interior hummocky zone on the crater floor, surrounded by an inward-facing annulus within the crater interior, and concentric fractures exterior to the crater (Mangold, 2003; Kress and Head, 2008; Baker et al., 2010; Head and Weiss, 2014; Levy et al., 2016). RMCs are interpreted to result from an impact into a massive ice deposit superposed by a thin debris cover (Kress and Head, 2008; Baker et al., 2010; Head and Weiss, 2014), and thus their presence is used to indicate the presence of buried ice deposits. We report on two RMCs, identified by Levy et al. (2016), that are associated with DLE craters. An 11.5 km diameter DLE crater is shown in Fig. 10A. An RMC is located in the southeastern section of its inner ejecta facies (Fig. 10B). Another DLE crater (15 km in diameter) is shown in Fig. 10C which exhibits an RMC crater located in the southeastern section of its inner ejecta facies (Fig. 10D). Because the RMCs are superposed on the ejecta of the DLE craters, (and are thus stratigraphically younger) the surface ice layer must have been present before the DLE craters formed. In this scenario, the ejecta of the DLE crater acts as the debris-cover required for RMC formation.

Note that the larger craters present to the southwest of the RMCs in Fig. 10A and 9B appear to pre-date the formation of the DLE craters, and thus would not be expected to

exhibit a ring-mold morphology. Several superposing craters are comparably sized to the RMC on the ejecta facies of the DLE crater in Fig. 10A but do not exhibit ring-mold morphology. Based on the observation that the ejecta proximal to the RMC in Fig. 10A appears substantially elevated relative to the rest of the ejecta, we raise the possibility that variations in ejecta or ice layer thickness may have prevented other superposing craters from being RMCs. In any case, these examples (Fig. 10) suggest that at the time that the RMCs formed, excess ice was present beneath the inner ejecta facies of each of these DLE craters. This observation is consistent with the glacial substrate model for DLE crater formation.

3.3. Expanded secondary craters

Recent SHARAD analyses and studies of terraced crater (Bramson et al., 2015) and “expanded” secondary crater (Viola et al., 2015) morphologies (Fig. 11A) are interpreted to indicate the presence of decameters-thick buried ice within Arcadia Planitia (~40-60°N). Unlike their more typical counterparts (e.g., Fig. 8A), some DLE craters that superpose this buried ice deposit exhibit observable secondary craters (e.g., Fig. 1A) (Viola et al., 2015). Many of these secondary craters have a unique “expanded” morphology (Fig. 11B) which is interpreted to form through impact of secondary material into debris-covered ice (Viola et al., 2015). Viola et al. (2015) interpret the ice deposits to have been emplaced by airfall snow deposition during periods of higher obliquity (e.g., Head et al., 2003), followed by sublimation of the ice during periods of lower obliquity. The debris-cover is interpreted to be a thin sublimation-lag deposit derived from small amounts of dust that was intermixed with the ice (Viola et al., 2015). Additionally, Viola

et al. (2017) found that expanded secondary craters are present in greater abundance within the ejecta facies of a 15 km diameter DLE crater than in the surrounding plains (Fig. 11B), which led these authors to conclude that massive ice was underlying the ejecta facies at the time of secondary crater formation. Viola et al. (2017) further established that the expanded secondary craters present within the inner ejecta facies exhibit a greater degree of expansion (relative to those in the outer facies), and suggested that the inner facies contained a higher abundance of excess ice when the secondaries formed.

These observations are consistent with the glacial substrate model for DLE crater formation, which also predicts excess ice to remain present below the ejecta facies of DLE craters after the obliquity shifts to lower values and the unarmored surface ice sublimates. Based on the presence of expanded secondary craters in the plains surrounding the craters characterized by Viola et al. (2015, 2017), it appears that the surface ice layer remains extant under a thin sublimation lag in Arcadia Planitia (e.g. Steinheim; Fig. 1A) (Viola et al., 2015). Weiss and Head (2013) suggested that the paucity of secondary craters around most DLE craters was caused by the sublimation and removal of the surrounding ice layer. The observations by Viola et al. (2015, 2017) thus suggest that in the rarer cases of DLE craters that do exhibit secondary craters, the surface ice may still be present in the surrounding terrain below a thin sublimation lag. Furthermore, the observation that the inner ejecta facies may have more underlying excess ice compared with the outer facies (Viola et al., 2017) is consistent with the results of Black and Stewart (2008), who showed that thicker ejecta more effectively inhibits sublimation of underlying surface ice deposits. For example, Table 3 in Black and

Stewart (2008) shows that for a typical mid-latitude surface temperature of 190 K (e.g., see Fig. 7 in Haberle et al., 2003), a debris cover of 10 cm is predicted to armor a 50 m thick ice deposit for 162 Ma, and a 10 m thick debris cover is predicted to armor a 50 m thick ice deposit for 780 Ma. In the context of the glacial substrate model, the thicker inner ejecta facies (Fig. 4A) is predicted to inhibit the sublimation of the underlying ice more effectively than the thinner outer facies (Fig. 4A), which may contribute to the thinner ice deposit below the outer ejecta facies currently observed by Viola et al. (2017). The expanded secondary craters characterized by Viola et al. (2017) penetrate to depths of ~2 m on average, which is sufficient to armor a 50 m thick ice deposit for ~290 Ma in the Black and Stewart (2008) model. We note that further work is required to determine if the ejecta thicknesses (particularly the inner facies) expected for DLE craters are consistent with the shallow depth to the ice layer inferred by Viola et al. (2017).

3.4. Excess ejecta craters

“Excess ejecta” craters are martian impact craters defined as having a larger observed volume of ejecta than is predicted to have been excavated by the impact which formed the crater (Meresse et al., 2006; Black and Stewart, 2008; Kadish and Head, 2011). The excess ejecta effect is interpreted to result from the presence of buried ice deposits underlying the ejecta, which armors the underlying ice from sublimation subsequent to an obliquity shift. In this scenario, the buried ice artificially increases the measured volume of ejecta subsequent to the sublimation of the surrounding unarmored ice in a later, different climate regime (Meresse et al., 2006; Black and Stewart, 2008; Kadish and Head, 2011). Because each of the reported excess ejecta craters are also classified as DLE

craters, Weiss and Head (2013) noted that the excess ejecta crater population raises the possibility that all DLE craters could have formed through impact into a target covered with surface ice. Our preliminary analysis of the excess ejecta crater population (Meresse et al., 2006; Black and Stewart, 2008; Kadish and Head, 2011) finds that, much like the entire DLE crater population, the excess ejecta craters exhibit a paucity (and more typically, a complete absence) of secondary craters, consistent with the suggestion that the unarmored ice has sublimated away (Weiss and Head, 2013).

Here, we illustrate how the excess ejecta effect is consistent with the presence of an ice layer underlying the ejecta by presenting the excess ejecta ratio, defined as the measured ejecta volume ($V_{observed}$) divided by the expected ejecta volume ($V_{expected}$), for the entire reported excess ejecta population (which are also DLE craters) from Black and Stewart (2008), Kadish and Head (2011), Schaefer et al. (2011), and Wulf and Kenkmann (2015). The expected ejecta volume for a given crater diameter (D) is found as the volume of a paraboloid, multiplied by a factor of 1.1 to match the Maxwell Z model (Maxwell, 1977; Weiss and Head, 2016): $V_{expected} = \frac{1.1\pi}{2} (D_T/2)^2 d$, where excavation depth, $d = 0.1D_T$ (Croft, 1980). The transient crater diameter is found as $D_T = D_{SC}^{0.15} D^{0.85}$ (Croft, 1985), where D_{sc} is the simple-complex crater transition (global average is 6 km on Mars; Robbins and Hynes, 2012). The excess ejecta ratios ($V_{observed}/V_{expected}$) of the entire reported excess ejecta crater population is shown as the colored squares in Fig 12. For comparison, we co-plot the predicted excess ejecta ratios for craters with a variety of different underlying ice thicknesses (t_i) and ejecta runout distances (red and blue lines in Fig. 12). The model lines are found as $\frac{V_{expected} + V_{ice}}{V_{expected}}$, where V_{ice} is the volume of the ice layer of a given radial extent, which is determined by

the ejecta runout distance (represented as ejecta mobility): $V_{ice} = t_i \pi \left(\frac{(EM+1)D}{2} \right)^2 - t_i \pi \left(\frac{D}{2} \right)^2$. The excess ejecta craters explored in this study are characterized by EM ratios which cluster around ~ 3.2 and ~ 8.4 (depicted as the colored squares in Fig. 12); we thus use EM values of 3 and 9 for the model lines.

The interpretation that these craters have ice underlying the ejecta is supported by the close correspondence between the observed excess ejecta ratios (squares in Fig. 12) and the model lines (red and blue lines in Fig. 12), which are within the range of ice thicknesses measured for pedestal craters (≤ 20 m to ~ 200 m; Kadish et al., 2010).

The result that the excess ejecta ratio decays to a constant value (typically between 1-2 for the average EM of 3 and ice thickness less than 50 m) at crater diameters above ~ 25 km demonstrates that the volume of ejecta vastly exceeds or matches the volume of underlying ice for larger craters. In the context of the glacial substrate model, this might explain why DLE craters are typically smaller than ~ 25 km in diameter (Fig. 2): at craters larger than ~ 25 km in diameter, the large volume of ejecta (relative to the surface ice) overwhelms the surface ice through bulldozing and assimilation by the process of ballistic sedimentation (Oberbeck, 1975). In this case, the obliteration of the lubricating surface ice layer could prevent the landslide of the near-rim ejecta for craters larger than ~ 25 km in diameter.

Note that Steinheim crater (Fig. 1A), marked by the dashed red arrow in Fig. 12, has an excess ejecta ratio of ~ 1 (Wulf and Kenkmann, 2015), and thus does not appear to have excess ice relative to the surrounding terrain. Wulf and Kenkmann (2015) argued that this indicates a lack of an ice layer below the ejecta beneath Steinheim and the martian DLE crater population. We note, however, that since the Wulf and Kenkmann

(2015) study was published, a recent study by Viola et al. (2015) found abundant expanded secondary craters in the plains surrounding Steinheim crater (Section 3.3). In concert with the low excess ejecta ratio observed for Steinheim crater (red arrow in Fig. 12) this may instead indicate that the surrounding ice has not yet sublimed away at Steinheim due to an armoring sublimation lag, preventing the excess ejecta effect. On this basis, we predict Steinheim crater (Fig. 1A) to be represented in the present day by the configuration depicted later in the text in Fig. 19E, with a thin sublimation lag overlying the excess ice in the surrounding plains.

Next, we summarize the results of our survey of ice-related features with DLE craters and assess the implications for DLE crater formation models.

3.5. Summary of ice associations with DLE craters

In summary, the presence of surface ice at the time of impact during DLE crater formation is consistent with:

(1) The observation of sublimation pits marginal to the ejecta facies of DLE craters. These marginal sublimation pits are interpreted to reflect the sublimation of a surface ice deposit for the relatively smaller pedestal craters (Kadish et al., 2008).

(2) The identification of ring-mold craters on the ejecta facies of DLE craters. Ring-mold craters are interpreted to result from an impact into a debris-covered ice deposit (Kress and Head, 2008; Baker et al., 2010; Head and Weiss, 2014; Levy et al., 2016).

(3) The concentration of expanded secondary craters on the ejecta facies of a number of DLE craters demonstrated by Viola et al. (2015, 2017). Expanded secondary craters are interpreted to form through the sublimation of a relatively pure near-surface ice layer

(Viola et al., 2015, 2017).

(4) The excess volume of material (interpreted to be buried ice) present beneath the ejecta of some DLE craters (Fig. 12) (Meresse et al., 2006; Black and Stewart, 2008; Kadish and Head, 2011; Schaefer et al., 2011; Weiss and Head, 2013, 2014).

(5) The concentration of DLE craters in the middle-to-high latitudes (Fig. 2), where non-polar surface ice deposits emplaced during periods of higher obliquity are common (e.g., Head et al., 2003, 2006a, 2006b, 2010; Kadish et al., 2009; Madeleine et al., 2009, 2014; Levy et al., 2014; Weiss and Head, 2014; Fastook and Head, 2014; Fassett et al., 2014).

These factors are consistent with the glacial substrate model (Weiss and Head 2013), but are inconsistent with the DLE crater formation model in which the ejecta is proposed to be emplaced on a rocky surface rather than an ice sheet (Wulf and Kenkmann, 2015). It is important to note that this hypothesis (Wulf and Kenkmann, 2015) was motivated in part by the suggestion that the ~26 km diameter terrestrial Ries crater (Fig. 13A) (which formed in water-rich sediments overlying a crystalline basement; Pietrek and Kenkmann, 2016) is a terrestrial analog for martian DLE craters (Sturm et al., 2013; Kenkmann et al., 2015). This analogy is based on the observation of a thin discontinuous distal ejecta facies at Ries (Fig. 13A), which was directly compared to the outer facies of martian DLE craters (Sturm et al., 2013) (Fig. 13B and C). Significantly, the more common single-layered ejecta (SLE) craters (Fig. 13D) also frequently exhibit thin distal ejecta beyond the single ejecta facies (see yellow arrows in Fig. 13D) that shares similar topography to that observed at Ries (black line in Fig. 13B). Coordinates of four other examples of SLE craters with thin distal ejecta deposits are listed in the caption of Fig.

13. As shown in Fig. 13A (adapted from Sturm et al., 2013), the ejecta beyond $\sim 1.12 R$ from the rim-crest of Ries is thin and highly discontinuous, much like that observed around SLE craters (Fig. 13D). Furthermore, while the typical ejecta mobility ratio of Ries crater (1.12; Sturm et al., 2013) is similar to the average ejecta mobility ratio of ~ 1.5 for martian DLE craters (Barlow, 2005; Weiss and Head, 2014; Li et al., 2015; Barlow, 2015), we point out that it is also similar to the average ejecta mobility ratio of ~ 1.2 for martian SLE craters (Barlow, 2005; Weiss and Head, 2014; Li et al., 2015; Barlow, 2015). Consequently, we suggest that SLE craters should also be considered as a candidate martian analog for the terrestrial Ries crater, a scenario in which the target substrate in which Ries crater formed should not guide interpretations for the target in which martian DLE craters form. Importantly, given the heavily degraded state of the Ries crater, analogies relating Ries to specific martian impact crater morphologies should be evaluated with care.

Next, we compare the ejecta morphology of DLE craters to landslides in order to assess whether the grooves and ramparts may form through an analogous manner.

4. Ejecta comparison with landslides

The comparison between landslide morphologies and crater ejecta has been instrumental in guiding previous interpretations of ejecta emplacement and flow (Carr et al., 1977; Schultz and Gault, 1979; Barnouin-Jha et al., 2005; Mouginiis-Mark and Baloga, 2006; Weiss and Head, 2013; Xiao and Komatsu, 2013; Wulf and Kenkmann, 2015). Although we argue in Section 2 (Figs. 6-8) that the grooves present on the ejecta facies of DLE craters are unlikely to have formed through atmospheric interactions, it

remains a question as to whether the grooves could have formed through shear/splitting in a landslide mode (Weiss and Head, 2013; Wulf and Kenkmann, 2015). We first compare the ejecta facies of DLE craters with longitudinal grooves observed on landslides to assess if the groove morphology and morphometry is consistent with formation through shear/splitting in a landslide mode (Section 4.1). We then compare crater ejecta and landslides from a kinematic perspective to evaluate whether the presence of landslide grooves on crater ejecta is plausible (Section 4.2). Finally, we evaluate the unusual ramparts of DLE craters and examine why they might be more subdued compared with ramparts from their more globally-distributed (but equatorially-concentrated) counterparts (SLE and MLE craters; Fig. 2) (Section 4.3).

4.1. Groove comparison of ejecta and landslides

As discussed in Section 2, three examples of DLE craters exhibit unique ejecta relationships with pre-impact topography (Figs. 6-8). These relationships appear to be inconsistent with the atmospheric-effects and scouring hypothesis for groove formation. Is the groove morphology of DLE craters consistent instead with that of a landslide? Although previous studies have shown that the dimensions of the grooves are comparable between ejecta and landslides (Weiss and Head, 2013; Boyce and Mouginis-Mark, 2014; Pietrek et al., 2016, 2017), it remains a question as to whether the morphology of DLE crater grooves are consistent with a landslide origin:

- (1) Some grooves on the inner ejecta facies of Bacalor crater have been reported to cross from the inner facies onto the outer facies (Boyce and Mouginis-Mark, 2006), which, if true, is seemingly inconsistent with a landslide origin. It might also be possible that

atmospheric blast winds could be channeled by the topography of the pre-existing landslide grooves, which may have resulted in continued erosion of grooves across the contact of the inner and outer ejecta facies. The observation of grooves crossing the contact between the two ejecta facies remains in dispute, however: Wulf and Kenkmann (2015) showed that grooves do not appear to cross the contact between the inner and outer facies at Bacalor crater. Further work is required to resolve this discrepancy.

- (2) The initial analysis by Boyce and Mouginis-Mark (2006) reported that grooves on the inner facies of DLE craters are exclusively linear, and do not follow the underlying topography, which may be inconsistent with a landslide origin. Wulf and Kenkmann (2015), however, identified curved grooves on the inner facies which do appear to have been deflected around topography.
- (3) Grooves have been observed which appear to cut through perpendicular troughs (Boyce and Mouginis-Mark, 2016). These authors argued that the grooves must have formed after the ejecta came to rest (and formed the troughs), which is inconsistent with a landslide origin for the grooves. We note, however, that in the shear/splitting mechanism for formation of grooves in landslides, the grooves are not merely a surficial feature, but are instead predicted to propagate down to some depth. In this case, the observations by Boyce and Mouginis-Mark (2016) do not rule out a landslide origin for the grooves because the troughs may simply be cutting through the grooves and exposing them at a shallow depth.
- (4) The density of grooves (number of grooves per km²) on the inner facies of DLE craters appears to decrease with increasing crater diameter (Boyce and Mouginis-Mark, 2014). Because near-rim ejecta thickness increases with increasing crater diameter (rim

height, $H_R \approx 0.04D_T$; Melosh, 2011; p.273), the results of Boyce and Mouginis-Mark (2014) point to increasing groove wavelength with increasing ejecta thickness. This trend is also observed in terrestrial landslides (De Blasio, 2014), and we thus note that the observation of decreasing groove density with increasing crater diameter is consistent with a landslide mode of formation for the grooves on the inner ejecta facies of DLE craters.

- (5) As noted by Boyce and Mouginis-Mark (2014) and Pietrek et al. (2016), grooves present on landslides can generally be traced from the beginning of the landslide to its terminus, and are relatively constant in along-strike width. This is in contrast to some grooves present on martian DLE craters, which are observed to increase in width with distance from the rim-crest, and to initiate and terminate discontinuously along the length of the ejecta facies (Boyce and Mouginis-Mark, 2014; Pietrek et al., 2016). Boyce and Mouginis-Mark (2014), Boyce et al. (2014), and Pietrek et al. (2016) further observed that terrestrial landslides accommodate lateral spreading by “forking” or “fanning out” of grooves, wherein a new groove forms by splitting off from an adjacent groove; the end result of this process is that grooves on landslides are frequently perpendicular to the flow terminus. In contrast, Pietrek et al. (2016) proposed that lateral spreading in DLE crater ejecta is likely accommodated by widening of the grooves with distance from the rim-crest and the formation of parallel grooves between already established grooves. It remains unclear if these observations preclude a landslide origin, or whether such morphologic differences can arise due to differences in flow thickness or velocity (Boyce and Mouginis-Mark, 2014). Consequently, these observations require further consideration in order to evaluate their consistency with a landslide origin for the grooves

(Boyce et al., 2014).

- (6) The grooves on the inner ejecta facies of DLE crater Bacolor exhibit two different scales (Mouginis-Mark et al., 2014): shallow, shorter-wavelength grooves that are superposed on top of deeper, longer-wavelength grooves. Fig. 14A shows a DLE crater that exhibits this bimodal groove wavelength-distribution (black arrows denote superposing grooves in Fig. 14B). Mouginis-Mark et al. (2014) argue that the two scales of grooves on the inner ejecta facies are inconsistent with their formation as a landslide. Critically, landslides within Valles Marineris also exhibit shorter-wavelength grooves superposed on larger grooves (Pietrek et al., 2016). For example, Fig. 14C and D show a landslide in Ganges Chasma which exhibits shorter-wavelength grooves superposed on larger grooves (black arrows in Fig. 14E and F). If the DLE crater groove wavelengths (Fig. 14B) are compared with those of the Ganges Chasma landslide (Fig. 14E and F), it is clear that smaller grooves superpose larger grooves on both the ejecta facies of DLE craters and landslide deposits (also see Fig. 1 in Pietrek et al., 2016). We conclude that the observation of a bimodal groove distribution on the inner ejecta facies of DLE craters (Mouginis-Mark et al., 2014) is consistent with a landslide origin.

In summary, observations of DLE crater groove morphology show general consistency with grooves on landslides. Future work is required to better understand some of the minor differences, but these differences do not necessarily preclude a landslide origin. Next, we compare crater ejecta with landslides to assess if DLE craters exhibit kinematic similarity with landslides that exhibit grooves.

4.2. Kinematic similarity with landslides

Considering how the grooves formed on the ejecta facies of DLE craters is critical in understanding the target structure and mechanics related to DLE craters formation. In this section, we evaluate the kinematic characteristics of landslides for comparison with martian crater ejecta in order to evaluate whether DLE crater grooves may plausibly form through the same shear/splitting mechanism. Landslide grooves (Fig. 1B and C) form through shear and splitting of adjacent substreams of debris (Shreve, 1966; Dufresne and Davies, 2009; De Blasio, 2011, 2014). Previous investigators have noted that whether a landslide displays longitudinal grooves or not depends on several different factors:

(1) *Speed*: Dufresne and Davies (2009) noted that grooves were most prominent in terrestrial flow features emplaced at high speeds, which they attributed to higher longitudinal speeds (compared with lateral speeds). High lateral speeds break up grooves (Dufresne and Davies, 2009), whereas high inertial energy favors groove formation (Lucchitta, 1979, 1987).

(2) *Flow thickness*: De Blasio (2014) noted that thin terrestrial landslides appear to develop shorter-wavelength grooves, while thicker landslides typically develop longer wavelength grooves.

(3) *Basal friction*: Low basal friction values facilitate groove formation (Shreve, 1966; Lucchitta, 1979; De Blasio, 2011, 2014) because low basal friction inhibits uniform spreading (Shreve, 1966), and subdues the vertical velocity gradient in the flow. High vertical velocity gradients promote vertical/turbulent mixing within the landslide, which prevents groove formation (De Blasio, 2011, 2014).

Do DLE craters satisfy these constraints? In order to compare martian DLE craters with martian and terrestrial landslides, we must examine the flows under identical

kinematic conditions and account for the different gravities present on different planetary bodies. Ejecta and landslides are free-surface flows dominated by the effects of gravity and inertia, and so we employ a dimensionless number, the Froude number (Fr), to evaluate kinematic similarity. The Froude number is the ratio of inertial to gravitational forces: $Fr = \frac{U}{\sqrt{gh}}$, where U is the flow speed, h is flow thickness, and g is gravity. The Froude number is frequently used to delineate flow regimes in free surface flows such as avalanches (Borstad, 2005), debris flows (Davies, 1986), and granular flows (Augenstein and Hogg, 1978; Louge and Keast, 2001; Ancey, 2001; Pouliquen and Forterre, 2002; GDR MiDi, 2004) under different gravities (Brucks et al., 2007). Kinematic similarity (i.e., the equivalence of the combined effects of flow thickness, velocity, and gravity) occurs when Fr of different flows are equal. By comparing Fr of landslides and ejecta as a function of basal friction (μ), we can evaluate the kinematic similarity of DLE ejecta and landslides, in reference to all of the factors recognized in longitudinal groove formation to determine if grooves associated with DLE craters may plausibly form in a landslide mode.

We reviewed the literature to examine the velocities, thicknesses, and basal friction values for a number of martian and terrestrial landslides and debris/rock avalanches (Table 1). We adopt the average velocities and friction coefficients found by numerical models where available. For the landslides where numerical models were unavailable, we use the established minimum or maximum bounds on velocity from potential energy estimates, and approximate the friction coefficient as $\mu = \tan\theta + \frac{H_0}{\Delta L}$ (Lucas et al., 2014), where θ is the slope on which the final deposit is resting, H_0 is the initial thickness of the material prior to the landslide, and ΔL is the distance traveled. We also consider Fr and μ

data from granular flow laboratory experiments. We evaluate the ejecta facies of DLE, SLE, and MLE craters for a wide range of ejecta thicknesses between 20 m and 100 m, and flow speeds between 10-100 m/s (e.g., Fig. 5F in Weiss and Head, 2014) to account for the wide range of crater diameters (e.g., Fig. 2). We adopt the μ ranges from Weiss and Head (2014): $\mu < 0.1$ for DLE craters, and $\mu = 0.1-0.6$ for SLE and MLE craters.

The Fr - μ relationships for these flow features are shown in Fig. 15. In the landslide data, Fr appears to increase as a function of μ , consistent with granular-flow experimentation and theory (Augenstein and Hogg, 1978; Louge and Keast, 2001; Pouliquen and Forterre, 2002; GDR Midi, 2004). The most noticeable feature in Fig. 15 is the relationship between Fr , μ , and presence of grooves (Fig. 15C and D). It appears there is a Fr - μ dichotomy between flows that form grooves, and those that do not. Landslides with low Fr form grooves only if they have low μ values. Conversely, flows with higher Fr may exhibit grooves even with higher μ values. Thus, for any given μ , flows that possess grooves are associated with higher Fr . The relationship between the competing effects of Fr and μ appears to be a central factor in forming grooves. This result supports the suggestion of previous investigators that velocity, flow thickness, and basal friction are important factors in groove formation (Shreve, 1966; Lucchitta, 1979; Dufresne and Davies, 2009; De Blasio, 2011, 2014). Our results suggest that groove formation is most favorable when inertial forces are high and basal friction is low. We interpret these results to indicate that when gravitational forces reach some strength relative to inertial forces (low Fr), gravitational spreading is enhanced, and high lateral velocities prevent groove formation (Dufresne and Davies, 2009). Likewise, when μ is high, the strong resulting vertical velocity gradient promotes turbulent mixing in the flow

front, which prevents groove formation (De Blasio, 2011).

In summary, the ratios of inertial to gravitational forces that allow grooves to form are dependent upon μ : grooves can form under conditions of high gravitational forces as long as μ is low, or under conditions of high μ as long as inertial forces dominate over gravitational forces. Although the large range in minimum/maximum Fr and μ values prevent meaningful statistical treatment (Fig. 15A), for the average Fr - μ data shown in Fig. 15, we find based on visual examination that the critical Froude number (Fr_c) above which grooves form appears to be of the approximate form, $Fr_c \approx 10\mu^{\frac{3}{5}}$.

We co-plot the predicted Fr - μ range of the layered ejecta craters in Fig. 15B. The predicted range of Fr - μ for DLE craters is shown in the honeycomb-region in Fig. 15B. The Fr and μ values of DLE crater ejecta are comparable and exhibit overlap with many landslides that exhibit grooves, but do not typically overlap landslides that do not exhibit grooves. The predicted Fr - μ range does extend below the critical Fr_c line (dashed black line in Fig. 15), but this remains consistent with the observation that some relatively fresh smaller DLE craters do not exhibit grooves on the inner facies (e.g., see Fig. 16).

Numerous martian landslides in Valles Marineris exhibit grooves, but it remains uncertain whether these landslides ran out onto ice sheets as proposed by De Blasio (2011, 2014), Gourronc et al. (2014), and Mazzanti et al. (2016). Out of the twenty terrestrial landslides plotted with grooves, however, seventeen of them ran out on a glacier or ice sheet. None of the ten terrestrial landslides without grooves ran out onto surface ice. Thus, while other models for DLE crater formation could possibly account for the low basal friction values, the kinematic similarity between DLE crater ejecta and terrestrial landslides which ran out on glaciers (and exhibit grooves) is highly suggestive

of a comparable target (i.e., surface ice), and sliding/flow kinematics.

Some of the other martian layered-ejecta craters also display grooves in their ejecta facies. For example, both SLE and MLE craters may in some cases exhibit longitudinal grooves (Barnouin-Jha et al., 2005; Boyce et al., 2014; Wulf and Kenkmann, 2015). We co-plot the predicted Fr - μ range of SLE and MLE craters as the cross-hatched region in Fig. 15B. Compared with DLE craters, SLE and MLE craters are predicted to have similar Fr , but SLE/MLE crater ejecta are interpreted to have been emplaced on a higher μ , rocky target, rather than surface ice (Carr et al., 1977, Mouginis-Mark, 1981; Costard, 1989; Barlow and Bradley, 1990; Barlow, 1994, 2005, 2006; Stewart et al., 2001; Baratoux, 2002; Barlow and Perez, 2003; Oberbeck, 2009, Weiss and Head, 2014; Jones and Osinski, 2015; Jones, 2015). The higher μ value of these craters places the SLE/MLE crater Fr - μ field (crosshatched field in Fig. 15B) largely beneath the predicted critical Fr_c value for groove formation. Thus, the lower ejecta runout distances and ejecta mobility values (interpreted to be due to their higher μ) would predict that SLE and MLE craters should exhibit grooves less frequently than DLE craters, consistent with observation.

We conclude that the ejecta facies of DLE craters are kinematically and frictionally comparable to landslides that formed on surface ice and exhibit grooves. This Fr - μ similarity strongly suggests that the landslide analogy for DLE craters (Weiss and Head, 2013; Wulf and Kenkmann, 2015) and the other layered ejecta craters (Barnouin-Jha et al., 2005) is appropriate, and that the longitudinal grooves on these features could have plausibly formed through the same shear/splitting mechanism as exhibited by landslides. The presence of grooves on the other layered ejecta craters (Barnouin-Jha et al., 2005; Boyce et al., 2014; Wulf and Kenkmann, 2015) may also be explained by kinematic

similarity to grooved landslides, but the paucity of grooves (compared with DLE craters) is likely to be due to their higher μ values, which can be interpreted to indicate a rocky target surface, rather than an icy one. Because groove formation requires either low basal friction or a high Froude number (Fig. 15), this model predicts that the efficiency of groove formation on the martian layered ejecta crater population should generally correlate positively with ejecta mobility, which can be tested with further morphologic observation.

Next, we discuss the unique conditions of DLE crater ejecta that may allow the formation of more subdued ramparts.

4.3. Distal rampart formation

Prominent distal ramparts tens to hundreds of meters thick are observed along the margins of the ejecta facies of many of the layered-ejecta craters (McCauley, 1973; Baloga et al., 2005; Baratoux et al., 2005; Mouginis-Mark and Baloga, 2006; Boyce et al., 2010). Can the distal ramparts around the layered ejecta craters provide information on the flow conditions? When observed with sufficiently high-resolution visible-wavelength images or thermal inertia data, distal ramparts on the layered ejecta craters frequently appear to be composed of larger particles than the rest of the ejecta facies (Fig. 17) (Baratoux et al., 2005; Mouginis-Mark and Baloga, 2006; Wulf et al., 2013; Jones et al., 2016). Baratoux et al. (2005) and Boyce et al. (2010) attributed this observation to kinetic sieving (Middleton, 1970; Savage and Lun, 1988; Pouliquen and Vallance, 1999), wherein gravity causes smaller particles in a granular media to move downwards where pore space is sufficiently large. Small particles percolate downwards more often because

they can infiltrate through smaller pores. Large particles thus segregate to the top of the flow, where the speed is highest, and are consequently transported to the front of the flow (Savage and Lun, 1988).

Ramparts have been compared to the distal rims observed in terrestrial debris flows and landslides (Barnouin-Jha et al., 2005; Baloga et al., 2005; Boyce et al., 2010), which are also characterized by blocky material (Shugar and Clague, 2011). In terrestrial debris flows, ramparts form as a result of “flow surges,” (Iverson, 1997 and references therein) wherein the process of kinetic sieving transports larger particles to the front of the flow. The larger particles are able to dissipate pore pressure more efficiently than finer material (Gray and Ancey, 2009). Because low pore pressure increases basal friction (Gray and Ancey, 2009) (and vice-versa), the presence of large particles at the flow-front thus causes the flow-front to decelerate relative to the rest of the flow, forming ramparts as the flow comes to a stop (Iverson, 1997). The suggestion that kinetic sieving and particle size control the formation of ramparts is supported by numerical modeling: debris-flow models that have low pore pressure at the flow-front can reproduce rampart morphologies (Iverson, 1997; Savage and Iverson, 2003), while models with uniform pore pressure instead produce a tapered flow front (Savage and Hutter, 1989; Savage and Iverson, 2003).

In the cases of single-layered and multiple-layered ejecta craters, the ejecta slides on a rocky surface with high friction (e.g., Weiss and Head, 2014). This is predicted to allow a strong vertical velocity gradient to develop within the ejecta, favoring kinetic sieving (left panels in Fig. 18). In this scenario, large particles are transported to the flow-front, and ramparts develop through preferential deceleration of the flow-front (left panels in Fig.

18).

Unlike SLE and MLE craters, the outer ejecta facies of DLE craters typically lack pronounced distal ramparts (Baratoux et al., 2005; Mouginis-Mark and Baloga, 2006; Barlow, 2006; Osinski et al., 2011); when present, they exhibit a more subdued morphology (e.g., compare the outer ramparts of DLE craters in the blue lines of Fig. 4A and 13B with the SLE rampart in the black line of Fig. 13B). Compared with SLE and MLE craters, DLE craters also exhibit fewer large particles on the margins of the outer ejecta facies. This is shown qualitatively with THEMIS nighttime infrared imagery in Fig. 17, where an SLE crater exhibits a strong thermal inertia signature in the distal ramparts (indicating the presence of larger particles), in contrast to the adjacent DLE craters which exhibit a relatively lower thermal inertia signature. This indicates the relative lack of larger particles in the distal ejecta of the DLE craters, despite the close proximity (~60 km) between the DLE and SLE craters (and thus likely identical bedrock target characteristics). In contrast, Jones et al. (2016) reported that several DLE craters exhibited larger particles at the ejecta margins, but our detailed review of these craters suggests that they are unlikely to be DLE craters and require updated classification. We suggest that the presence of surface ice at the time of DLE crater formation is ultimately responsible for the observations discussed above. The intervening icy layer allows DLE crater ejecta to encounter lower basal friction than SLE and MLE craters (Weiss and Head, 2014). Low basal friction values would reduce the vertical velocity gradient in the flowing ejecta (Wada and Barnouin-Jha, 2006) and would thus be expected to reduce the efficiency of kinetic sieving (right panels Fig. 18). This would produce a lower frequency of large particles at the margins of the outer ejecta facies of DLE craters relative to SLE

and MLE craters (e.g., Fig. 17). The relative absence of larger particles at the flow-front of DLE crater ejecta leads to less frictional resistance within the sliding/flowing ejecta (due to the higher pore pressures), generating a more tapered flow front and ramparts that appear subdued compared with those of SLE and MLE craters (Fig. 18). We conclude that the subdued nature of DLE crater ramparts are consistent with formation on a low-friction surface snow/ice layer. We now review the proposed chronology of ejecta deposition in the glacial substrate model.

5. Chronology of ejecta deposition

In summary, the glacial substrate model for DLE crater formation consists of several phases (Fig. 19):

(1) Decameters-thick snow and ice deposition during periods of higher martian obliquity ($\sim 35^\circ$), generally in the mid-high latitudes (Fig. 19A) (Head et al., 2003, 2006a, 2006b, 2010, Madeleine et al., 2009, 2014). Formation of DLE craters closer to the equator would occur in areas that accumulate snow deposits during periods of even higher obliquity ($\sim 45^\circ$; Forget et al., 2006).

(2) Impact of a meteoroid onto the martian surface (Fig. 19A).

(3) The resulting shock wave excavates ejecta from below the surface ice (Fig. 19B); impacts that do not excavate from below the ice instead form pedestal craters (Kadish and Head, 2011; Weiss and Head, 2014).

(4) A high-angle ejecta curtain forms due to the volatile-rich nature of the substrate (Fig. 19B) (e.g., Greeley et al., 1980).

(5) The near-rim ejecta is ballistically emplaced on the ice-covered rim of the

transient crater cavity (Fig. 19C). The ejecta speeds are sufficient to overcome the static friction values of the icy target surface. The landslide that forms the inner ejecta facies thus begins immediately during ejecta emplacement (Fig. 19C) and accelerates down the slope of the structurally uplifted rim (Fig. 4C). Shear of adjacent debris substreams (Shreve, 1966; De Blasio, 2011) and splitting in the azimuthally expanding ejecta produces linear grooves on the inner ejecta facies (Fig. 19D).

(6) Concurrent with the landsliding of the near-rim ejecta, which forms the inner ejecta facies, distal ejecta is being ballistically emplaced beyond the rim (Fig. 19C), and forms the outer ejecta facies. The ejecta comprising the outer facies continues to slide/flow on the low-friction surface ice (Fig. 19D). The relatively thinner outer ejecta facies produces shorter-wavelength grooves, while topographic deflection produces more sinuous grooves.

The inner facies may overthrust the outer facies to form the inner/outer facies contact (Wulf et al., 2012; Weiss and Head, 2013; Wulf and Kenkmann, 2015), resulting in rapid deceleration. The deceleration of the inner facies is interpreted to produce the tensile fissures observed along the margins of the inner ejecta facies (e.g., Fig. 1A, 16B and D).

Immediately following both inner and outer ballistic ejecta emplacement, the underlying snow and ice deposits will compress (Senft and Stewart, 2008) and frictionally melt (Sosio et al., 2008; De Blasio, 2014). Ballistic erosion (Oberbeck, 1975) and continued sliding will further entrain surface snow and ice deposits. The ejecta temperatures of craters in this diameter range are generally predicted to be less than 273K (Weiss and Head, 2016), preventing any surface-ice melting through conduction.

(7) Transient crater rim collapse occurs. Rim-collapse is believed to occur

immediately after, or even during the formation of the transient cavity (e.g., Melosh, 1989; Turtle et al., 2005; Senft and Stewart, 2007), and so rim collapse is likely to occur while ejecta sliding is still ongoing.

(8) Following ejecta emplacement, the DLE crater is predicted to be surrounded by the surface ice layer (Fig. 19E).

(9) The martian obliquity shifts to different values, whereupon the mid-high latitude surface ice is no longer stable: any surface ice not armored by overlying ejecta or debris sublimates away (Fig. 19F). Any secondary craters present in these deposits that are not large enough to excavate through the ice layer to the underlying regolith/rock are removed as the ice layer sublimates away (Fig. 19F). The nature of periodic, high-obliquity excursions within the martian Amazonian period (Laskar et al., 2004) implies that this cycle of snow deposition, meteoroid impact, DLE crater formation, and ice sublimation may have occurred numerous times, in a matter analogous to the periodic formation of pedestal craters (Kadish and Head, 2014).

SLE and MLE craters, therefore, are predicted to form in the areas not covered by sufficient surface ice: the equatorial regions during higher obliquity ($\sim 35^\circ$) excursions and near-globally during periods of lower obliquity ($< 35^\circ$) (e.g., Fig. 11 in Weiss and Head, 2014). This is consistent with (1) their near-global but equatorially concentrated latitudinal distribution (e.g., Fig. 2) (Barlow and Perez, 2003; Weiss and Head, 2014, 2017; Li et al., 2015), (2) their lower ejecta runout distances relative to DLE craters (Barlow, 2005; Weiss and Head, 2014; Li et al., 2015); and (3) their presence adjacent to DLE craters (Fig. 17). For example, the DLE craters in Fig. 17 are infilled with concentric crater fill (interpreted as debris-covered ice; e.g., Levy et al., 2010; Fastook

and Head, 2014), while the adjacent SLE crater appears to have minimal crater fill. We thus interpret the SLE crater in Fig. 17 to be younger and to have formed following the termination of the glacial period during which the adjacent DLE craters formed.

6. Conclusions

The target stratigraphy and formation mechanism of martian double-layered ejecta (DLE) craters (Fig. 1) have been debated (Fig. 5) since their discovery (Carr et al., 1977). In this contribution, we test different models for their formation (with special focus on the landslide model).

Our morphological analysis (Section 2) finds that longitudinal grooves (Fig. 3) are present on the surfaces of the near rim-crest ejecta of DLE craters only where an inner facies is fully developed. In cases where the near rim-crest ejecta gradually decays with increasing distance outward, grooves are not observed near the rim-crest, and a contact between an inner and outer facies is not observed. (Figs. 6-8). This strongly suggests that the formation of grooves and the emplacement of the inner ejecta facies are genetically linked. This is inconsistent with an atmospheric-effects origin for the grooves and formation of the inner facies (Schultz and Gault, 1979; Schultz, 1992; Boyce and Mouginis-Mark, 2006; Komatsu et al., 2007; Harrison et al., 2013), but is consistent with a landslide origin (Weiss and Head, 2013; Wulf and Kenkmann, 2015).

In order to test whether the landslide occurred over a target superposed by surface ice (Weiss and Head, 2013) or a rocky target (Wulf and Kenkmann, 2015), we evaluated the association of DLE crater ejecta with established ice-related morphologic features (Section 3). We provide examples where DLE crater ejecta is associated with marginal

sublimation pits (Fig. 9), and is superposed by ring-mold craters (Fig. 10) and expanded secondary craters (Fig. 11). All three of these features have previously been interpreted to reflect the presence of debris-covered ice (Kadish et al., 2008; Kress and Head, 2008; Head and Weiss, 2014; Viola et al., 2015, 2017). Their identification associated with DLE crater ejecta, in tandem with the relationships between DLE craters and excess ejecta craters (Fig. 12), suggests that massive ice (remnant from a surface ice layer present at the time of impact), is present beneath the ejecta of DLE craters.

We also assessed whether the grooves on DLE crater ejecta are consistent with a shear/splitting mechanism in an analogous manner to landslides (Section 4). We find that the morphology of the grooves on the inner ejecta facies of DLE craters is broadly consistent with a landslide origin. Critically, we find that ejecta facies of DLE craters are kinematically and frictionally comparable to terrestrial landslides that exhibit grooves (most of which also form on glaciers; Fig. 15), indicating that landslide grooves are indeed predicted to form on DLE crater ejecta based on kinematic and frictional similarity. We find that the relationship between the Froude number and basal friction heavily influences whether or not a flow feature exhibits grooves (Fig. 15), confirming the suggestion that flow velocity, thickness, gravity, and basal friction affect groove formation in gravity-driven free surface flows. Our observations suggest that a change in flow regime occurs for landslides, debris avalanches, and martian impact ejecta at a critical Froude number, $Fr_c \approx 10\mu^{\frac{3}{5}}$, above which longitudinal grooves are able to form. Finally, the outer ejecta facies of DLE craters may display more subdued ramparts containing fewer large particles than those of MLE and SLE craters (Fig. 17) due to the relatively lower basal friction for the DLE deposits, which inhibits kinetic sieving and

flow-front deceleration (Fig. 18).

We emphasize that we consider the presence of surface ice at the time of impact, the landslide of the near-rim ejecta, and the superposing grooves to be mutually inclusive. We do not favor landslide models which suggest the absence of surface ice (Wulf and Kenkmann, 2015) because of the numerous lines of evidence that point to surface ice at time of impact (Section 3 and 4). A landslide of near-rim ejecta is not predicted to occur without the lubricating surface ice. This is evidenced by the martian SLE craters, which are interpreted to form in a target rich in pore-ice (without surface ice) but do not exhibit any inner ejecta facies derived from a landslide of near-rim ejecta. Finally, the volatile-rich landslide model for DLE crater formation, which proposes that the ejecta is emplaced on a rocky surface (Wulf and Kenkmann, 2015) does not explain why DLE craters exhibit a relatively higher ejecta runout distance ($EM = \sim 3$; e.g., Barlow, 2005; Weiss and Head, 2014; Li et al., 2015) compared with other martian crater populations (SLE crater $EM = \sim 1$, MLE crater $EM = \sim 2$; e.g., Barlow, 2005; Weiss and Head, 2014; Li et al., 2015) whose ejecta is interpreted to have been emplaced on a rocky target. Further, we do not favor landslide models which accept the presence of surface ice at the time of impact, but favor an atmospheric-origin for the grooves (Boyce et al., 2016). This is because our examples in Section 2 are inconsistent with an atmospheric origin for the grooves, and because we consider the grooves to be the natural consequence of a landslide overriding an ice sheet, as evidenced by the kinematic and frictional similarity of DLE crater ejecta to terrestrial landslides which overran glaciers (Section 4.2 and Fig. 15).

We conclude that DLE craters form by impact and excavation through a surface snow

and ice layer (~50 m thick on average) that was deposited in the mid-latitudes of Mars during periods of higher obliquity than today (Fig. 19). The low-friction icy surface layer is fundamentally responsible for the formation of longitudinal grooves, the landslide which produces the inner facies, the long runout distance of the outer facies, and the subdued ramparts.

If this hypothesis is correct, the presence of DLE craters could have predictive value for assessing the distribution and timing of surface snow and ice deposits during more ancient climate regimes on Mars. Near rim-crest landslides may be a common process on planetary bodies which have both icy surfaces and craters with sufficient structural uplift (e.g., Ganymede; Boyce et al., 2010). Further examination of grooves on the ejecta of DLE craters may enhance understanding of longitudinal groove formation in free-surface flow features elsewhere in the Solar System.

The glacial substrate model can be further tested by: (1) comparing the rim heights of DLE craters to those of SLE and MLE craters (the glacial substrate model predicts that the rim-crest of DLE craters should have less ejecta, and therefore be lower than those of SLE and MLE craters for a given crater diameter); (2) SHARAD analysis of the ejecta facies of DLE craters. The glacial substrate model predicts that the surface ice may still underlie the ejecta, which may be observable for sufficiently thick ejecta/ice thicknesses given the ~10 m vertical resolution of SHARAD; and (3) evaluating the presence of grooves on the layered ejecta craters as a function of their ejecta mobility; the kinematic similarity proposed in Section 4.2 suggests that the presence of grooves should be favored in craters with high ejecta mobility. Future work is required to better understand several morphologic differences between DLE crater ejecta and terrestrial landslides

(Section 4.1) (Boyce and Mouginis-Mark, 2014; Pietrek et al., 2016, 2017) to fully assess their consistency with a landslide origin.

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Figures, tables, and captions:

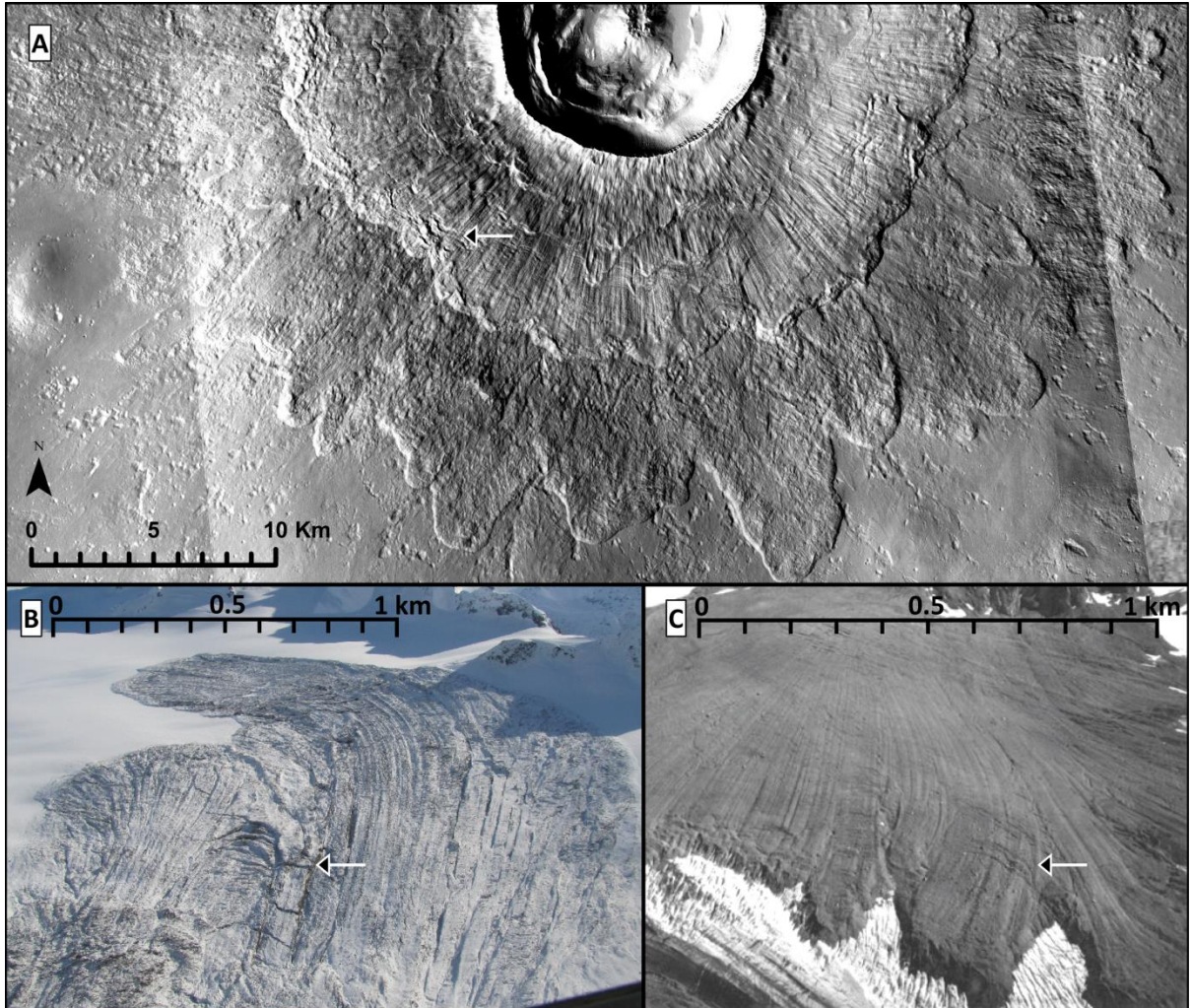


Figure 1. (A) Radial grooves and transverse fissures (white/black arrow) on the southern inner ejecta facies of the martian Steinheim crater (190.6°E, 54.5°N; CTX image P21_009160_2348). (B) The Mt. La Perouse, Alaska landslide ran out on an icy surface and displays prominent longitudinal grooves and transverse fissures (white/black arrow); flow direction is up. Landslide surface was covered in ~10 cm of snow subsequent to the landslide on Feb. 16, 2014 (Drake Olson personal communication, 2014). Photo courtesy Drake Olson. (C) The 1964 Sherman landslide in Alaska ran out on a glacier surface and displays prominent longitudinal grooves and transverse fissures (white/black arrow) (photo by Austin Post, U.S.G.S, 1965). Scale bars for (B) and (C) are approximate.

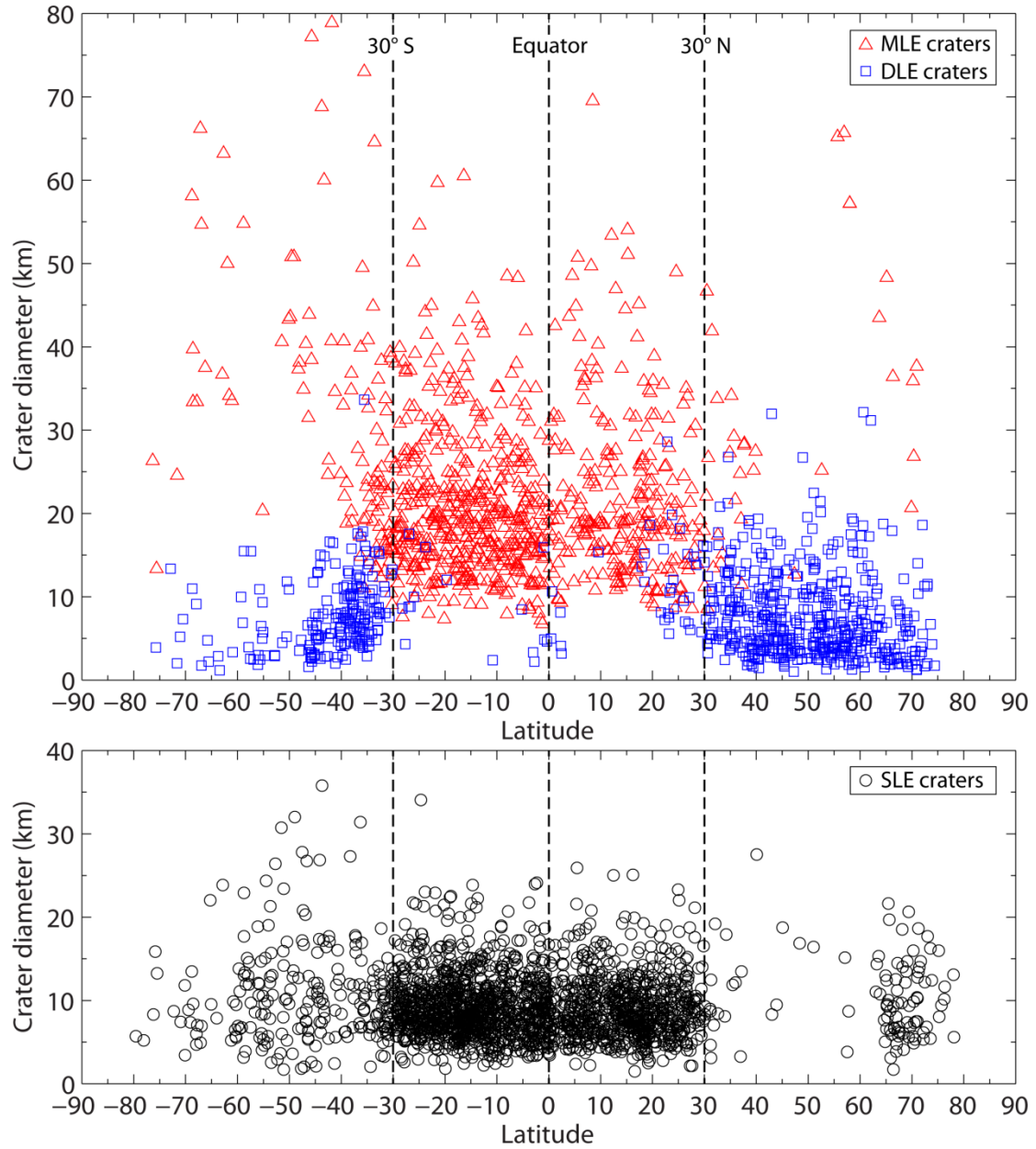


Figure 2. Diameter-latitude relationships for multiple-layered ejecta (MLE), double-layered ejecta (DLE) craters, and single-layered ejecta (SLE) craters from Weiss and Head (2014, 2017). DLE craters are concentrated in the mid-high latitudes, where non-polar surface ice deposits emplaced during periods of higher obliquity are common (e.g., Kadish et al., 2009; Madeleine et al., 2014; Levy et al., 2014; Fastook and Head, 2014; Fassett et al., 2014).

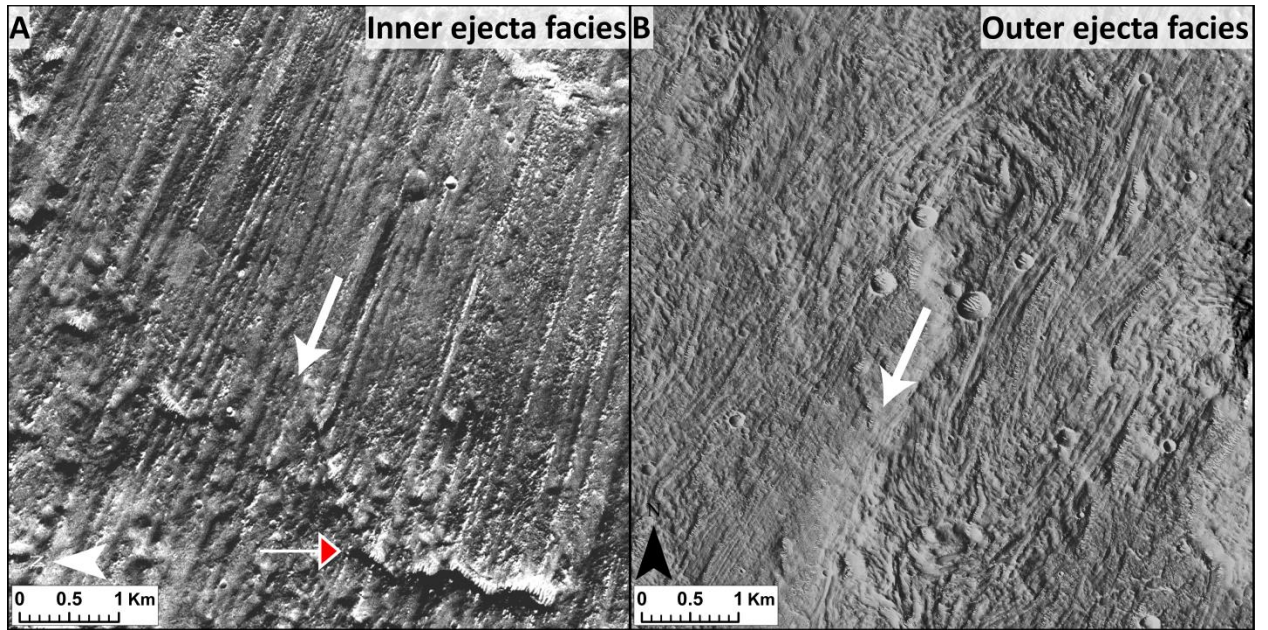


Figure 3. Ejecta facies of the 20.5 km diameter Bacolor crater (118.3°E, 32.4°N). (A) Northwestern section of the inner ejecta facies; flow direction (white arrow) is SE to NW; transverse fissures (red arrow). CTX image P17_007752_2140. (B) Southwestern edge of the outer ejecta facies; flow direction is NE to SW. HiRISE image PSP_007752_2125. Grooves on the outer ejecta facies are shorter-wavelength and more sinuous than those of the inner facies.

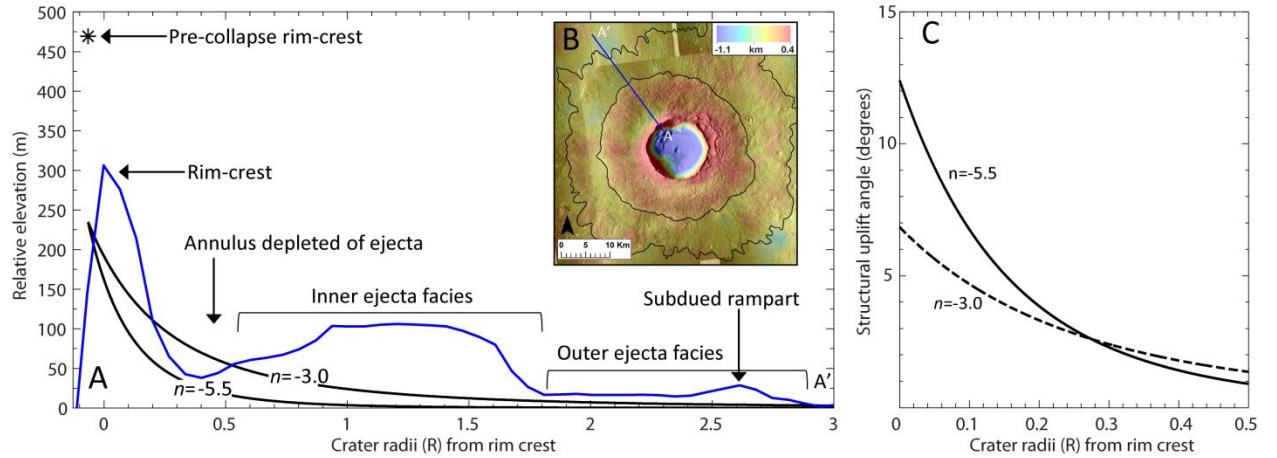
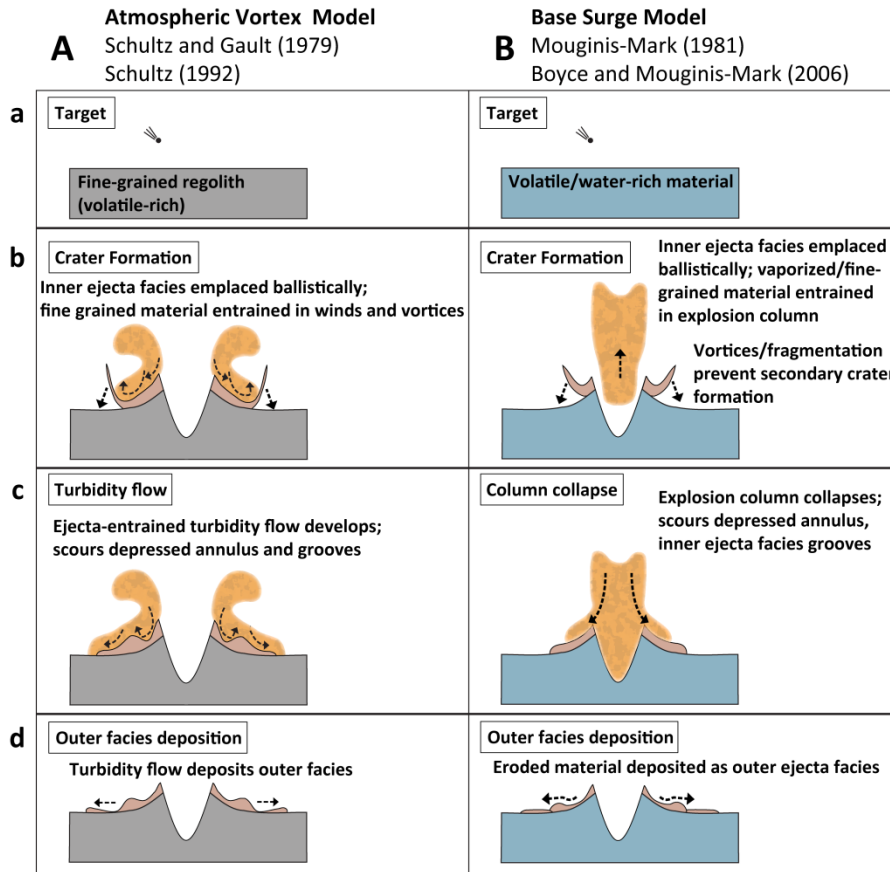


Figure 4. (A) Altimetric profile (blue) of a 12.5 km diameter DLE crater (34.7°N, 120.5°E). Value of 0 on the X axis corresponds to the location of the rim crest. Structural uplift height as a function of crater radii (R) from the rim crest (r/R) is shown as black lines from the structural-uplift height (Su) function of Stewart and Valiant (2006), where $Su \propto (r/R)^n$, with a decay exponent, $n=-3.0$ and -5.5 . (B) Map view of the crater with HRSC DEM (250 m/pixel) overlain; profile is shown as the blue line, A to A'. (C) Structural-uplift angle (θ) for $n=-5.5$ (solid) and $n=-3.0$ (dashed) for all crater diameters (angle shown is before crater collapse). Ejecta landing on the slopes of the crater rim continues to accelerate. This is the result of the interaction between the low basal μ values of the intervening icy layer and the slope of the structural uplift: in the glacial substrate model, the ejecta will accelerate when $\tan(\theta) > \mu$. The landslide begins to decelerate when $\tan(\theta) < \mu$. CTX images P19_008477_2150, G20_026186_2157, G20_025975_2135, and G22_026753_2164, HRSC DEM h2878_0000.

Atmospheric-effects models



Landslide models

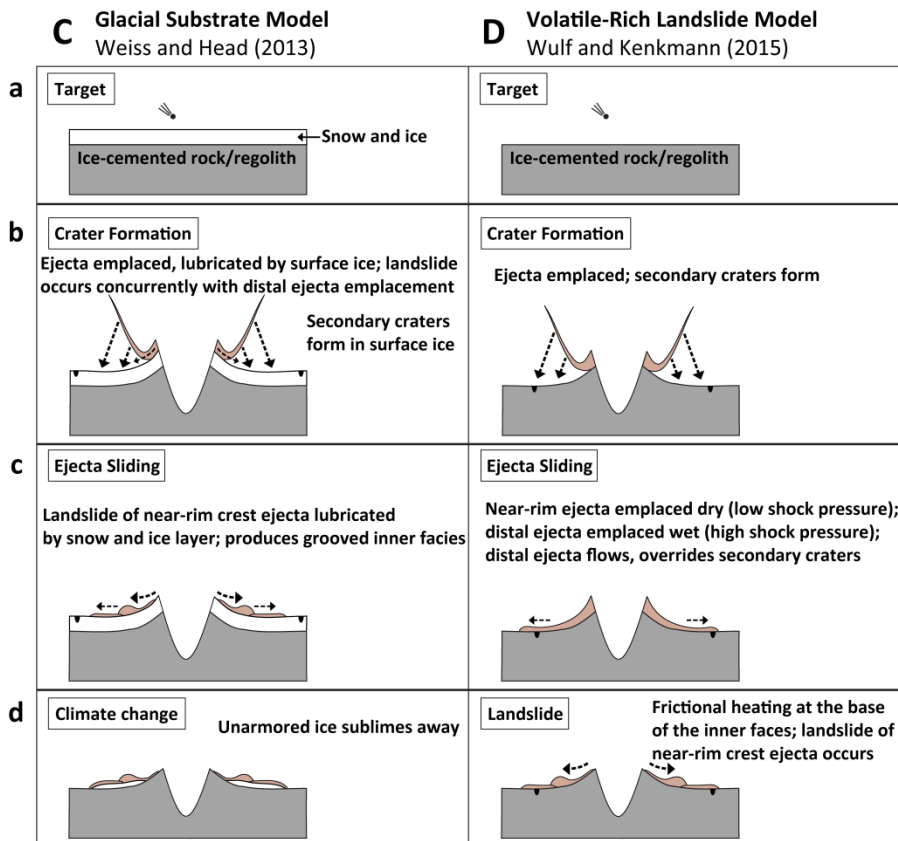


Figure 5. The four different end-to-end models of double-layered ejecta crater formation. Other hypotheses for the target structure and ejecta emplacement (Wohletz and Sheridan, 1983; Costard, 1989; Barlow and Bradley, 1990; Barlow and Perez, 2003; Osinski, 2006; Komatsu et al., 2007; Senft and Stewart, 2008; Oberbeck, 2009; Osinski et al., 2011; Jones and Osinski, 2015) have not been included because end-to-end models have not yet been formulated.

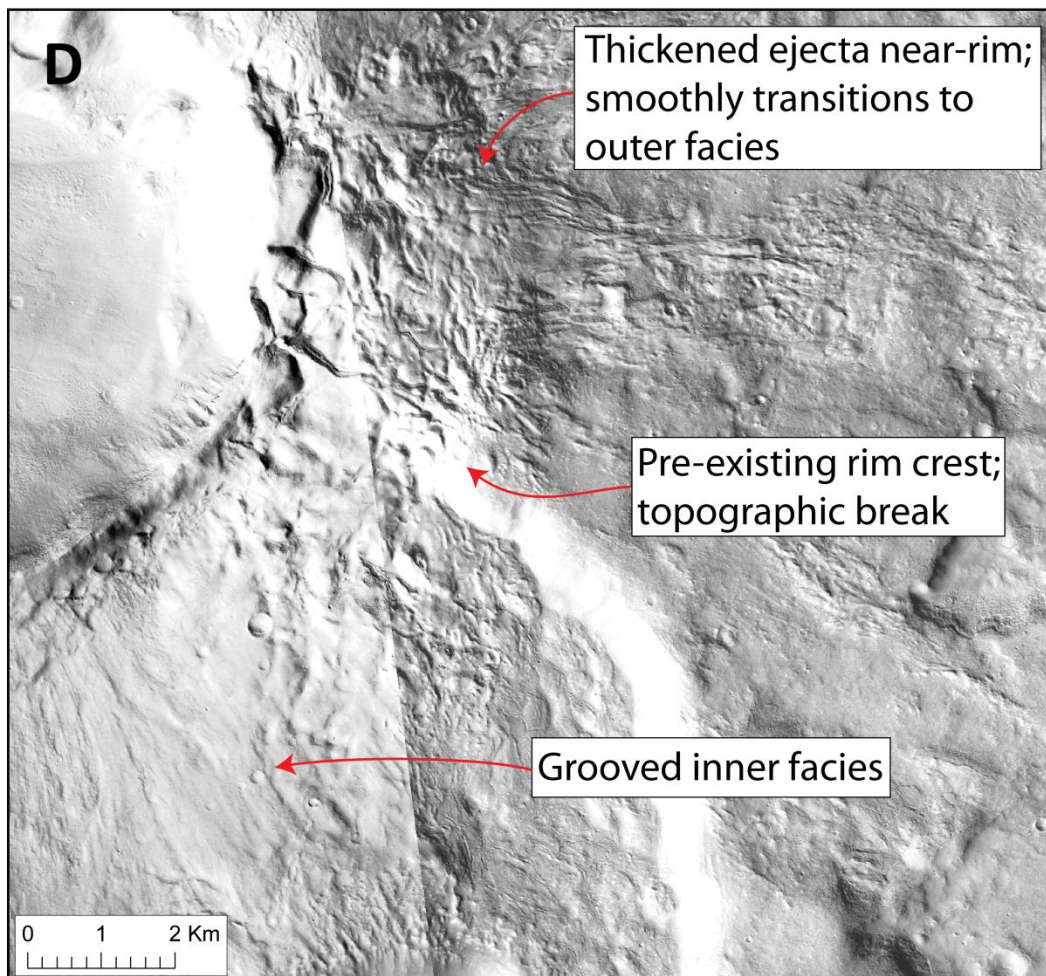
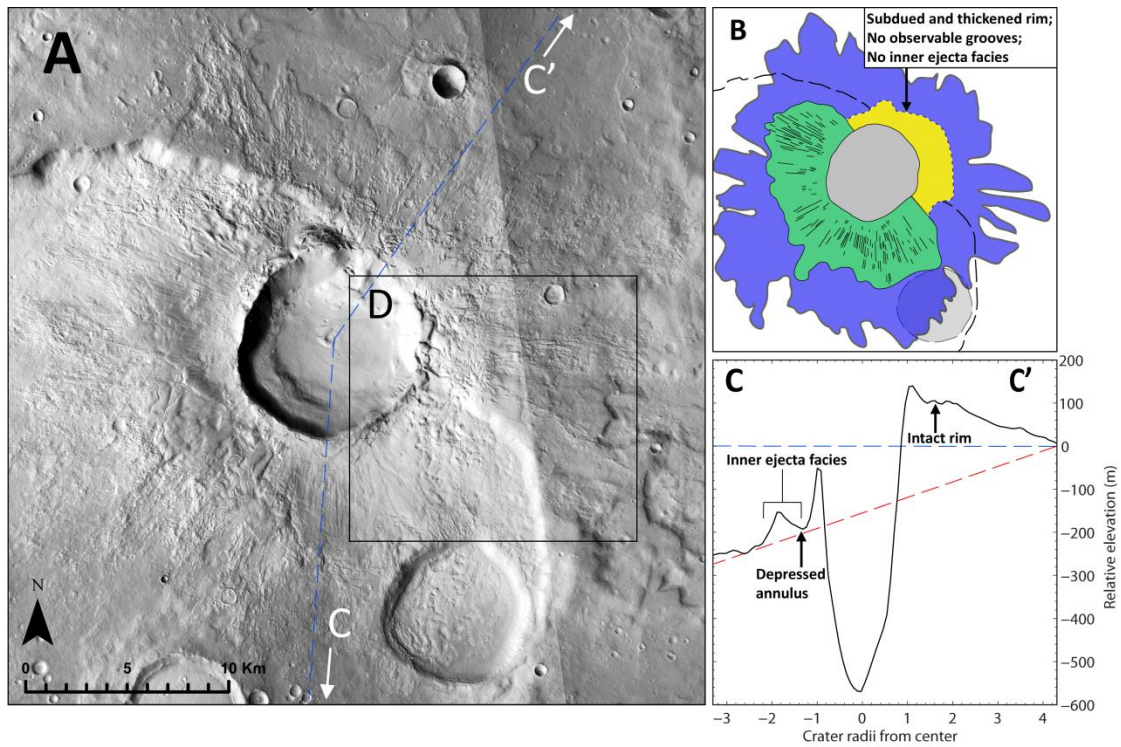


Figure 6. (A) An example of a DLE crater in which the topography of the target has inhibited the formation of the inner ejecta facies in the north-east half (42.8°N , -64.8°E). (B) Sketch map of (A) illustrating the relationship of grooves to inner ejecta facies. Craters (grey), outer ejecta facies (blue), inner facies (green), grooves (black lines), thickened near-rim ejecta (yellow), and rim crest of large craters (dashed black curve). The boundary of the thickened near-rim ejecta (yellow) is dashed to reflect that there is no clear observable contact between the near-rim ejecta and the more distal ejecta. (C) MOLA gridded topographic profile (~ 463 m/pixel; shots in Fig. S1A of supporting online material) from C to C', zero relative elevation (dashed blue line) and elevation difference (dashed red line) in (A). (D) Zoomed in black panel from (A) showing the smooth transition between the near rim-crest ejecta and the outer facies in the northeastern zone of ejecta. CTX images B18_016791_2212 and B19_017147_2230.

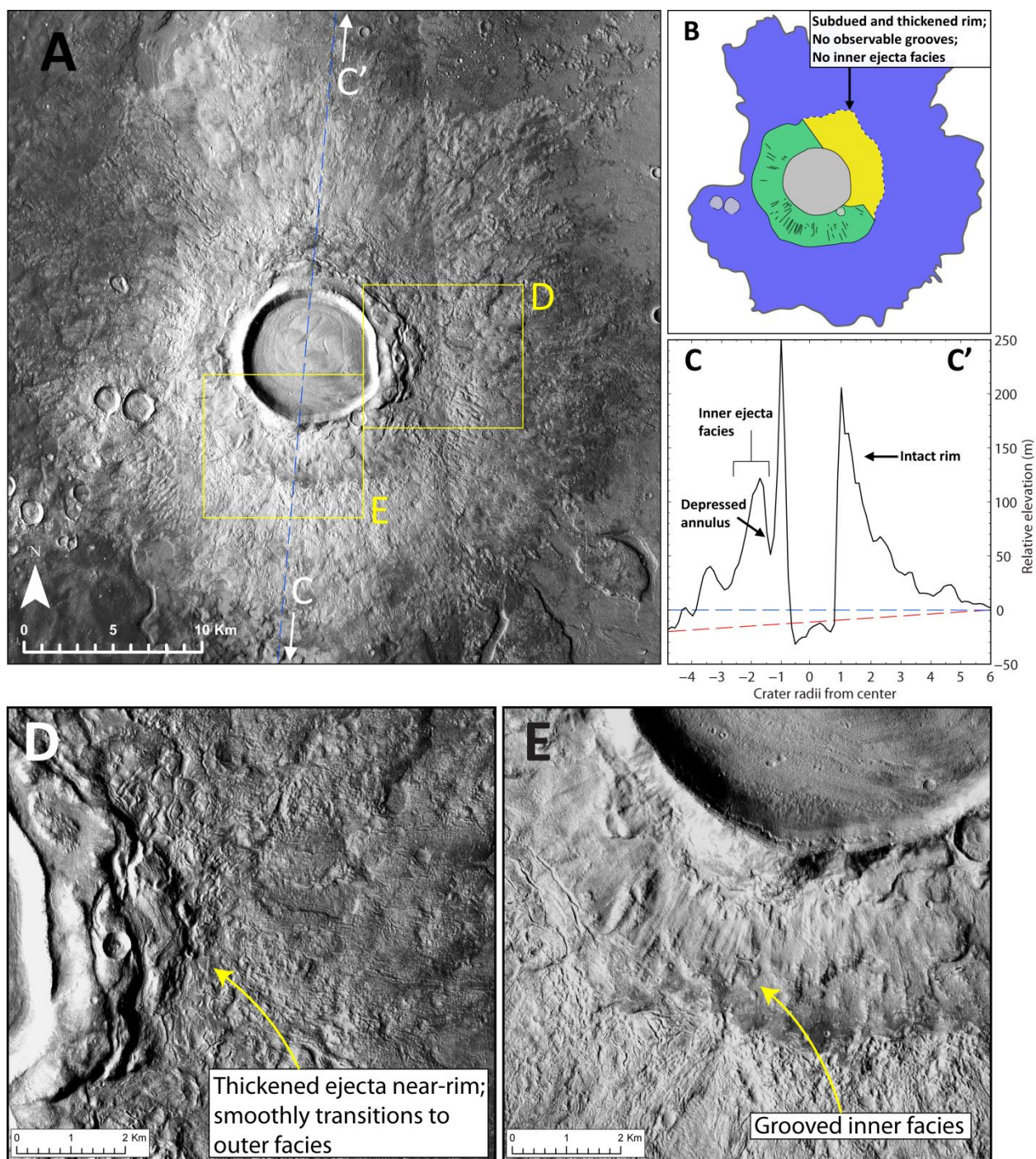


Figure 7. (A) An example of a DLE crater in which the topographic gradient has inhibited the construction of the inner facies in the northeast half (36°N , 134.4°E). (B) Sketch map of (A) illustrating the relationship of grooves to inner ejecta facies. Craters (grey), outer ejecta facies (blue), inner facies (green), grooves (black lines), thickened near-rim ejecta (yellow). The boundary of the thickened near-rim ejecta (yellow) is dashed to reflect that there is no clear observable contact between the near-rim ejecta and the more distal ejecta. (C) MOLA gridded topographic profile from C to C', zero elevation relative to surrounding terrain (dashed blue line) and elevation difference (dashed red line) in (A);

shots in Fig. S1B of supporting online material. (D) Zoomed in yellow panel from (A) showing the smooth transition between the near rim-crest ejecta and the outer facies in the upslope northeastern zone of ejecta. (E) Zoomed in yellow panel from (A) showing the grooved inner present facies in the downslope southwestern zone of ejecta. CTX images B19_016942_2170, P20_008727_2155, and P02_001831_2150.

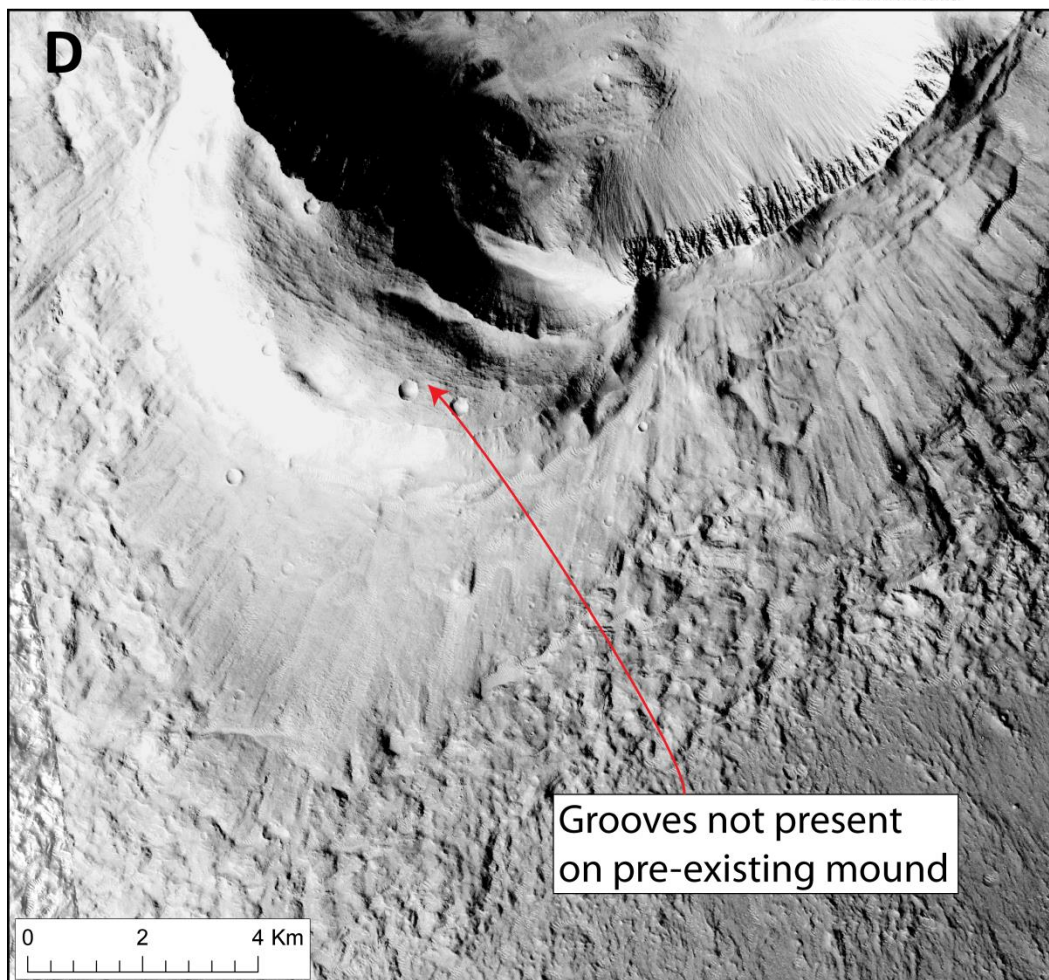
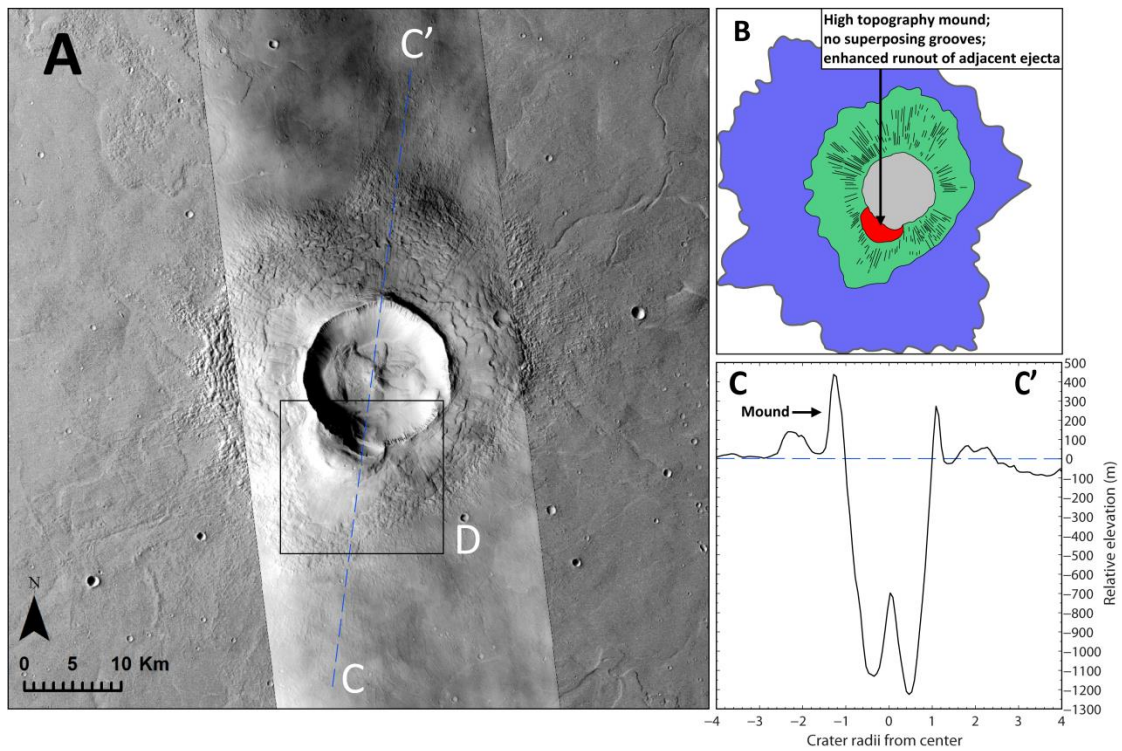


Figure 8. (A) An example of a DLE crater in which the impact into a high-topography mound in the southern rim has displaced the inner facies by a distance equivalent to the mound width (28.0°N, 116.7°E). No grooves are observed on the top of the mound. (B) Sketch map of (A) illustrating the relationship of grooves to inner ejecta facies and mound. Crater (grey), outer ejecta facies (blue), inner facies (green), grooves (black lines; some visible at lower map scales and different image stretches), high topography mound (red). (C) MOLA gridded topographic profile from C to C', zero elevation relative to surrounding terrain (dashed blue line) in (A); shots in Fig. S1C of supporting online material. (D) Zoomed in view of the black panel in (A) showing the lack of grooves on the surface of the mound, as would be expected if they formed through an atmospheric scouring effect. (A) CTX images P15_007027_2099 and P15_006816_2100, and HRSC image h2922_0000.

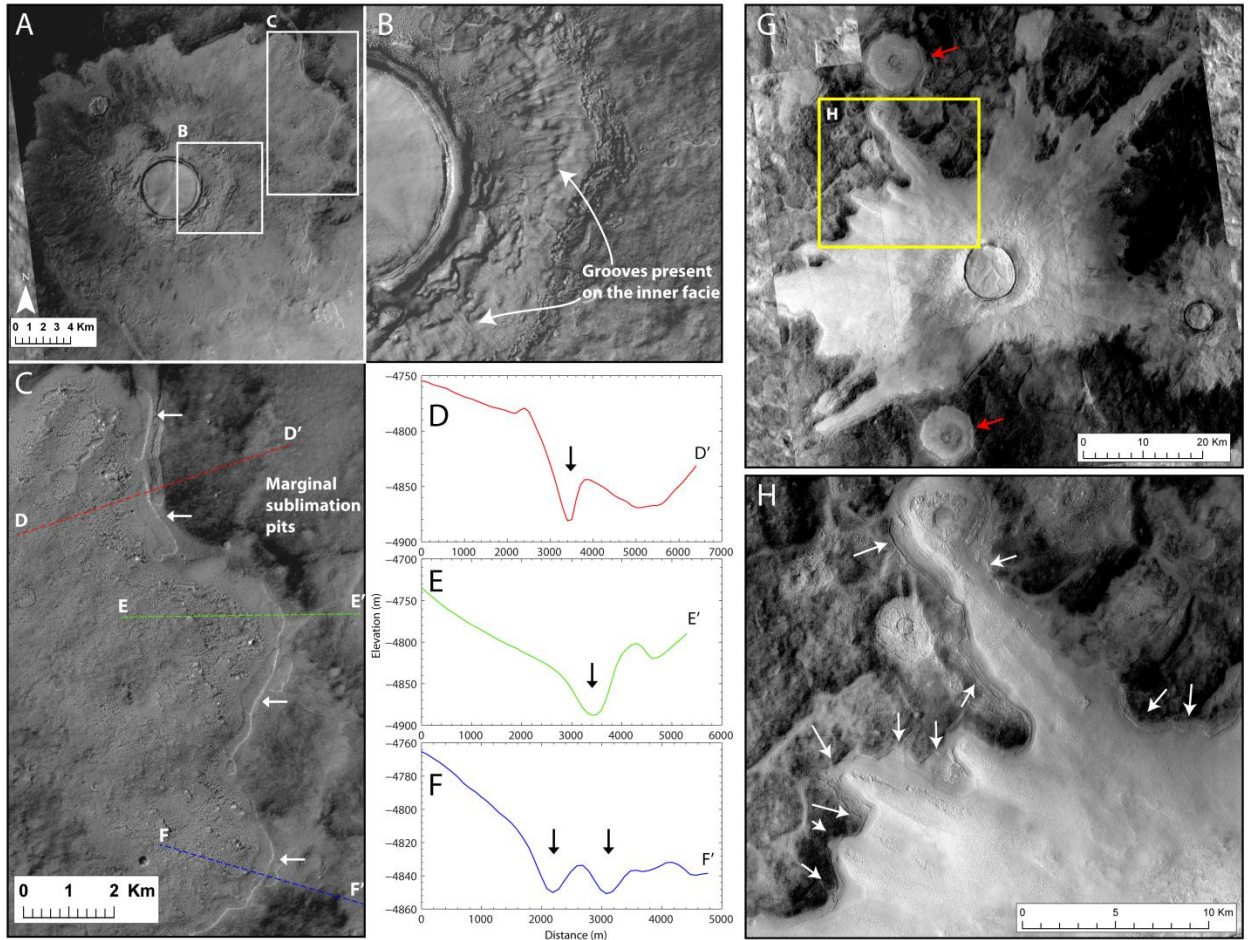


Figure 9. Examples of DLE craters which exhibit marginal sublimation pits. (A) This 4.8 km diameter crater in Utopia Planitia (43.2°N, 99.4°E) is interpreted to be a DLE crater (A, B), and exhibits sublimation pits (white arrows in C) marginal to the outer ejecta facies, much like those observed along the pedestal margins of pedestal craters (red arrows in panel G) (Kadish et al., 2008). HRSC DEM topographic profiles are shown in (D, E, and F) with sublimation pits (black arrows). (G) Another example of a DLE crater (9 km in diameter; 44.86°N, 99.95°E) with marginal sublimation pits (H) is shown. Fig. 16A shows another example of a DLE crater with marginal sublimation pits. These marginal sublimation pits are interpreted to indicate the presence of an underlying icy substrate (Kadish et al., 2008). (CTX image G21_026490_2229; HRSC DEM h1398_0000).

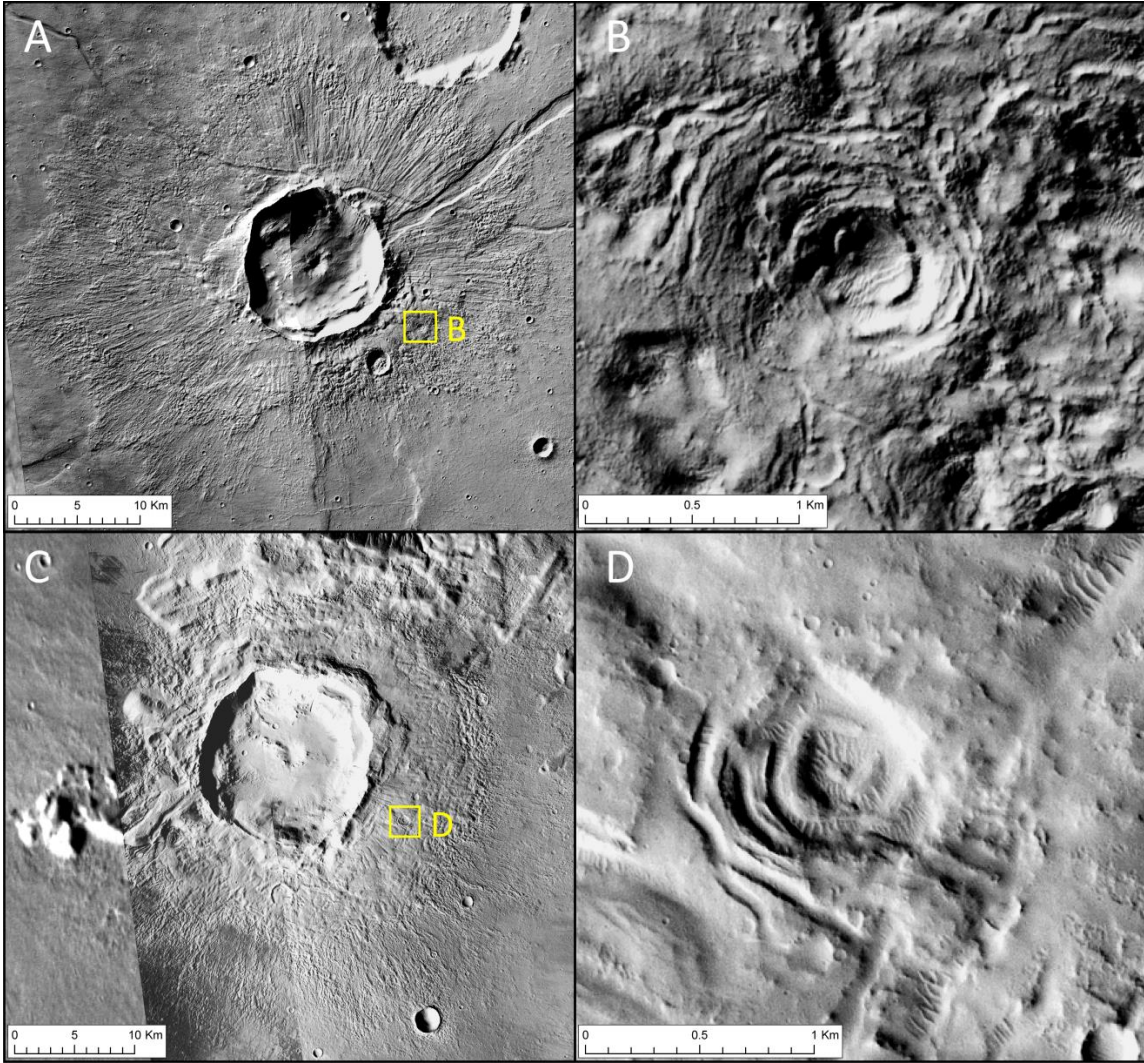


Figure 10. Examples of two DLE craters with ring-mold craters (RMCs) superposed on the ejecta facies identified by Levy et al. (2016). (A) 11.5 km DLE crater (36.5°S, -145.7°E). (B) ~700 m RMC located on the inner ejecta facies of the DLE crater in (A). (C) 15 km DLE crater (30.4°N, -86.3°E). (D) ~800 m RMC located on the inner ejecta facies of the DLE crater in (C). RMC craters are interpreted as impact craters that formed when the projectile impacted a debris-covered ice substrate (Kress and Head, 2008; Baker et al., 2010; Head and Weiss, 2014). Their presence on the ejecta of these craters suggests that massive ice was present below the ejecta facies at the time the RMCs formed. CTX images for (A): F01_036178_1423, P13_006259_1437. CTX images for (B): P15_006863_2129, D22_035780_2101.

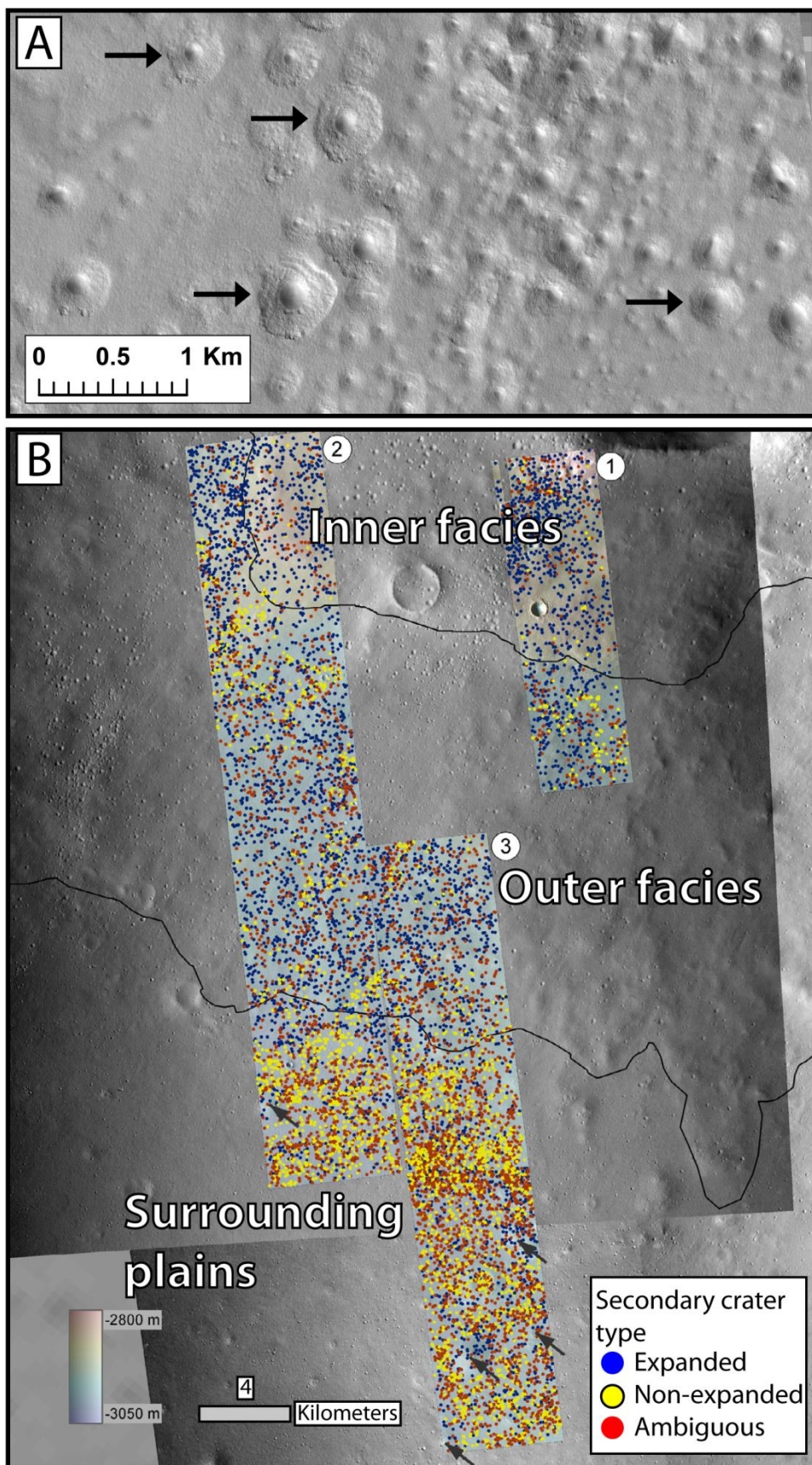


Figure 11. (A) Example of “expanded” secondary craters characterized by Viola et al. (2015) (black arrows) (52.7°N, 217.7°E, HiRISE image ESP_028688_2330). (B) The secondary craters mapped by Viola et al. (2017) associated with this 15 km diameter DLE crater (50.4°N, 219.6°E). Panel (B) is adapted from Viola et al. (2017). Expanded secondary craters are frequently associated with the ejecta of DLE craters (Viola et al., 2015, 2017) and are interpreted to result from secondary impact into buried surface ice (Viola et al., 2015, 2017).

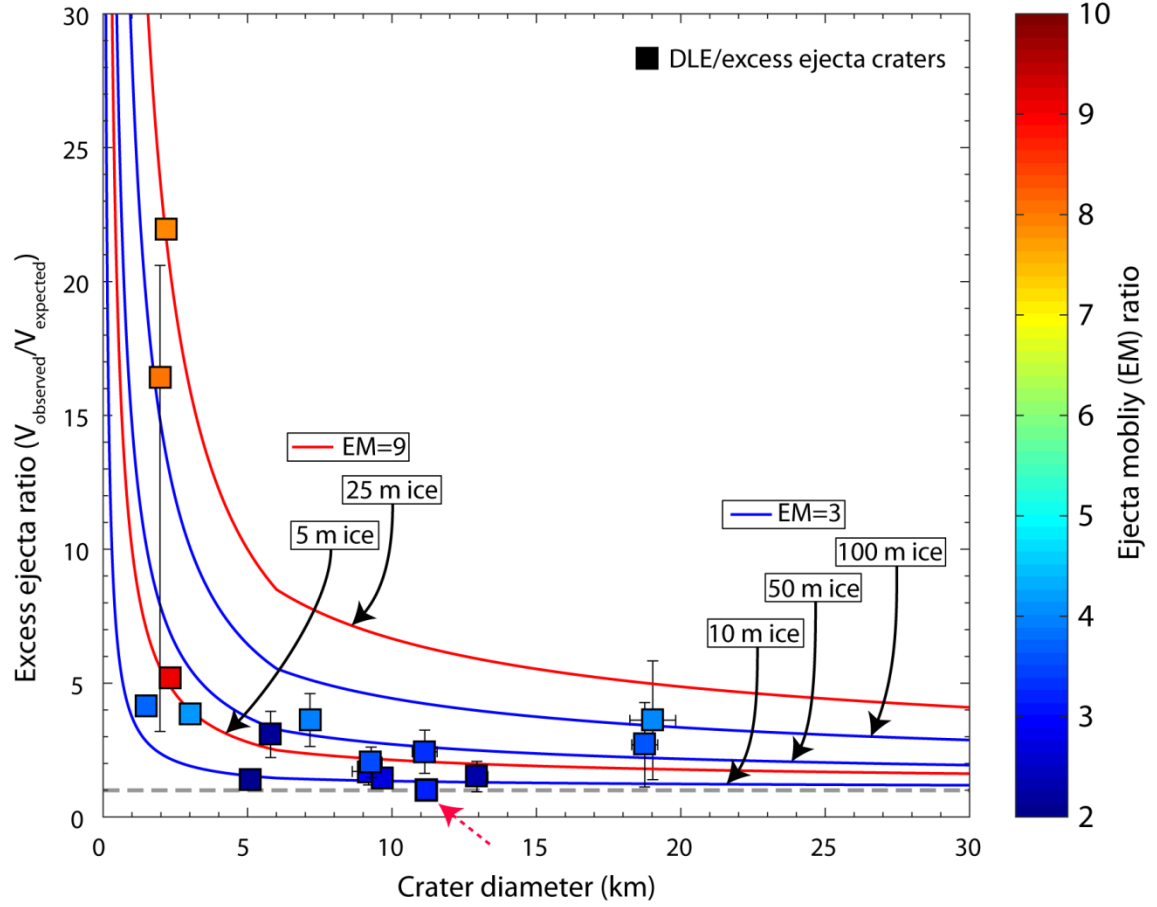


Figure 12. Excess ejecta ratio is shown as the measured ejecta volume (V_{observed}) divided by the expected ejecta volume (V_{expected}) as a function of crater diameter for 16 DLE craters from Black and Stewart (2008), Kadish and Head (2011), Schaefer et al. (2011), and Steinheim crater from Wulf and Kenkmann (2015) (marked with the dashed red arrow; see Section 3.4 for explanation). The color of these markers corresponds to their ejecta mobility (EM) ratio shown in the colorbar. The model lines show the predicted excess ejecta ratio for ice thicknesses between 5 m and 100 m and an EM ratio of 3 (blue lines) and 9 (red lines). The dashed grey line marks the excess ejecta ratio of 1, where the observed volume of ejecta matches that expected without any underlying ice deposits. The classification of a number of DLE craters as excess ejecta craters is interpreted to be due to the underlying ice substrate, which well reproduces the observed trend of excess ejecta ratio versus crater diameter for these craters, and supports the suggestion that an ice layer is present beneath the ejecta facies of DLE craters.

Ries crater, Earth

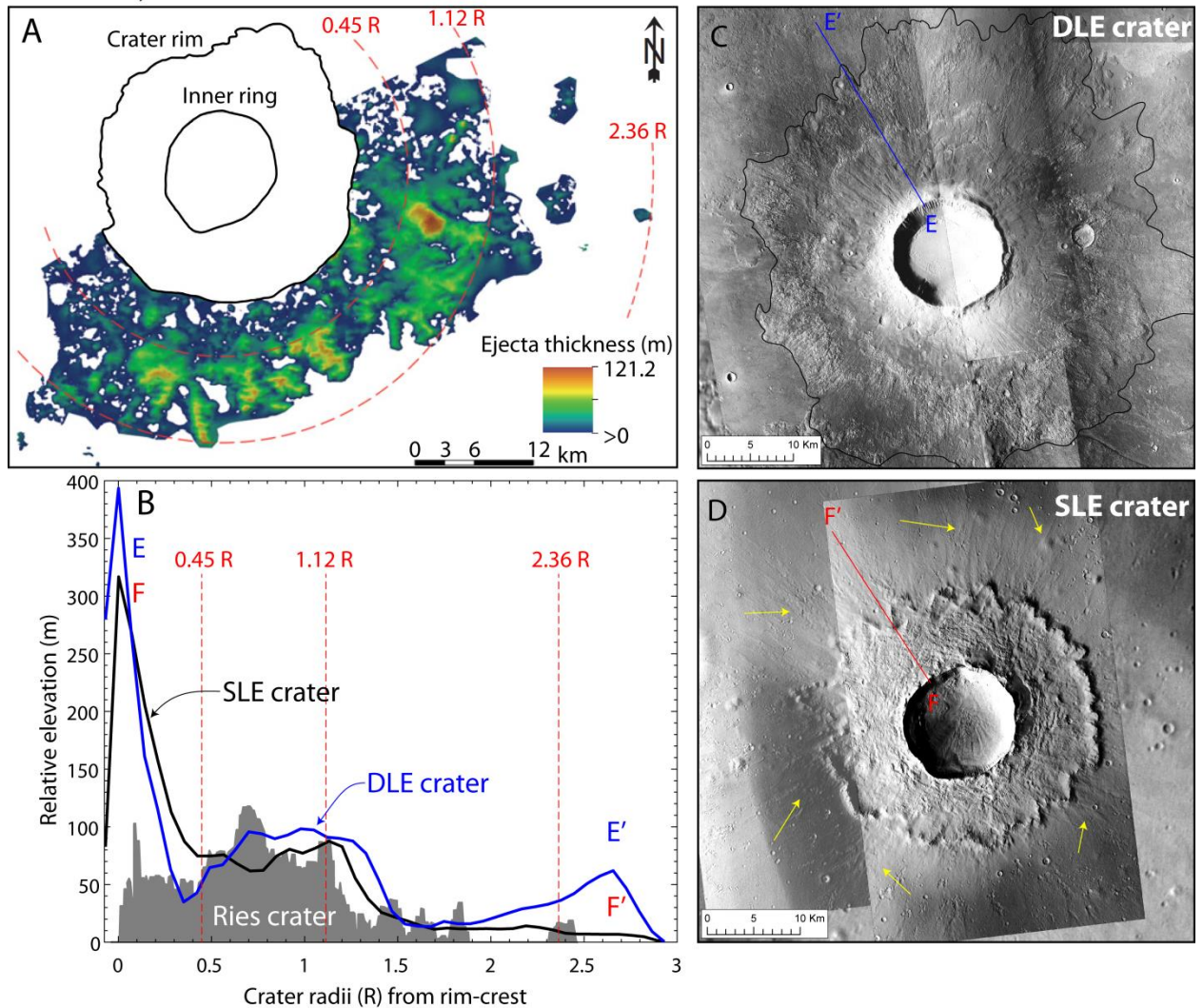


Figure 13. Comparison between martian layered ejecta craters and the ~26 km diameter terrestrial Ries crater. (A) Ries crater Bunte Breccia interpolated ejecta thickness adapted from Fig. 2 of Sturm et al. (2013). Dashed red lines show distances of 0.45 crater radii (R) from the rim (the distance of the depressed annulus of ejecta surrounding the rim), 1.12 R (the distance of most of the ejecta), and 2.36 R (the maximum distance of Ries ejecta) (Sturm et al., 2013). (B) Comparison between the ejecta topography of a martian DLE crater (blue line; E to E' in panel C), a martian SLE crater (black line; F to F' in panel D), and the ejecta thickness of the terrestrial Ries crater from Sturm et al. (2013) (shaded grey area). The ejecta topography of the Ries crater (data from Sturm et al., 2013) was found as the minimum ejecta thickness from 12 different topographic profiles across the Bunte Breccia. The ejecta topography of Ries crater appears to be more similar to SLE craters than DLE craters, and so we do not consider target properties of Ries crater to be representative of martian DLE craters (e.g., Sturm et al., 2013; Wulf and Kenkmann, 2015; Kenkmann et al., 2015). (C) 12.7 km diameter martian double-layered ejecta crater (34.6°N, 125.6°E). Blue line is the MOLA gridded topographic profile in panel (B). (D) 12 km diameter martian single-layered ejecta crater (36.0°N, 80.5°E). Red

line is the MOLA gridded topographic profile in panel (B). Yellow arrows indicate thin distal ejecta beyond the primary single ejecta facies. Four other SLE craters with thin distal ejecta deposits can be found at: 19.6°N, 143.3°E, and 19.5°N, 141.2°E, and 16.8°N, 156.3°E, and 9.4°S, 107.8°E. CTX images in (A): P17_007554_2154, B18_016639_2159, B18_016639_2159, D02_027887_2162, P18_008200_2142. CTX images in (B): B04_011287_2163, P18_008083_2177.

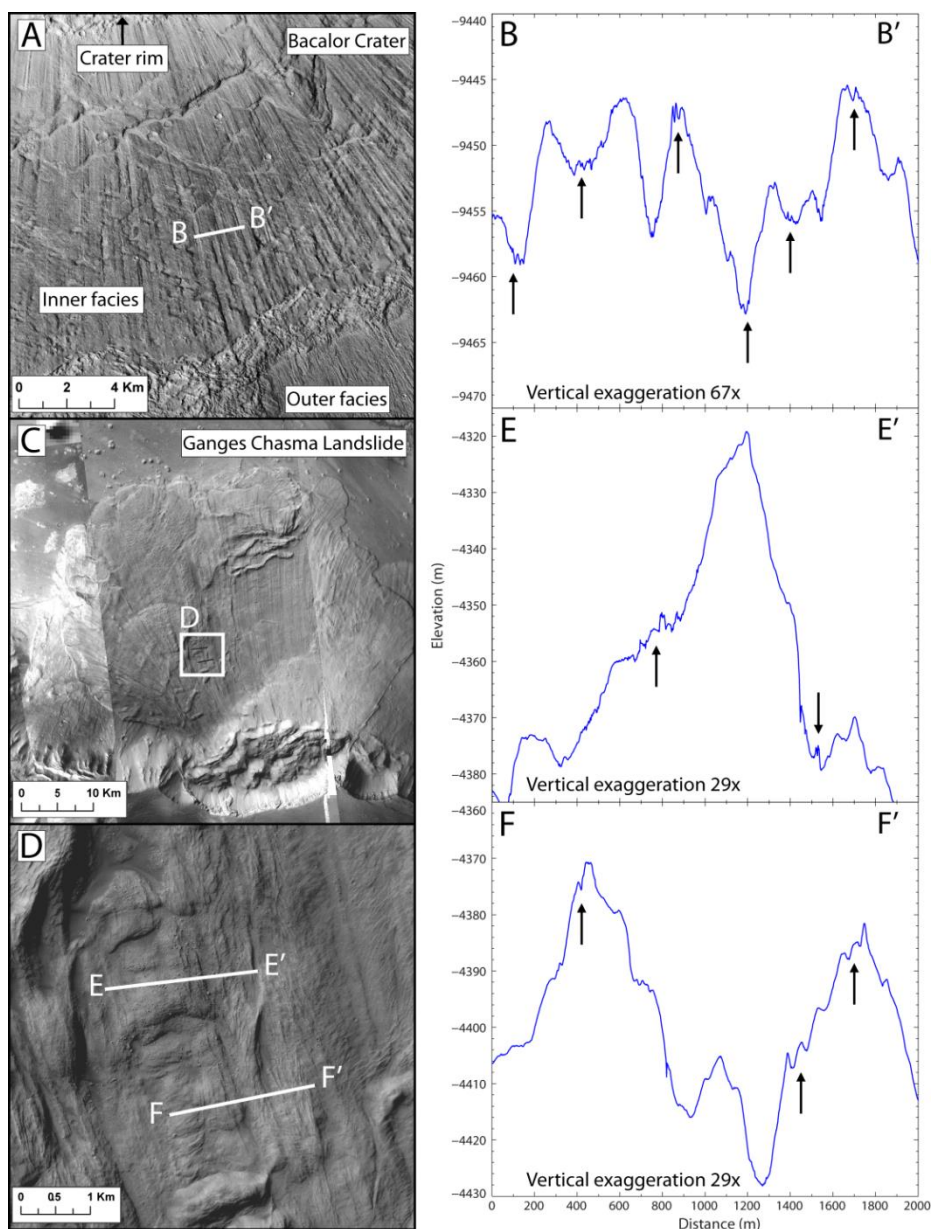


Figure 14. Landslide groove morphometric comparison between a DLE crater (A and B), and a martian landslide (C, D, E, and F) using HiRISE DEMs; the superposing grooves (which are ~1-2 m in depth) are resolvable in HiRISE DEMs, which have vertical uncertainties on the order of tens of centimeters. Both the DLE crater ejecta and the landslide surface display a bimodal groove wavelength distribution; small grooves superposing larger grooves are denoted by black arrows. Although the vertical exaggeration differs between the landslide example and the DLE crater, the scale of the grooves (~5-30 m) and superposing grooves (~1-2 m) are comparable, and differences in thickness are likely to explain the minor variation. CTX images P16_007462_2133, P21_009248_1726, B10_013718_1714, P02_001996_1725, B11_013784_1715, B17_016263_1713, HiRISE DEMs PSP_007462_2130_PSP_006750_2130, ESP_024333_1715_ESP_025401_1715.

Flow feature	μ	Fr	Note	Grooves	Reference
DLE craters	0.01-0.1	0.52-11.6		x	This study
SLE/MLE craters	0.1-0.6	0.52-11.6		Frequently	This study
Sherman, 1964	0.05	$5.87^{+23.5}_{-3.30}$		x	Shreve (1966), McSavaney (1975), Sosio et al. (2012a)
Punta Thurweiser, 2004	0.05	$8.28^{+2.10}_{-2.57}$		x	Sosio et al. (2008)
Black rapids West, 2002	0.16	$9.09^{+1.07}_{-0.79}$		x	Jibson et al. (2006)
Black Rapids East, 2002	0.16	$9.09^{+1.07}_{-0.79}$		x	Jibson et al. (2006)
Mt Iliamna, 2003	0.12	$12.51^{+2.81}_{-1.68}$		x	Schneider et al. (2010)
Mt Cook, 1991	0.13	$15.71^{+1.85}_{-1.37}$		x	Schneider et al. (2010)
Kolka Glacier, 2002	0.05	$5.64^{+23.1}_{-4.00}$		x	Haerberli et al. (2004), Huggel et al. (2005), Zaporozhchenko (2006), McDougall (2006)
Mt. Steller, 2005	0.09	$9.03^{+1.07}_{-0.79}$		x	Huggel et al. (2008)
Mt. Munday, 1997	0.07	$2.87^{+13.8}_{-0.00}$		x	Delaney and Evans (2014), Sosio et al. (2012a)
McGinnis Peak, 2002	0.18*	$10.61^{+1.58}_{-1.58}$	min	x	Jibson et al. (2006)
Hast, Mt. Cook, 2013	0.18*	$8.78^{+11.5}_{-3.72}$	avg	x	Hancox and Thomson. (2013)
Vampire, 2008	0.39*	$10.55^{+6.50}_{-4.16}$	avg	x	Cox et al. (2008)
Shivulich, 1965	0.13	$7.22^{+0.85}_{-0.63}$	max	x	Belousov (1995), Belousov et al. (1999) Sosio et al. (2012b)
Llullaillaco	0.07	$4.54^{+0.50}_{-3.23}$	max	x	Richards and Villeneuve (2001), Sosio et al. (2012b)
Mt. Steele, 2007	0.05	$5.31^{+17.0}_{-0.28}$		x	Lipovsky et al. (2008)
Little Tahoma Peak, 1963	0.21	$10.23^{+4.88}_{-4.88}$		x	Sheridan et al. (2005)
Socompa Proximal	0.04	$4.52^{+1.87}_{-3.75}$		x	Kelfoun and Druitt (2005), Kelfoun et al., (2008), Davies et al. (2010)
Mt. Meager, 2010	0.02	$2.97^{+10.1}_{-2.54}$		x	Guthrie et al. (2012)

Tsar Mountain, 2000	0.11	$7.56^{+6.80}_{-5.34}$	avg	x	Jiskoot (2011)
Mt. Fletcher, 1992	0.19	$8.08^{+30.2}_{-4.78}$		x	McSavaney (2002), Allen et al. (2009)
Blackhawk	0.23*	$2.89^{+0.65}_{-0.85}$	min		Shreve (1968)
Acheron	0.51	$1.46^{+2.21}_{-0.09}$			Smith et al. (2006)
Pink Mountain, 2002	0.42*	$3.39^{+6.19}_{-2.67}$			Geertsema et al. (2006)
Nagano, 2011	0.26*	$1.43^{+0.57}_{-0.00}$	max		Yamasaki et al. (2014)
Niumiangou Creek, 2008	0.25*	$3.44^{+1.70}_{-1.94}$	min		Zhang et al. (2011)
Wulong, 2009	0.26*	$2.48^{+0.00}_{-0.73}$			Xu et al. (2010)
Frank, 1903	0.14	$3.01^{+2.26}_{-1.03}$			Poisel et al. (2008)
Six des Eaux Froides, 1945	0.13	$3.59^{+1.66}_{-0.17}$	max		Hungr and McDougall (2009)
Socompa distal reflected	0.04	$0.82^{+1.73}_{-0.49}$			Kelfoun and Druitt (2005), Kelfoun et al., (2008), Davies et al. (2010)
Little Salmon Lake, 2008	0.58	$7.90^{+12.5}_{-4.37}$	min		Brideau et al. (2009)
Ganges Chasma, Mars	0.07*	$2.54^{+0.57}_{-0.78}$		x	Quantin (2004)
Ius east, Mars	0.05*	$1.61^{+0.40}_{-0.33}$		x	Lucchitta (1987)
Melas, Mars	0.08	$3.22^{+1.75}_{-1.48}$		x	Lucchitta (1987), Mazzanti et al. (2016)
Ius Center, Mars	0.12*	$0.99^{+0.09}_{-0.04}$			Lucchitta (1987)
Ophir Labus	0.08	$1.81^{+0.22}_{-0.36}$			Mazzanti et al. (2016)
Granular flow experiment	0.87	$8.30^{+1.45}_{-1.1}$		x	Forterre and Pouliquen (2001)
Granular flow experiment	0.29-0.36	2.21-4.14			Louge and Keast (2001)
Granular flow experiment	0.36-0.67	0.12-5.27			Ancey (2001)

Table 1. Ejecta, landslide, avalanche, and granular flow features from Fig. 15. The Note column indicates whether the Froude number (Fr) is a minimum, average, or maximum bound based on the velocity estimates, otherwise, the Fr range for each flow feature corresponds to the average, plus or minus some value corresponding to the maximum and minimum velocities. The * symbol denotes cases where friction coefficients (μ) were

measured in this study. Thickness and velocity data can be found in Table S1 of the supporting online material.

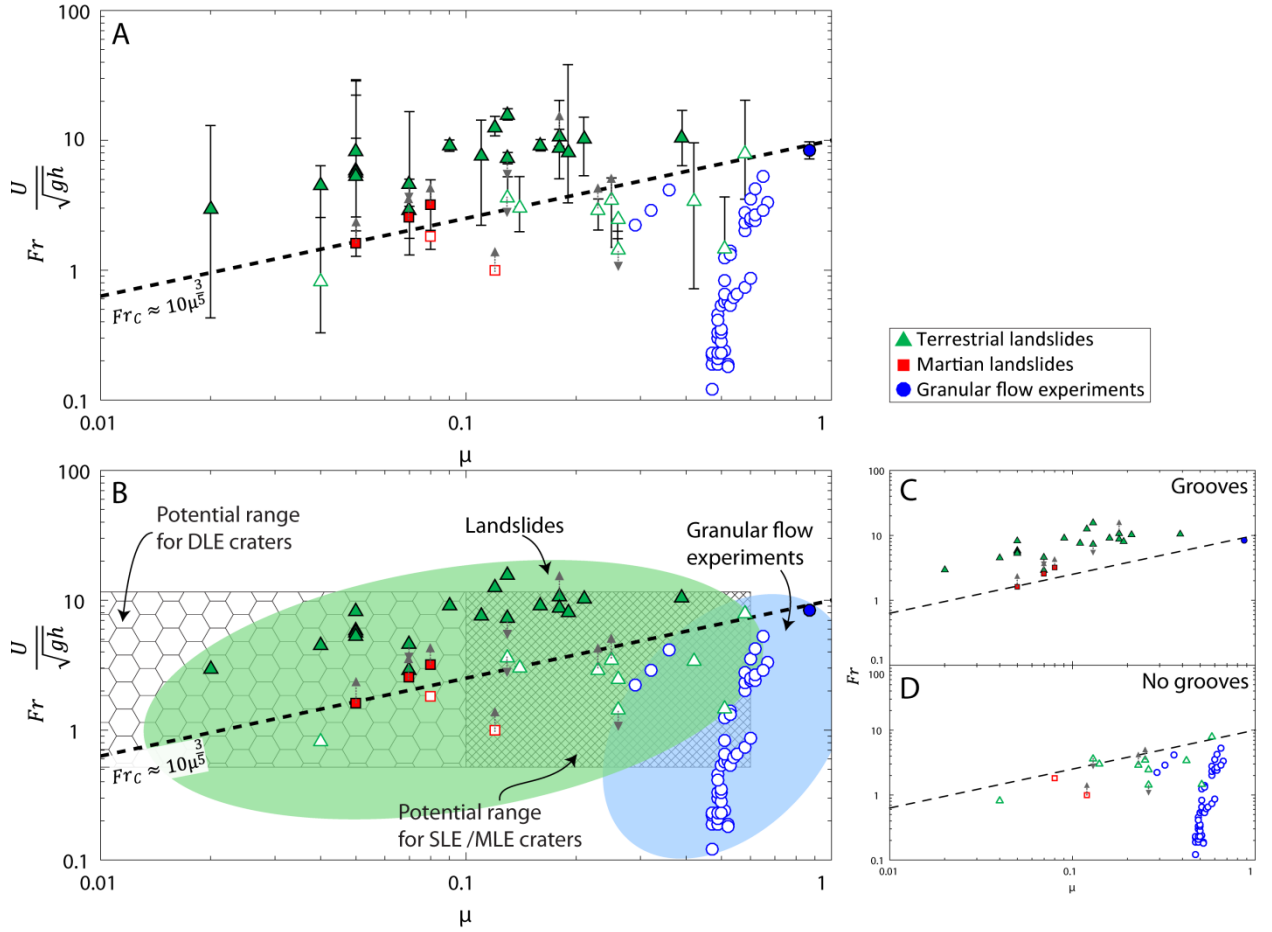


Figure 15. Kinematic and frictional comparison between landslides and martian layered ejecta deposits. (A) Froude number versus friction coefficient (μ) for a number of terrestrial landslides (green triangles), martian landslides (red squares), and granular flow experiments (blue circles) from Table 1 (also see Table S1 in supporting online material). Filled markers exhibit longitudinal grooves (C), while open markers do not (D). Dashed line is the critical Froude number (Fr_c) above which flow features generally exhibit grooves. Each point represents the average value; the error bars represent the maximum range in Fr due to the various speed and thickness estimates (and not the error in the measurement). Dashed gray arrows indicate markers that are minimum or maximum Fr bounds, and the direction the marker would move if the minimum or maximum derived velocities were below/above the average velocity, respectively. (B) Same as (A) but with the flow features grouped by origin. The honeycomb-patterned region is the predicted range of DLE crater Fr - μ values for an ejecta speed range of 10-100 m/s, ejecta thickness between 20-100 m, and μ between 0.01 and 0.1 (Weiss and Head, 2014). The cross-hatched-patterned region is the predicted range of SLE and MLE crater Fr - μ values for μ between 0.1 and 0.6 (Weiss and Head, 2014). The shaded green region encompasses the landslide data, while the shaded blue region encompasses the data from the granular flow laboratory experiments. DLE crater ejecta is kinematically and frictionally comparable to terrestrial landslides with grooves (most of which ran out on glaciers), supporting the hypothesis (Weiss and Head, 2013) that the ejecta slid on a surface ice layer and that the

grooves formed through a manner analogous to landslides. DLE craters are generally predicted to be above the Fr_c , while SLE/MLE craters are generally predicted to be below the Fr_c , consistent with the large abundance of grooves reported for DLE craters but lower frequency of grooves reported for SLE and MLE craters.

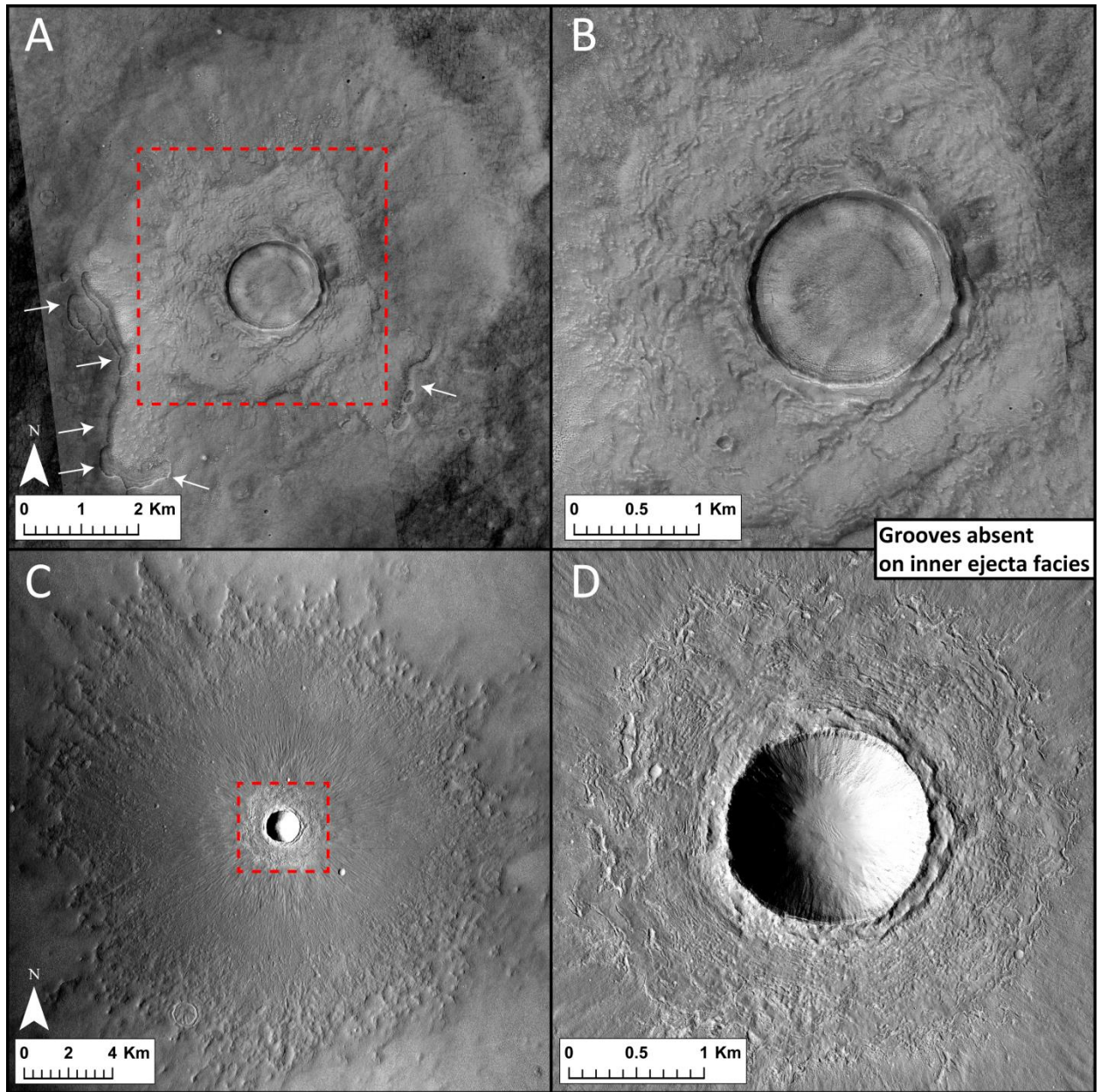


Figure 16. Examples of DLE craters surveyed which exhibit tensional features (transverse ridges and fissures) but not grooves on the inner ejecta facies. (A) 1.6 km diameter DLE crater (42.9°N, 96.3°E). HiRISE image ESP_035048_2235 overlain on CTX image B18_016495_2244. Dashed red box denotes zoomed-in area in (B). White arrows indicate marginal sublimation pits, indicative of sublimation of massive ice at the ejecta margins (Section 3.1). (C) 1.6 km diameter DLE crater (39.4°N, 75.2°E). CTX image G23_027173_2212. Dashed red box denotes zoomed in area in (D); HiRISE image PSP_010611_2210.

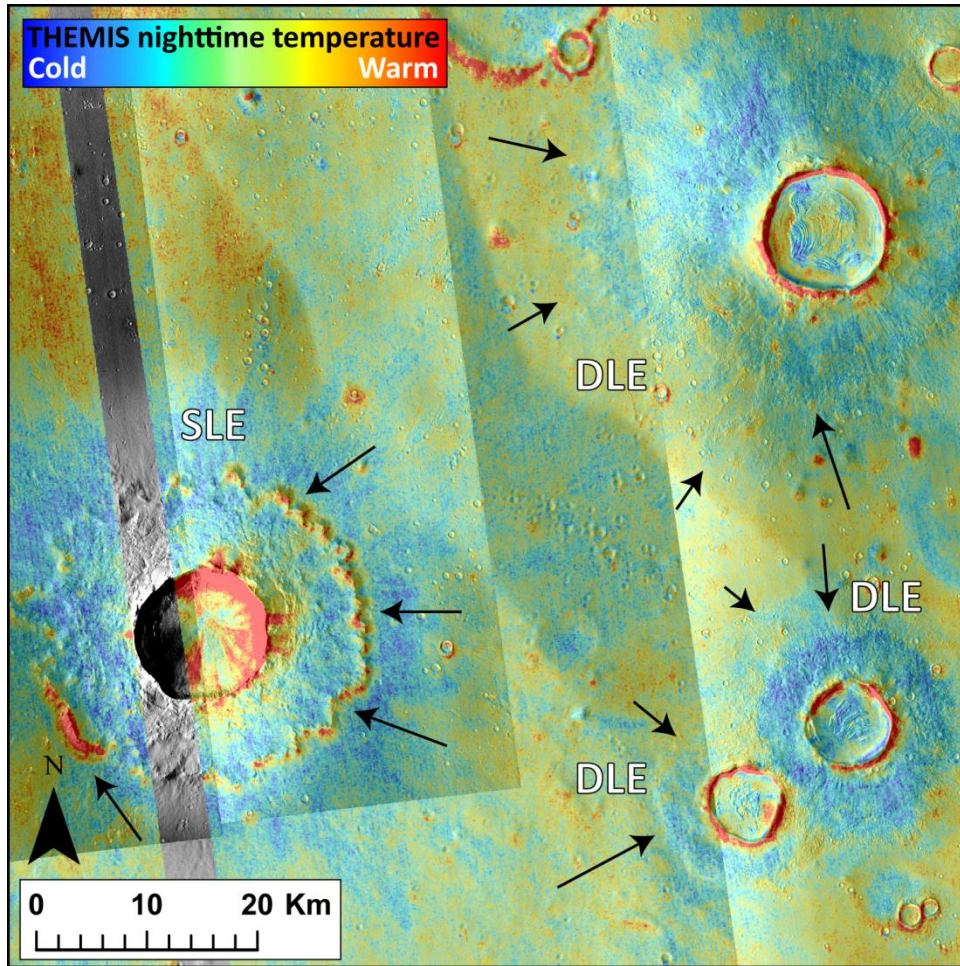


Figure 17. Colorized Thermal Emission Imaging System (THEMIS) nighttime infrared imagery (Edwards et al., 2011) overlain on CTX images. The single-layered ejecta crater (from Fig. 13D) qualitatively exhibits higher thermal inertia on the distal ejecta ramparts than adjacent DLE craters (black arrows) (35.6°N, 80.5°E). This indicates the presence of larger blocks present in the distal ramparts of the SLE crater ejecta relative to the adjacent DLE crater ejecta. DLE and SLE craters may be observed side-by-side because SLE craters form on the rocky surface during periods of low obliquity, while DLE craters are interpreted to form during periods of higher obliquity, when surface snow/ice deposits are present (Weiss and Head, 2014). Based on the concentric crater fill infilling the DLE craters and lack of crater fill within the SLE crater, we interpret the SLE crater to be younger, and to have formed after the glacial period during which the DLE craters formed. CTX images P15_007081_2183, P16_007371_2175, P17_007872_2151, and P18_008083_2177.

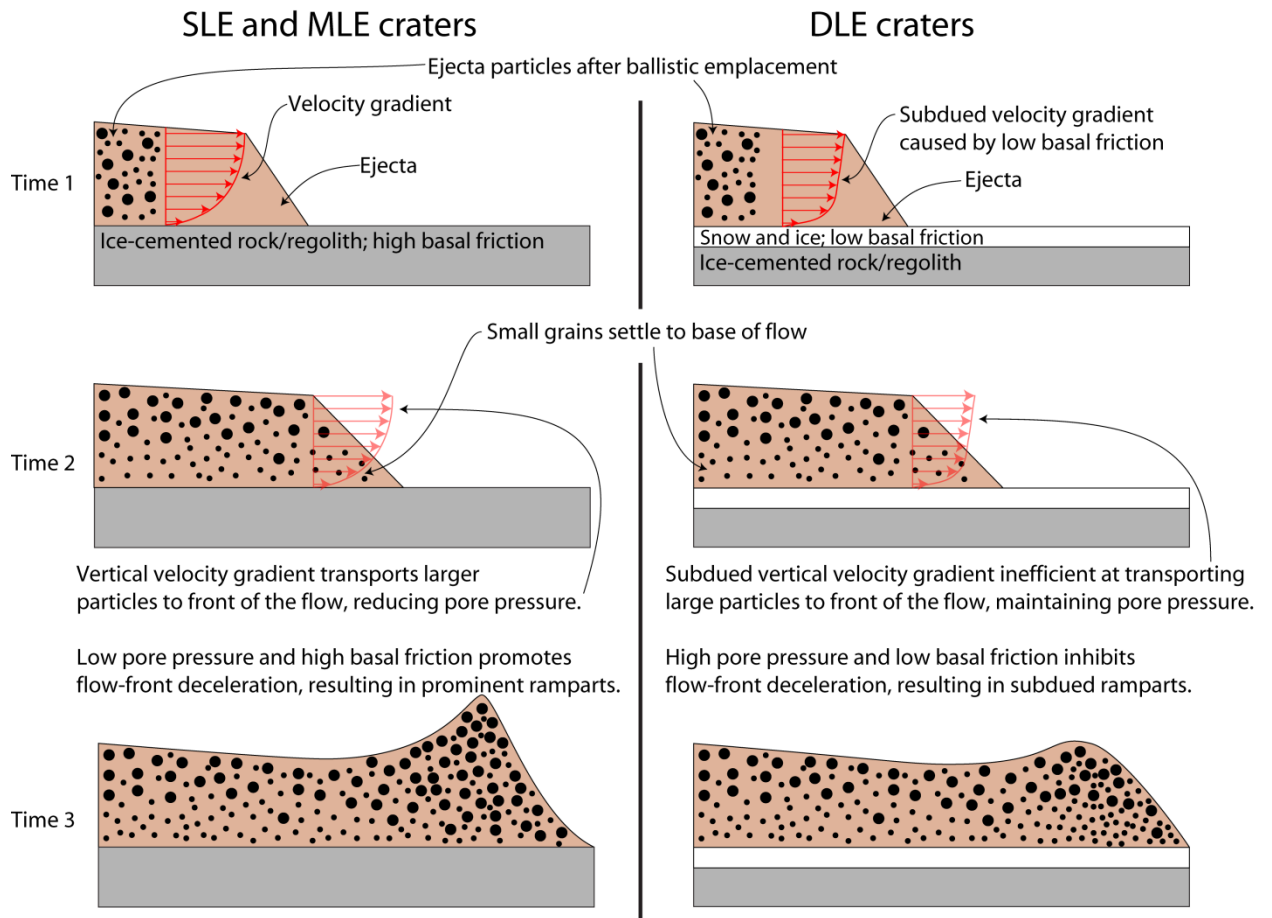


Figure 18. Proposed flow characteristics for the distal ejecta of single- and multiple-layered ejecta craters (SLE/MLE) (left panels) and DLE craters (right panels). Top panels show the initial flow conditions following ballistic emplacement. The icy substrate underlying the DLE crater ejecta leads to low basal friction values and a subdued vertical velocity gradient compared with SLE and MLE craters. The middle panels show that in both cases, the small particles are predicted to settle to the base of the flow, resulting in a high concentration of large particles at the shallower depths of the flow. Due to the higher flow velocities near the top of the flow predicted for SLE and MLE craters, the large particles are transported to the flow front more efficiently through kinetic sieving (Middleton, 1970; Savage and Lun, 1988; Pouliquen and Vallance, 1999). Large particles are not predicted to be transported to the flow front as efficiently for DLE craters because the vertical velocity gradient is subdued. The bottom panels show the final rampart morphologies generated, which are different due to the presence of a surface ice later for DLE craters, which subdues the vertical velocity gradient compared with SLE and MLE craters. This inhibits kinetic sieving and flow-front deceleration for the outer ejecta facies of DLE craters, leading to subdued ramparts with fewer large particles compared with SLE crater ejecta and the outer ejecta facies of MLE craters.

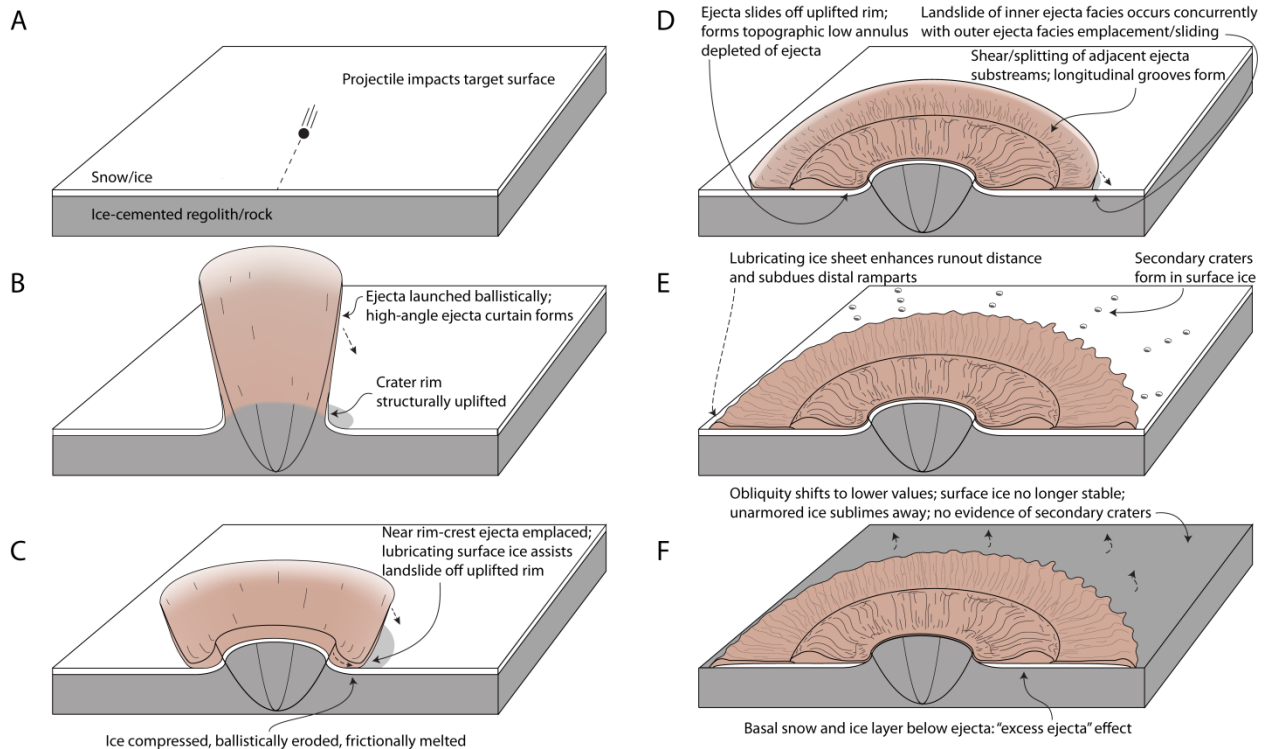


Figure 19. 3D schematic of the glacial substrate model for double-layered ejecta crater formation (Weiss and Head, 2013). (A) A projectile impacts the target surface. (B) Excavation flow generates an expanding ejecta curtain. (C) Near rim-crest ejecta material is ballistically emplaced and begins to slide off the ice-lubricated crater rim. (D) The landslide of the near rim-crest ejecta forms the inner ejecta facies during the emplacement and sliding of the outer ejecta facies. Crater rim collapse (not shown) occurs during this phase for complex craters. (E) The crater is depicted following the impact event. (F) The crater as seen in the present day, after the surrounding surface ice has sublimated away.

Supporting material

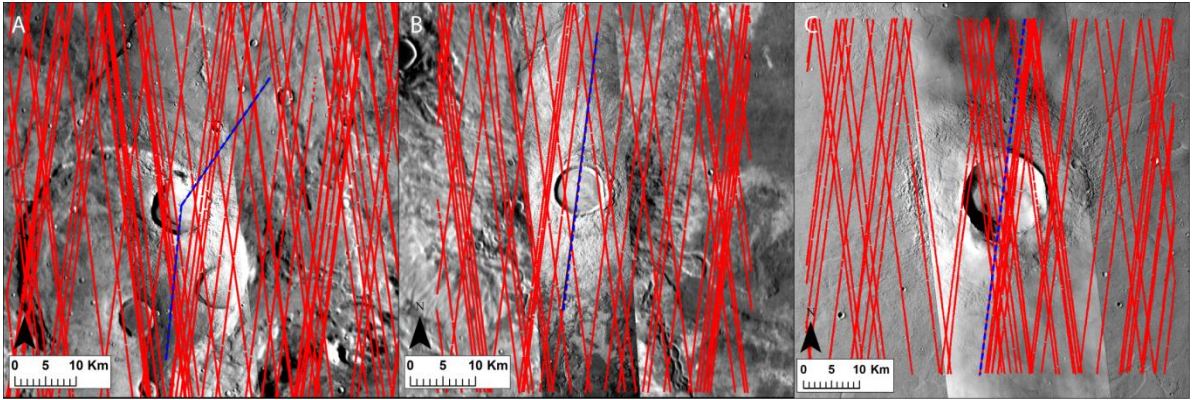


Figure S1. MOLA shots (red dots) used to generate profiles in Fig. 6, 7 and 8. (A) MOLA tracks and profile (dashed blue line) used to generate altimetric profile in Fig. 6C. (B) MOLA tracks and profile (dashed blue line) used to generate altimetric profile in Fig. 7C. (C) MOLA tracks and profile (dashed blue line) used to generate altimetric profile in Fig. 8C. Profiles were taken in areas of the highest MOLA shot density in the regions of interest.

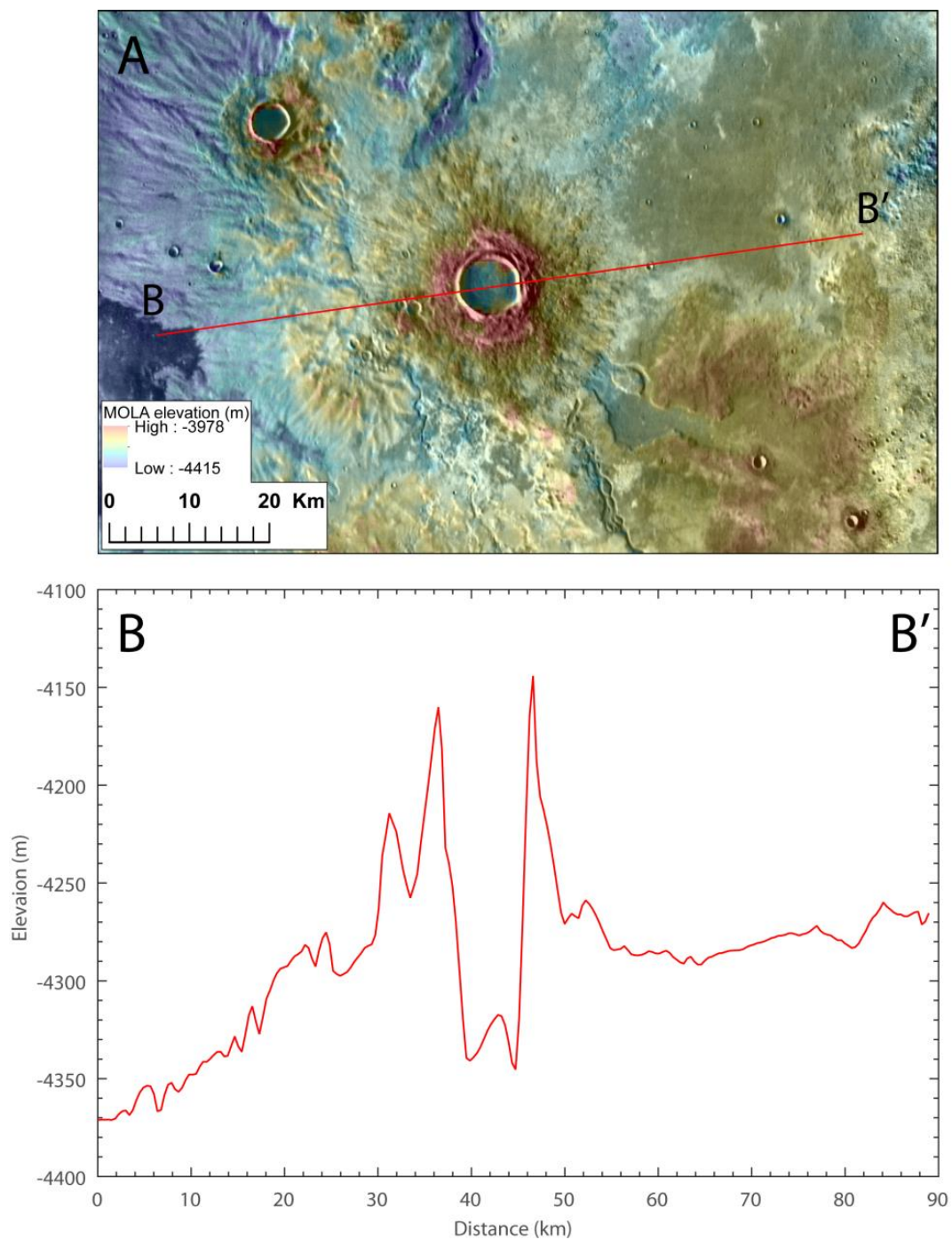


Figure S2.MOLA topographic data overlain on THEMIS global daytime imagery showing the east-west regional slope in Fig. 7.

Flow feature	μ	$h_{\min}$ (m)	h_{avg} (m)	$h_{\max}$ (m)	$U_{\min}$ (m s ⁻¹)	U_{avg} (m s ⁻¹)	$U_{\max}$ (m s ⁻¹)	Note	Reference
Sherman, 1964	0.05	0.5	2	5	18	26	65		Shreve (1966), McSavaney (1975), Sosio et al. (2012a)
Punta Thurweiser, 2004	0.05	4	4.5	5	40	55	65		Sosio et al. (2008)
Black rapids West, 2002	0.16	2	2.5	3		45			Jibson et al. (2006)
Black Rapids East, 2002	0.16	2	2.5	3		45			Jibson et al. (2006)
Mt Iliamna, 2003	0.12	1	1.5	2		48			Schneider et al. (2010)
Mt Cook, 1991	0.13	1	1.25	1.5		55			Schneider et al. (2010)
Kolka Glacier, 2002	0.05	1	8	15	20	50	90		Haeberli et al. (2004), Huggel et al. (2005), Zaporozhchenko (2006), McDougall (2006)
Mt. Steller, 2005	0.09	10	12.5	15		100			Huggel et al. (2008)
Mt. Munday, 1997	0.07		1.5			11	64		Delaney and Evans (2014), Sosio et al. (2012a)
McGinnis Peak, 2002	0.18*		2		40	47	54	min	Jibson et al. (2006)
Hast, Mt. Cook, 2013	0.18*	0.5	2.5	7	42	43.5	45	avg	Hancox et al. (2013)
Vampire, 2008	0.39*	0.5	1	2	28	33	38	avg	Cox et al. (2008)
Shivulich, 1965	0.13	10	12.5	15			80	max	Belousov (1995),

									Sosio et al. (2012b)
Llullaillaco	0.07		10	95	40	45	50	max	Richards and Villeneuve (2001), Sosio et al. (2012b)
Mt. Steele, 2007	0.05	4	22		35	78	140		Lipovsky et al. (2008)
Little Tahoma Peak, 1963	0.21		3		29	55.5	82		Sheridan et al. (2005)
Socompa Proximal	0.04	25	50	70	20	100	100		Kelfoun and Druitt (2005), Kelfoun et al., (2008), Davies et al. (2010)
Mt. Meager, 2010	0.02	5	15	50	9.5	36	91.2		Guthrie et al. (2012)
Tsar Mountain, 2000	0.11	1	2	10	22	33.5	45	avg	Jiskoot (2011)
Mt. Fletcher, 1992	0.19	1	10	60		80	120		McSavaney (2002), Allen et al. (2009)
Blackhawk	0.23*	10	15	30	35			min	Shreve (1968)
Acheron	0.51	16	30	34		25	46		Smith et al. (2006)
Pink Mountain, 2002	0.42*	1	2	5	5	15	30		Geertsema et al. (2006)
Nagano, 2011	0.26*		5			10	14	max	Yamasaki et al. (2014)
Niumiangou Creek, 2008	0.25*	5	7	20	21	28.5	36	min	Zhang et al. (2011)
Wulong, 2009	0.26*		30	60		42.5			Xu et al. (2010)
Frank, 1903	0.14		18		10	40	70		Poisel et al. (2008)
Six des Eaux Froides, 1945	0.13	28	60	66			87	max	Hungr and McDougall (2009)
Socompa distal	0.04	25	60	90	10	20	40		Kelfoun and Druitt

reflected									(2005), Kelfoun et al., (2008), Davies et al. (2010)
Little Salmon Lake, 2008	0.58	0.3	2	10	35			min	Brideau et al. (2009)
Ganges Chasma, Mars	0.07*	40	60	125	38	40	60		Quantin (2004)
Ius east, Mars	0.05*	320	500	790	69.4	320	500		Lucchitta (1987)
Melas, Mars	0.08	170	195	220	49.8	86.73	125		Lucchitta (1987), Mazzanti et al. (2016)
Ius Center, Mars	0.12*	1600	1900	2050	83.3	1600	1900		Lucchitta (1987)
Ophir Labus	0.08	450	520	750	76.7	79.8	83.4		Mazzanti et al. (2016)
Granular flow experiment	0.87	0.0021	0.0025	0.0026	1.15	1.3	1.4		Forterre and Pouliquen (2001)

Table S1. Landslide, avalanche, and granular flow features from Fig. 15. This table is an expansion of Table 1. Note column indicates whether the velocity estimates are minimum, average, or maximum bounds, otherwise, the velocity range for each flow feature corresponds to the minimum, average, and maximum velocities. * denotes where friction coefficients (μ) were measured by this study as $\mu = \tan\theta + \frac{H_0}{\Delta L}$ (Lucas et al., 2014), where θ is the slope the deposit is resting on, H_0 is the thickness of the released mass, and ΔL is the distance traveled.

Chapter 2:
Crater degradation in the Noachian highlands of Mars: Assessing the hypothesis of regional snow and ice deposits on a cold and icy early Mars

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Abstract

The presence of valley networks and the highly degraded state of Noachian highland craters has led to the interpretation that Mars was once warmer and wetter. Recent climate models have suggested, however, that the extremely cold climate in the Noachian would be unlikely to support liquid water precipitation. The presence of a thicker atmosphere thermally coupled to the surface is predicted instead to concentrate surface snow and ice deposits in the higher-altitude southern highlands, producing a Late Noachian Icy Highlands (LNIH) characterized by hundreds of meters of relatively continuous ice cover. In this study we test this hypothesis by reevaluating the degradation state of Noachian highland craters to assess whether their degradation state might be attained in such a cold and icy climate. We review the characteristics of Amazonian-aged impact craters hypothesized to form in surface snow and ice layers (excess ejecta, EE; double-layered ejecta, DLE; and pedestal, Pd, craters) to provide the potential initial conditions of craters forming in Late Noachian surface snow and ice layers. We then examine modification processes active in the Amazonian that may have played a role in crater degradation in the late Noachian. In addition, we examine the potential morphometric effects of impacting into a thick surface ice deposit, and the potential erosive effects of backwasting, top-down melting, basal ice melting, and atmospheric warming pulses on the morphology of Noachian highland craters. We find that several aspects of the highly degraded state of Noachian craters could be accounted for in the context of a cold and icy climate, and we outline further tests of the hypothesis.

1. Introduction

The presence of a faint young sun (Newman and Rood, 1977, Gough, 1981) has led to

the supposition that early Mars was cold, with mean annual temperatures well below freezing (e.g., Kasting, 1991; Haberle et al., 1993; Carr and Head, 2003; Gaidos and Marion, 2003; von Paris et al., 2014). The presence of valley networks (Carr and Clow, 1981; Hynek et al., 2010), the degraded state of Noachian highland craters (Mangold et al., 2012), and the relationship of increasing surface age with elevation in Noachian terrains (Craddock and Maxwell, 1993; Irwin et al., 2013), however, has led previous investigators to suggest that the martian climate in the Noachian ($>\sim 3.6$ Ga; ; Hartmann, 2005; Werner and Tanaka, 2011) was warm and wet, and that liquid water precipitation (rainfall) and surface runoff is the most likely cause of fluvial activity and crater degradation (Craddock and Maxwell, 1993; Craddock et al., 1997; Hynek and Phillips, 2001; Craddock and Howard, 2002; Irwin et al., 2005a, b; Howard et al., 2005; Hoke and Hynek, 2009; Hynek et al., 2010; Hoke et al., 2011), although snow precipitation and melting has not been ruled out (Howard et al., 2005).

Recent climate models, however, have shown that climatic conditions in the Late Noachian (~ 3.6 to 3.8 Ga; Hartmann, 2005; Werner and Tanaka, 2011) may not have been able to support liquid water precipitation (Wordsworth et al., 2013), and that regional snow and ice deposits, much like those inferred to be present at various latitudes in the Amazonian (Head et al., 2003; 2005; 2006a; b; 2010), could have been pervasive in the Late Noachian southern highlands (Head and Marchant, 2014; Fastook and Head, 2015). The models show that even a slight increase in atmospheric pressure enables the martian atmosphere to thermally couple to the surface (Forget et al., 2013; Wordsworth et al., 2013; Scanlon et al., 2013), a scenario in which the Noachian highlands acts as a cold trap and preferentially accumulates atmospheric snow and ice deposits (Fastook et al.,

2012; Head and Marchant, 2014). This set of cold and icy conditions is known as the Late Noachian Icy Highlands (LNIH) scenario (Wordsworth et al., 2013; Head, 2013).

Concentrating the current Amazonian ice supply from the polar caps above an equilibrium line altitude (ELA) of 1 km, is volumetrically equivalent to a thickness of ~280 m of ice concentrated in the southern highlands (Head and Marchant, 2014; Fastook and Head, 2015), although this estimate may be significantly underestimated based on the unknown quantity of water lost from Mars since the late Noachian (Jakosky et al., 1994; Kass and Yung, 1999; Hodges, 2002; Greenwood et al., 2008; Andrews-Hanna and Lewis, 2011; Carr and Head, 2014).

Wordsworth et al. (2015) further conducted a variety of 3D global climate models of both warm/wet and cold/icy scenarios. In order to produce warm/wet conditions, they artificially increased the solar flux. In this case, Wordsworth et al. (2015) found poor correlation between the zones predicted to exhibit liquid water precipitation and those which host the valley networks (e.g., Hynek et al., 2010). In contrast, the models which assumed nominal cold Late Noachian conditions (mean annual temperature of 225.5 K) showed that the regions with high valley network densities correlated much better with zones predicted to accumulate snow (Wordsworth et al., 2015). Thus, the LNIH model appears robust enough to be further tested.

In this paper, we examine modes of crater modification expected in the LNIH scenario for comparison with the observed degraded morphology of Noachian highland craters. Our aim is not to prove or disprove conclusively whether the LNIH early Mars climate model is correct or not; rather, we provide observational tests of the climate model of Wordsworth et al. (2013) using one geologic line of evidence, the morphology

and degradation state of Noachian highland craters, noting that future work is required to test whether other features of the Late Noachian are consistent with such a model. We first outline and document the set of morphologic features and morphometric associations that characterize impact craters formed in the Noachian highlands (Section 2). Next, we look to the nature of impact craters formed in snow and ice in the later, Amazonian period of the history of Mars as a guide to candidate formation and degradation processes that might have operated in a LNIH climate scenario (Section 3). We then use these characteristics and guidelines to assess candidate processes that might have operated during both crater formation and later modification of Noachian craters (Sections 4, 5, and 6). Finally, we examine these candidate characteristics and processes to assess whether they can successfully explain the characteristics of Noachian highland craters (Section 7). We find that several aspects of Late Noachian crater degradation states, commonly thought to be explained by “warm and wet” climatic conditions, are also consistent with formation and modification in the context of Late Noachian icy highland climate conditions. We conclude with suggestions for further tests and investigations.

2. Characteristics and degradation state of Noachian craters

Martian Noachian highland craters differ markedly from fresh martian craters (Table 1; Fig. 1 through 4) in that they are highly degraded (Craddock and Maxwell, 1990; Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Mangold et al., 2012). Noachian highland craters exhibit subdued crater rims when compared with fresh craters (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002). Craddock et al. (1997) compared

the morphometry of fresh and degraded crater rim heights and found that the least squares fit for rim heights of degraded craters was systematically lower than fresh craters by factors of ~ 3 to ~ 10 , and that many degraded craters lacked any observable rim topography. These results highlight the importance of either significant and pervasive degradation processes, or anomalous formation processes, or a combination of both.

Noachian highland craters possess flat and shallow crater floors (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002), in contrast to deeper and rough-floored (often with central peaks) craters characteristic of young impacts. Craddock et al. (1997) compared the morphometry of fresh and degraded crater depths and found that the least squares fit for crater depths of degraded craters was systematically shallower than fresh craters by factors of ~ 1.6 to ~ 5 , an observation supported by a more recent study of Noachian craters in North Hellas Planitia and Southern Margaritifer Terra (Mangold et al., 2012). While Craddock et al. (1997) showed that rim erosion can contribute to the apparent shallowing of these craters, the flat nature of the floor and anomalously shallow depths point to major contribution by additional processes infilling the crater (Warner et al., 2010). Mangold et al. (2012) measured and compared the slopes of Noachian highland craters to more well-preserved, early Amazonian craters in North Hellas Planitia and Southern Margaritifer Terra and found the curvature of the interior crater wall to be substantially lower for the degraded Noachian craters, confirming that Noachian highland craters possess unusually flat floors. Furthermore, craters present on Early Noachian-aged terrains (>3.96 Ga; Hartmann, 2005; Werner and Tanaka, 2011) tend to be deeper than those found on younger Noachian terrains (Irwin et al., 2013). These results emphasize the major difference in

morphology and morphometry between Noachian and younger craters.

Many Noachian highland craters possess channels superposing crater rims (Masursky et al., 1977; Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002). These channels may be limited to small gullies on the crater wall interior or branching valleys within 2 crater radii (R) from the rim (Mangold et al., 2012). The presence of gullies suggests a role of fluvial activity from rainfall or snowmelt in the modification of Noachian craters.

Noachian highland craters are characterized by a relative absence of discernible ejecta facies (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Mangold et al., 2012), and an absence of secondary craters and crater chains. Channels frequently superpose near-rim areas which were likely to have been previously covered by ejecta deposits (Mangold et al., 2012). The general lack of ejecta and secondaries further emphasizes the unusual characteristics of Noachian craters.

Additionally, diameter-frequency distributions of the Noachian crater population as a whole shows a paucity of degraded craters $< \sim 10\text{-}20$ km in diameter (McGill and Wise, 1972; Jones, 1974; Craddock and Maxwell, 1990), and a complete absence of Noachian-aged craters ≤ 4 km in diameter (Irwin et al., 2013). The paucity of small craters has been previously hypothesized to relate to a deficiency in crater production (Whitaker and Strom, 1976; Barlow, 1990) or erosion/deposition (Strom et al., 1992; Craddock and Maxwell, 1993; Jones, 1974; Chapman, 1974; Chapman and Jones, 1977; Irwin et al., 2013).

In summary, the morphology of Noachian highland impact craters (Table 1; Fig. 1 and 4) suggests that they have been heavily degraded (Craddock and Howard, 2002;

Mangold et al., 2012), although the mode of degradation has been debated (see Craddock and Howard, 2002). These morphologic characteristics have been variously explained by: (1) burial by air fall deposits (Hartmann, 1971; Wilhelms and Baldwin, 1989; Grizzaffi and Schultz, 1989; Grant and Schultz, 1990; Moore, 1990; Barlow, 1995), (2) erosion by groundwater sapping (Gurnis, 1981; Pieri, 1980), (3) fluvial erosion (Barlow, 1995; Forsberg-Taylor et al., 2004) by rainfall and surface runoff (Craddock and Maxwell, 1993; Craddock et al., 1997; Hynek and Phillips, 2001; Craddock and Howard, 2002; Irwin et al., 2005a, b; Howard et al., 2005; Hoke and Hynek, 2009; Hynek et al., 2010; Hoke et al., 2011), (4) impact-induced seismic liquefaction (Clifford, 1997), (5) a complex interweaving of erosion, deposition, and cratering (Malin and Edgett, 2001), or (6) erosion from melted snowpack (Kite et al., 2011).

In contrast to this wide range of degradation processes, generally unaddressed has been the role that formation in unusual substrates might have played in forming the initial morphology and morphometry of Noachian craters. We now examine Amazonian craters that have been interpreted to have formed in snow and ice to assess their application to the degradation state of Late Noachian impact craters.

3. Amazonian Crater Characteristics

In the Amazonian, the presence of regional snow and ice deposits concentrated in the middle to high latitudes has been shown to affect the impact cratering process (e.g., Wrobel et al., 2006; Barlow, 2006; Meresse et al., 2006; Senft and Stewart, 2008; Black and Stewart, 2008; Kadish and Head, 2011; Weiss and Head, 2013; 2014; 2015); impact into decameters-thick snow and ice deposits (Fig. 5) has been proposed to produce

pedestal (Pd; Wrobel et al., 2006; Kadish et al., 2008; 2009; 2010), excess-ejecta (EE; Black and Stewart, 2008; Kadish and Head, 2011), and perched (Pr) craters (Meresse et al., 2006). More recently, the low-aspect-ratio layered ejecta (LARLE; Boyce et al., 2015; Barlow et al., 2014), and double-layered ejecta (DLE) crater (Weiss and Head, 2013; 2013a; 2015) morphologies have also been attributed to impact into icy surface layers (Fig. 2). Could the characteristics of these craters provide predictions as to the initial state of impact craters formed in a Late Noachian Icy Highlands?

Pedestal craters (Fig. 3) are typically $< \sim 2$ km in diameter, are present in the middle to high latitudes, and are characterized by a crater perched near the center of an elevated plateau surrounded by a semi-circular outward-facing scarp that extends on average $\sim 3R$ from the rim crest (Kadish et al., 2009; 2010). Pedestal craters are proposed to form when the impactor is of insufficient size to excavate substantial ejecta from beneath the surface snow and ice layer (Wrobel et al., 2006; Kadish et al., 2009; Weiss and Head, 2014), or as an erosional remnant of a thin base surge deposit around LARLE craters which penetrate into an ice-rich fine grained mantling deposit (Barlow et al., 2014; Boyce et al., 2015). The icy material proximal to the crater may be armored from sublimation by a blast wave (Wrobel et al., 2006) or the deposition of fine-grained base surge material (Boyce et al., 2015). The unarmored icy material surrounding the pedestal can then sublime away during a later, different climate regime, leaving the pedestal perched several tens of meters above the surrounding terrain (Wrobel et al., 2006; Barlow, 2006; Kadish et al., 2008, 2009). The post-formation climate regime is still cold and icy and not sufficiently warm to cause melting of the ice beneath the pedestal (Kadish et al., 2009; 2010).

Excess-ejecta (EE) craters are craters that possess ejecta facies with a volume in excess of that of the crater cavity by factors of 2.5 to 28, and are hypothesized to form when an impact excavates through a decameters-thick surface snow and ice deposit (Black and Stewart, 2008; Kadish and Head, 2011). When the impact occurs, ejecta deposited on top of the surface icy deposits armor the underlying ice from sublimation during a later climate regime (thus producing the “excess ejecta” effect). Perched craters are hypothesized to have an identical origin to EE craters (Meresse et al., 2006), but subsequent infill of the crater cavity has left the crater cavity perched above the surrounding terrain.

A relatively common morphological class of craters, known as double-layered-ejecta craters (Fig. 2), has also been proposed to share the theme of formation in surface icy layers (Weiss and Head, 2013; 2014; 2015). Double-layered-ejecta craters are typically $< \sim 25$ km in diameter (7.8 km average), are concentrated in the middle to high latitudes (Barlow and Perez, 2003; Boyce and Mouginis-Mark, 2006; Weiss and Head, 2014; Li et al., 2015), and are characterized by two ejecta facies: the inner ejecta facies typically extends $1.5R$ from the rim crest, is semi-circular, and displays longitudinal grooves emanating from the rim crest (Barlow, 2005); the thinner outer ejecta facies is characterized by smaller, sinuous grooves and an anomalously high runout distance ($\sim 3R$ from the rim crest on average; Boyce and Mouginis-Mark, 2006; Weiss and Head, 2014). DLE craters are proposed to form when the impactor is of sufficient size to excavate through the surface snow and ice layer and into the deeper, ice-cemented rock/regolith. In this scenario, the inner ejecta facies may be formed through a landslide of the near-rim ejecta, facilitated by lubricating snow and ice below the ejecta (Weiss and Head, 2013).

The excess-ejecta and perched craters thus far documented have all been classified as DLE craters (Kadish et al., 2011; Weiss and Head, 2013). DLE craters have alternatively been suggested to form through impact into a volatile-rich target (e.g., Mouginis-Mark, 1981; Barlow and Bradley, 1990; Barlow; 1994, 2005; Barlow and Perez, 2003; Senft and Stewart, 2008; Jones and Osinski, 2015) followed by atmospheric vortex interactions (Schultz and Gault, 1979; Schultz, 1992), the collapse and base surge of an explosion column (Mouginis-Mark, 1981; Boyce and Mouginis-Mark, 2006; Harrison et al., 2013), impact melt overtopping the rim (Osinski, 2006; Osinski et al., 2011); or a landslide of near-rim ejecta on a rocky target (Wulf and Kenkmann, 2015).

Could the nature of these crater types, formed in snow and ice and ice-cemented regolith, provide insights into crater formation and morphology in a hypothesized Late Noachian Icy Highlands environment? In the following sections, we use these guidelines to examine the potential effects of regional snow and ice deposits (on the order of hundreds of meters thick as predicted by the LNIH model), in order to assess whether they are applicable to the formation and subsequent modification of Noachian highland craters.

4. Amazonian crater formation into a Late Noachian Icy Highlands

Impact into regional snow and ice deposits in the southern highlands during the Noachian could form craters analogous to several morphologies of Amazonian craters (DLE and pedestal craters, Fig. 2 and 3). Noachian highland craters which formed on regional snow and ice deposits will differ from their Amazonian counterparts given that the Amazonian regional average snow/ice thickness is ~50 m (Kadish et al., 2010;

Fastook and Head, 2013), as opposed to a lower-bound Noachian average snow/ice thickness of ~280 m in the LNIH model (Fastook and Head, 2015). Because some of the depth of these “fresh” Noachian craters would be accommodated by the surface snow and ice layer, some of the rim structural uplift would be present in the hectometers-thick snow and ice substrate. Subsequent sublimation and/or melting of the near-rim ice in a later, different climate regime would lower the observed rim height of these craters, potentially by hundreds of meters. Larger craters that penetrate through the ice will produce a smaller crater cavity in the underlying rocky substrate relative to their diameter. This is shown in Fig. 6, where a 16 km diameter pedestal crater located in the Dorsa Argentea Formation (Fig. 6A) is compared with a 12 km diameter single-layered ejecta crater (Fig. 6B). If the pedestal (red line, Fig. 6C) were removed, both craters would appear to be 12 km in diameter, but the cavity remaining in the bedrock from the pedestal crater would be substantially shallower than for typical craters (blue line, Fig. 6C). Consequently, impact craters that are predicted to penetrate through the surface ice in the LNIH model will produce anomalously shallow crater cavities in the underlying bedrock relative to their given diameter, and may thus be infilled and shallowed more rapidly (Fig. 8A) by settling and loss of the icy substrate, and related erosion of the silicate debris.

Are DLE craters useful analogs to craters forming in a LNIH ice sheet? Amazonian-aged DLE craters are much smaller (typically < 25 km in diameter) than most craters forming in the Late Noachian. It has been proposed that larger impact craters forming in surface ice avoid classification as DLE craters because they may not typically generate an observable inner ejecta facies (Weiss and Head, 2014). Consequently, the relatively

larger Late Noachian craters forming in an LNIH ice sheet are not predicted to exhibit an inner ejecta facies (right panel in Fig. 8). The DLE crater analogy for LNIH craters is useful to examine the initial conditions following the impact because they are both interpreted to form in surface ice. The majority of the degraded highland craters in the LNIH scenario, however, may instead have appeared morphologically similar to the multiple-layered ejecta (MLE) crater class, because the MLE morphology is comparable to the outer facies of double-layered ejecta craters (arrows in Fig. 7; Weiss and Head, 2014).

5. Modification processes

What modification processes might further degrade craters that impacted into hectometers-thick surface snow and ice deposits in the Late Noachian Icy Highlands model? In this section, we examine several erosional processes that would be expected to modify Noachian highland craters following impact into surface snow and ice deposits.

5.1 Backwasting

Concentric crater fill (CCF) deposits, which display a characteristic concentric ridge texture on their surface, infill Amazonian-aged craters in the middle to high latitudes of Mars (Levy et al., 2010). CCF craters are interpreted to form through regional snow and ice precipitation, followed by cold-based glacial flow into craters due to steep inner rim-crest slopes (Fastook and Head, 2013). Once the crater rim-crest is exposed, backwasting of material from the crater rim crest (gravitationally-induced mass wasting of the inner rim crest) armors the surface of the glacial deposits as they flow into the crater interior

(Fastook and Head, 2013). Subsequently, climate change causes sublimation and removal of the surrounding unarmored icy deposits, leaving the remnant debris-covered crater fill (the CCF; Fastook and Head, 2013). Consequently, the preservation of CCF deposits dating from a glacial epoch provides an observational foundation that material from crater rims is constantly being backwasted and infilling the craters in the absence of fluvial activity. This same process could be active during a cold and icy Late Noachian and would contribute to shallowing of crater floors and subduing of the rims of Noachian highland craters (Fig. 8B). Craddock et al. (1997) used a volume conservation model for fresh craters and found that backwasting could account for a depth decrease of ~23 – 47% for moderately degraded craters between 10-50 km in diameter (5-10% enlargement by backwasting), increasing up to ~54 – 65% for heavily degraded craters between 10-50 km in diameter (5-10% enlargement by backwasting). This indicates that backwasting may play a moderate role in crater infill, particularly when assisted by fluvial runoff (Craddock and Howard, 2002).

5.2 Top-down melting

Snow is expected to accumulate on craters following impact, both from locally (Kite et al., 2011) and regionally derived (Wordsworth et al., 2013) atmospheric precipitation. Any solar-insolation induced top-down melting (e.g. Wordsworth et al., 2013; Cassanelli and Head, 2014; Fastook and Head, 2015), such as that proposed for the formation of gullies in the cold and dry Amazonian (Christensen, 2003; Milliken, 2003; Bleamaster and Crown, 2005; Bridges and Lackner, 2006; Dickson et al., 2007; Head et al., 2008; Dickson and Head, 2009; Schon and Head, 2011), or albedo-enhanced melting from a

thin dust-cover over ice (Clow, 1987; Williams et al., 2009) could generate runoff from the melting of surface snow and ice deposits, thereby further eroding the rim and ejecta facies, infilling the crater, and generating interior wall channels on Noachian highlands craters (Fig. 8B). Future climate modeling is required to characterize the presence, location and magnitude of melting environments produced by solar insolation in the LNIH scenario, but top-down melting in a LNIH model provides an additional candidate process for crater degradation.

6. Melting processes and further tests of the hypothesis

The previous sections outlined how impact into the hectometers-thick surface snow and ice envisioned in the LNIH model is predicted to naturally produce smaller crater cavities and subdued rims prior to icy rim removal, and how subsequent modification processes (backwasting, solar-insolation induced top-down melting) could further erode rims and ejecta, infill craters, and generate interior wall channels. If dry backwasting and gully formation were the major contributors to Noachian highland crater degradation, however, some Amazonian craters would be expected to share the numerous degraded characteristics of Noachian highland craters, which is not observed. Consequently, other modification processes are necessary to explain the degraded state of Noachian highland craters.

Craddock et al., 1997 modeled rock weathering, fluvial backwasting, and linear diffusion creep in the degradation of martian impact craters using the MARSSIM model (Howard, 1994) and found creep to be insufficient to explain the morphologies; the scale-inefficient nature of creep (low transport distances) prevented larger craters from

exhibiting large degrees of degradation. Fluvial erosion, however, was found to be scale-efficient and could explain larger degraded craters and steep crater walls (Craddock et al., 1997; Craddock and Howard, 2002; Howard et al., 2005). Forsberg-Taylor et al. (2004) used the same model (Howard, 1994) and tested the relative effects of aeolian infill and fluvial backwasting. They (Forsberg-Taylor et al., 2004) found that while aeolian processes could contribute to infilling the crater, fluvially enhanced backwasting was required to erode the rim and produce the steep inner crater wall slopes. Could fluvial activity have acted to erode crater rims and infill craters in the cold and icy late Noachian scenario? In the following sections, we examine potential mechanisms that might generate fluvial activity around impact craters that formed in hectometers-thick surface snow and ice deposits in the LNIH scenario. We then assess whether these mechanisms are sufficient to explain the observations.

6.1 Melting from hot ejecta

When an impact occurs, a hemispherical shock-wave propagates outward through the target material, subjecting the target material to elevated pressures and temperatures (e.g., Melosh, 1985; Keil et al., 1997, pg. 353-354). Although craters less than ~30 km in diameter do not produce high enough shock pressures to substantially heat their continuous ejecta facies (e.g., Stewart et al., 2004; Black and Stewart, 2008; Senft and Stewart, 2008; Ivanov and Pierazzo, 2011; Stöffler et al., 2013), larger craters may produce ejecta hot enough to melt snow/ice. In this case, snow landing on the hot ejecta of larger craters would melt, and could erode material through debris flows (Kite et al., 2011). The hot ejecta of these larger craters would also melt some of the underlying

surface ice deposits, which could produce fluvial channels and erosion of the ejecta similar to that observed for some Amazonian (<3 Ga; Hartmann, 2005; Werner and Tanaka, 2011) and Hesperian (~3 to 3.6 Ga; Hartmann, 2005; Werner and Tanaka, 2011) impact craters (see Morgan and Head, 2009; Mangold, 2012). Melting of pore ice (Stewart et al., 2004) within the ejected target material is expected to be enhanced during in the LNIH scenario (due to the warmer temperatures relative to Amazonian conditions; e.g., Stewart et al., 2004); drainage of this melted pore ice might also contribute to erosion of the ejecta.

6.2 Basal melting

In the LNIH scenario, another potential source of water available for fluvial erosion is the snow and ice hypothesized to be present underlying the ejecta. Could basal melting of this snow and ice also contribute to crater modification? Although the geothermal gradient in the Noachian remains poorly constrained (McGovern et al., 2004; Solomon et al., 2005), previous investigators (Carr and Head, 2003; Zent, 1999; Fastook et al., 2012; Cassanelli and Head, 2015) have shown that it may be possible to generate basal melting only in instances of kilometers thick snow and ice deposits. The Noachian highlands ice budget may be sufficient to generate ~280 m+ average thickness of snow and ice (Fastook and Head, 2015), and thus the average LNIH regional snow/ice deposits alone would be of insufficient thickness to generate basal melting (Fastook and Head, 2015; Cassanelli and Head, 2015).

Ejecta deposition on top of the regional snow and ice deposits, however, has not been previously considered, and could potentially be important. In the LNIH scenario, tens to

hundreds of meters of silicate regolith and bedrock ejecta would be deposited on the snow and ice deposits as a result of an impact cratering event; these icy deposits would remain present during crater modification. The low thermal conductivity of the ejecta (Warren, 2011) overlying the snow and ice deposits, in concert with an elevated Late Noachian geothermal heat flux, could potentially raise basal temperatures sufficiently to generate melting of the underlying snow and ice. A cycle of runoff, evaporation, snow deposition and cratering might allow this process to continue across Late Noachian terrains for extended periods of time.

We undertake here some preliminary calculations to examine the potential for basal melting in the near-rim zone of newly produced craters. In order to address this, we apply the steady state one-dimensional heat conduction equation to examine the basal temperatures of ice superposed by tens to hundreds of meters of ejecta. The steady state basal temperature (T) of ice can then be calculated:

$$T_{(z)} = T_{(z-1)} + \Delta T \quad (1)$$

$$\Delta T = \frac{Q\Delta z}{\kappa_z} \quad (2)$$

where $T_{(z=0)}$ is surface temperature. The basal temperature of ice is then a function of the geothermal heat flux, Q , thermal conductivity of ejecta and ice, $\kappa_{(z)}$, ejecta thickness, and ice thickness (z). To examine the potential for basal melting occurring in the thicker near-rim ejecta, we use fresh ejecta thickness from the maximum rim height functions of Stewart and Valiant (2006):

$$H_{Min} = 0.138 D_C^{0.23} \quad (3)$$

$$H_{Mid} = 0.091 D^{0.47} \quad (4)$$

$$H_{Max} = 0.133 D_C^{0.51} \quad (5)$$

where H is the least-squares fit to a number of fresh martian crater measurements: H_{Min} is from Solis Planum (N=33), H_{Mid} is from Lunae Planum (N=48), and H_{Max} is from Isidis Planitia (N=24). These different rim height functions are used to illustrate the range in fresh crater ejecta thicknesses observed on Mars between ~2 to 50 km in diameter. These functions agree with the results of Craddock et al. (1997) who performed similar measurements (N=264), and are used to capture the wide topographic variability exhibited by crater rim heights. D_C is crater diameter (for crater diameters > 8 km). By subtracting the structural uplift at the rim (~50% the rim height; Stewart and Valiant, 2006) from H , we estimate the maximum ejecta thickness at the rim. While Sharpton (2014) found that the structural uplift composed ~80% of the rims of lunar craters, this does not appear to be the case for martian craters (Mouginis-Mark and Boyce, 2012; Mouginis-Mark, 2015; Sturm et al., 2014). Indeed, the volatile-rich nature of the martian crust in a cold and icy early Mars scenario would lead to higher excavation angles (e.g., Greeley et al., 1980; O’Keefe et al., 2001; Stewart et al., 2001; Stewart et al., 2004; Senft and Stewart, 2008), and thus thicker near-rim ejecta deposits (relative to the Moon) are expected.

We use κ_E of 0.3 W/m·K (typical of lunar breccia; Horai and Winkler, 1980; Warren and Rasmussen, 1987; Warren, 2011) and a κ_i of 0.7 W/m·K (firn), 1.5 W/m·K (firn/ice), and 2.25 W/m·K (ice; Carr and Head, 2003; Cassanelli and Head, 2015), for Late

Noachian Q values of 45, 55, and 65 mW/m² (McGovern et al., 2004; Clifford et al., 2010; Fastook et al., 2012). We assume that the excavated rocky ejecta below the surface ice are 10% pore ice by volume (e.g., Hannah and Phillips, 2005). We use a surface temperature of 225 K, which is representative of the Late Noachian southern highlands mean annual surface temperatures for atmospheres of 1 bar (Wordsworth et al., 2013, 2015). By examining the range of crater diameters that have sufficient ejecta thicknesses to produce basal temperatures in excess of 273 K (the pure ice-melting isotherm expected from atmospherically emplaced snowfall), we can establish the minimum crater diameter expected to produce basal melting over the range of parameters.

We find (Fig. 9) that the minimum crater diameter for the onset of basal melting ranges from 20 km to many hundreds of kilometers depending on the rim height function, heat flux, ice thickness used, and compaction state of the icy surface (e.g. firn versus ice; Cassanelli and Head, 2015). Figure 9A shows that for the minimum rim thickness function (eq. 3), basal melting can only occur for thick firn (250+ m) and moderately high Q values (55-65 mW/m²) in the crater diameter range of interest. The middle rim height function (Fig. 9B; eq. 4) shows that the onset of basal melting can occur for smaller crater diameters (40-180 km) depending on the assumed heat flux and whether the ejecta is overlying firn, firn/ice, or ice. In this scenario, the low thermal conductivity of firn produces a relationship between increasing ice thickness and decreasing basal melting onset diameter. The max rim height function (Fig. 9C; eq. 5) generally eliminates this trend due to the presence of an enhanced thickness of lower thermal conductivity ejecta. Figure 9C shows that the onset of basal melting ranges in this scenario from ~20-80 km in diameter, although it clusters around ~40 km in diameter over the range of ice

thicknesses, thermal conductivities, and geothermal heat fluxes. Deposition of thick deposits of low thermal conductivity ejecta on regional snow and ice deposits may thus help to produce melting of regional snow and ice deposits in the Noachian. Such ice melting could serve to assist in the erosion of the near-rim ejecta and rims of Noachian highlands craters. Such melting would also have the potential to generate channels through fluvial erosion, and to assist in infilling the crater.

Craddock and Maxwell (1993) and Craddock et al. (1997) found that Noachian craters possess a variety of degradation states at any given size and concluded that degradation was not a short-lived catastrophic event (in which case highland craters might all have the same state of degradation), but that it decreased in intensity through time and lasted until the late Hesperian. Can these observations be explained in the context of the LNIH model? In the LNIH scenario the extent of erosion any crater can experience by basal melting is dependent on (1) surface temperature (which is latitude- and elevation-dependent); and (2) the thickness of ice and overlying ejecta at the time of impact. Because rim heights (and thus near-rim ejecta) naturally vary on the order of ~10-100+ m for craters within the same degradation state (e.g. eq. 3-5; Stewart and Valiant, 2006), basal melting is not expected to be equivalent for all craters (i.e., not all craters above ~40 km in diameter would be expected to produce basal melting for any given ice thickness and heat flux; Fig. 9). Furthermore, because snow continually deposited on the crater and ejecta facies subsequent to the impact cannot be melted until another impact event occurs (in the absence of top-down melting), the ice thickness at time of impact can limit the supply of melt available for erosion. Greater degrees of crater degradation would thus be favored by lower elevation, lower latitude, larger impacts into thicker ice, and

impacts that deposit more ejecta near the crater rim. Thus, in the LNIH scenario, variations in geographic location, ejecta thickness, and ice thickness may introduce significant variations in the degradation state of Late Noachian craters. Previous investigators (Craddock and Maxwell, 1993; Craddock et al., 1997) have interpreted the variety of observed degradation states of Noachian highland craters to reflect long-lived erosion that decreased in intensity through time. On the other hand, in the LNIH scenario, the initial conditions at the time of impact (e.g., ice thickness, elevation) may also contribute to the variety of observed degradation states, and so long-lived erosion in an LNIH climate scenario could have been accompanied by large-scale erosive events.

6.3 Potential for large-scale erosional events

Jones (1974) analyzed the crater size-frequency distributions for three regions within the Noachian highlands (Fig. 10) and found: (1) at larger crater diameters (>40 km) the frequency distributions of highly, moderately, and slightly degraded craters were parallel; (2) a sharp drop-off in frequency below the maximum frequency occurred for all regions and degradation states at smaller crater diameters; and (3) the maximum frequency of craters shifted to larger diameters with decreasing crater degradation. Jones (1974) modeled the effects of crater production and obliteration on crater size-frequency distributions and concluded that these features (Fig. 10) pointed to an intense short-lived episode of crater “obliteration” followed by a return to approximately the prior erosion rate, and that regional variations in degraded crater densities in the Noachian highlands resulted from differences in obliteration rates. Chapman (1974) similarly modeled the effects of an obliteration episode and found that one short period of obliteration can

match the observed Noachian incremental crater size-frequency distribution of Noachian highland craters. Much like Jones (1974), Chapman (1974) concluded that the Noachian highlands exhibited a relatively constant erosion rate punctuated by one short period of obliteration (coincident with the construction of the Hesperian ridged plains; see Chapman and Jones, 1977). This is consistent with the observation that there exists a bimodal depth/diameter distribution between Noachian and Hesperian-aged mid-latitude craters, which is interpreted to result from a short-lived erosional event followed by a decreased rate of degradation (Boyce and Garbeil, 2007). Indeed, more recent studies of crater degradation analyzed Noachian and Hesperian crater size-frequency-distributions and concluded that Noachian crater degradation occurred during a short period between ~ 3.8 - 3.6 Ga (Warner et al., 2010), and that a sharp transition in degradation mode occurred to at ~ 3.6 - 3.7 Ga (Mangold et al., 2012; Warner et al., 2010). These ages are also in concurrence with the cessation of valley network ages at ~ 3.75 Ga on average (Neukum system) (Fassett and Head, 2008; see Fig. 4A in reference for age uncertainties), but post-date the end of the heavy bombardment period (~ 3.9 Ga; e.g., Chapman et al., 2007; Fassett and Head, 2011). Are any of these observations and trends consistent with the LNIH scenario?

The widespread presence of thick surface snow and ice in the southern highlands (as in the LNIH climate model) offers a source for fluvial erosion during an “obliteration” event (Fastook and Head, 2015). If atmospheric warming pulses occurred near the Noachian-Hesperian boundary, it is plausible that these could have influenced the modification of Noachian craters. The emplacement of the Late Noachian-Early Hesperian ridged plains might have generated planet-wide warming (Phillips et al., 2001;

Johnson et al., 2008; Ehlmann et al., 2011; Halevy and Head, 2014). Volcanism- or impact-induced greenhouse warming (Halevy and Head, 2014; Toon et al., 2010) could potentially create short-lived “thermal pulses” which could significantly affect Noachian highland crater morphology in the presence of regional snow and ice deposits. If an atmospheric thermal pulse occurred, fluvial erosion could erode crater rims, ejecta facies, infill the crater, and generate superposed channels. Ice related to Noachian highlands craters could occur both underlying the ejecta facies (following impact) and overlying the ejecta facies (from emplacement during periods of snow deposition). Smaller craters forming exclusively within the surface ice layers, analogous to Amazonian pedestal craters (Barlow, 2006; Kadish et al., 2009; 2010), are predicted to be entirely melted and eliminated in this scenario because heating may melt and remove the icy “pedestal” and the crater cavity contained within. Thus, some of these observations of crater populations and degradation states could be consistent with a LNIH climate model.

6.4 Paucity of small craters

The size frequency distributions of lunar, mercurian, and martian impact craters each exhibit a deficiency in craters $< \sim 70$ km in diameter (e.g., Barlow, 1988, 1990), which has led to the suggestion that the paucity of craters in this diameter range reflects the size distribution of the impacting population (Whitaker and Strom, 1976; Barlow, 1990). Although the Moon and Mercury exhibit similar downturns at small crater diameters ($< \sim 70$ km), the impact crater population between 20-128 km in diameter on Mercury is deficient in craters when compared to the Moon (Strom, 1977; Fassett et al., 2011). Fassett et al., (2011) noted that because the Moon and Mercury are believed to have the

same impactor population and cratering efficiency (defined as the ratio of the mass of the ejected/displaced material to that of the projectile; e. g., Holsapple and Schmidt, 1979) (Strom et al., 2005; Marchi et al., 2009), the deficiency of smaller mercurian craters relative to the Moon is more likely to be related to volcanic resurfacing on Mercury (e.g., Strom, 1977; Woronow, 1982; Strom et al., 2008).

The observation that younger martian craters (which have not experienced substantial erosion) exhibit a downturn in crater size frequency similar to that observed for older martian craters (which are highly degraded) has also been interpreted to support the suggestion that the downturn in the crater size frequency distribution for Mars is a result of the impacting population (Barlow, 1990). It is important to note, however, that the martian crater size-frequency distribution is deficient in craters $<\sim 40$ km in diameter relative to the Moon. On the basis of a similar impactor population found for the near-Earth asteroids and young martian craters, Strom et al. (2005) argued that the Moon and Mars did have the same impactor population. In concert with the observation that Mars exhibits a paucity of small craters relative to the Moon (Strom et al., 2005; Rodríguez et al., 2005), this suggests that while the downturn in the crater size-frequency distribution for small crater diameters on the Moon, Mars, and Mercury is reflective of the impacting population, both Mercury and Mars (which are deficient in small craters relative to the Moon) experienced enhanced crater obliteration relative to the Moon. Consequently, one interpretation of the paucity of small Late Noachian impact craters ($<\sim 10$ -20 km in diameter) is the role of erosional processes (Strom et al., 1992; Craddock and Maxwell, 1993; Jones, 1974; Chapman, 1974; Chapman and Jones, 1977). Recently, Irwin et al. (2013) found that the mid-Hesperian production crater population from 1 to 4 km in

diameter supports a post-Noachian resurfacing event in the highlands (i.e., surface features in that spatial scale were resurfaced). The removal of any regional highlands ice sheets (such as the Late Noachian-Early Hesperian expanded south polar ice sheet, the Dorsa Argentea Formation; Head and Pratt, 2001; Ghatan and Head, 2002, 2004; Fastook et al., 2012; Scanlon and Head, 2014; Kress and Head, 2015; see Fig. 11 for Hesperian polar unit boundaries) could account for such a resurfacing event. Smaller craters, which may form pedestal craters, are predicted to be eliminated subsequent to the removal of the surface ice (by melting and/or sublimation), and may expected to contribute to the paucity of small craters.

What size distribution of craters might be effected by the removal of hundreds of meters of surface ice? On the basis of the heights of the preserved ice-rich pedestals (Kadish et al., 2008; 2010), Amazonian-aged pedestal craters (which are typically $< \sim 2$ km in diameter) are interpreted to have resulted from impacts into an ice deposit which ranges from ~ 20 -200 m thick, and is ~ 50 m on average (Kadish et al., 2010). Previous workers (Barlow, 2006; Wrobel et al., 2006; Kadish et al., 2009; Kadish and Head, 2011) have suggested that pedestal craters form when the impactor is of insufficient size to excavate substantial ejecta from beneath the surface snow and ice layer. This hypothesis predicts that (1) pedestal craters should begin to exhibit non-ice silicate regolith ejecta at larger diameters; and (2) thicker surface ice deposits should allow for the formation of larger pedestal craters. Thus, by assessing the diameter distribution and presence of ejecta on pedestal craters, we may estimate the diameters of pedestal craters in the LNIH scenario. In this research, we use the pedestal crater database generated by Kadish and Head (2011), who measured the diameters and thicknesses of 2287 pedestal craters. We

evaluate a random subset of this population composed of 1093 pedestal craters. We examined each of these craters using CTX images (~ 6 m/pixel) to determine whether the crater exhibited resolvable ejecta present overlying the pedestal (e.g., Fig.3), indicating that the crater did excavate a small amount of non-ice silicate regolith from beneath the surface ice. In order to find the maximum pedestal crater diameter at a given ice thickness, we binned the data in 1 m thickness increments and plotted the maximum diameter for each bin.

The data show that there exists a weak trend of increasing crater diameter as a function of ice thickness (black squares and black line, Fig. 12A). Pedestal craters which exhibit resolvable ejecta (accounting for $\sim 22\%$ of pedestal craters; blue squares, Fig. 12A) are larger than pedestal craters with no resolvable ejecta by ~ 300 m on average, consistent with the suggestion that progressively larger impactors excavate increasing amounts of ejecta from below the surface ice (e.g., Kadish and Head, 2011; Weiss and Head, 2014). We find that the maximum pedestal crater diameter (D_{max}) in each ice thickness (z) bin (red squares, Fig. 12A) increases with increasing ice thickness, and can be linearly fit with the function, $D_{max} = 11.1z + 1648$ (in meters), with an R^2 value of 0.30. Interestingly, the slope of this line (11.1) is similar to the slope predicted for the depth of excavation (10 ± 2): the excavation depth (d) of an impact crater of transient diameter (D_T) can be approximated as $d = 0.1(\pm 0.02)D_T$ (Croft, 1980; Melosh, 1989; Spudis, 1993; Wieczorek and Phillips, 1999; Hikida and Wieczorek, 2007; Potter et al., 2012). This suggests that the depth of excavation in relation to the ice sheet thickness may play a large role in the formation of pedestal craters: i.e., thicker surface ice sheets can accommodate larger pedestal craters. In order to validate these results for larger

diameters and ice thicknesses, we co-plot the 23 km diameter McMurdo crater in Fig 10B, where the transient crater diameter, $D_T = D_{SC}^{0.15 \pm 0.4} D_R^{0.85 \pm 0.04}$ (Croft, 1985); D_{SC} is the simple-complex crater transition diameter (~ 6 km on Mars; Robbins and Hynek, 2012), and D_R is the rim-to-rim crater diameter. McMurdo crater (Fig. 13) (Tanaka et al., 2000; Schaller et al., 2005) formed exclusively in the ~ 1.5 km thick ice/dust south-polar layered deposits, and like pedestal craters, is an example of a large impact crater that did not excavate material from beneath the icy target. McMurdo crater appears to follow the D_{max} trend remarkably well (Fig. 12B), supporting the hypothesis that thicker ice sheets can accommodate larger pedestal craters.

We now use the D_{max} linear fit function to estimate the pedestal crater diameters that may result due to thicker surface ice deposits, such as in the LNIH scenario. Pedestal craters in the Amazonian typically impact into ~ 50 m of ice and are $< \sim 2$ km in diameter (Kadish et al., 2009). We estimate that thicker snow and ice deposits (for example, ~ 300 m) in the LNIH scenario (Fastook and Head, 2015) may increase the maximum sizes of pedestal craters to ~ 5 km (Table 2). Similarly, all craters under ~ 13 km in diameter would be predicted to be pedestal craters if 1000 m of ice were above the 1 km ELA; this is equivalent to ~ 10 times the current Amazonian surface/near-surface ice supply (Carr and Head, 2014). We find that the maximum diameter of pedestal craters able to form in surface ice in the LNIH model is highly dependent upon the thickness of the ice. For the most plausible LNIH scenario ice thicknesses considered by Fastook and Head (2015) ($< \sim 700$ m), pedestal craters would be predicted to form with diameters up to ~ 10 km (Table 2).

What would be the fate of such craters? Craters of this size are predicted to be

entirely eliminated subsequent to the removal of surface ice in the highlands (by melting and/or sublimation). This may contribute to the observed paucity of smaller Noachian highland craters (Fig. 10). For example, if the post-Noachian removal of craters ≤ 4 km in diameter documented by Irwin et al. (2013) was caused by the removal of LNIH surface ice (instead of a discrete resurfacing event), it would require that a regional ice sheet ~ 300 m thick was present on the highlands prior to their erasure. Furthermore, impact craters that penetrate through the surface ice will produce much shallow crater cavities in the underlying bedrock relative to normal craters of their diameter (as shown in Fig. 6), and could drastically effect the observed size-frequency distributions on Late Noachian-aged terrains.

This model predicts that larger impact craters may appear smaller after the surface ice sheet is removed (e.g., Fig. 6C). How might the removal of surface ice effect the observed crater size-frequency distributions? In order to address this question, we model the effects of impact into surface ice on a martian crater size-frequency distribution. In order to find the final diameter (D_F) following the removal of a given constant thickness of surface ice (z), we equate each impact crater (with diameter, D) as a parabola (e.g., Garvin et al., 2000), such that:

$$D_F = \left(\frac{D^2(d-z)}{d} \right)^{0.5} \quad (6)$$

where d is crater depth. We use d/D functions for rocky targets as opposed to icy ones (Schenk, 2002) because the thickness of the surface ice is generally minor compared with the depth of larger craters. Schenk (2002) found that smaller craters ($\sim < 10$ km in

diameter) forming on the Galilean satellites exhibit d/D ratios that are virtually identical to their rocky lunar counterparts, and so in the case of smaller craters (in which the surface ice will comprise a larger fraction of the crater depth), a rocky d/D ratio is expected to be representative. We determined d from the globally averaged martian crater depth functions found by Robbins and Hynek (2012): $d = 0.097 \left(\frac{D}{2}\right)^{1.061}$ for simple craters (<6 km in diameter on average; Robbins and Hynek, 2012), and $d = 0.250 \left(\frac{D}{2}\right)^{0.527}$ for complex craters. Results for this function are shown in Fig. 14A. For comparison, the large pedestal crater located in the Dorsa Argentea Formation (Fig. 6A) and McMurdo crater (Fig. 13) are co-plotted. Our simple model results are in good agreement with the observed diameters, final diameters, and icy substrate thicknesses for these two observations. We now apply this function to assess how the removal of surface ice might affect the observed crater-size frequency distribution. We model a surface starting with an age of 4 Ga that is accumulating craters through time (with the frequency-chronology function from Neukum et al. (2001) and Hartmann (2005)). In our model, all impact craters that accumulate between 4 Ga and 3.5 Ga are formed in surface ice. We then simulate a resurfacing event at 3.5 Ga by removing the surface ice for comparison with the observed Late Noachian Highlands crater-size frequency data (Irwin et al., 2013; Robbins and Hynek, 2012). The crater size-frequency data (using the Hartmann (2005) isochrons) is modified by the ice removal to account for the diameter change in the crater (i.e., D to D_F). Our results are shown in Fig. 14B for surface ice thicknesses of 300 m, 700 m, 1 km, and 1.5 km. Our model shows that this scenario, wherein impact craters form in surface ice which is later removed, shallows the observed crater size-frequency distribution substantially for craters <32 km in diameter (Fig. 14B),

even without any erosional/infilling processes. Thicker surface ice sheets appear to produce more shallowing of the crater size-frequency distribution at progressively larger diameters. Interestingly, the model with a 1.5 km thick ice sheet appears to well-reproduce the observed Late Noachian highland unit data (Irwin et al., 2013; Robbins and Hynek, 2012) below 32 km in diameter (red squares; Fig. 14B). On this basis, we suggest that impact craters forming in surface ice sheets in the LNIH scenario may plausibly contribute to the observed crater-size frequency distribution (e.g., paucity of small craters). Future work is required to better determine robust ice thicknesses in order to assess the degree to which this process could contribute to the observed crater size-frequency distribution in the LNIH scenario. In summary, the LNIH climate model provides several new options for the interpretation of degradation processes in Noachian craters.

6.5. Elevation control on degradation

Craddock and Maxwell (1993) performed crater counts on terrains at different elevations in the Noachian highlands (Fig. 15). By plotting $N(x)$ (where $N(x)$ = number of craters $\geq (x)$ km diameter per 10^6 km^2), they showed that the relatively fresh portion of the crater population (those with sharp rims and observable ejecta) increased in frequency with increasing elevation for $N(2)$ and $N(5)$ crater counts. They showed further that this trend of more fresh craters at higher elevations was not observed for larger craters ($N(16)$ and $N(50)$). Finally, they showed that the degraded crater population increased with increasing elevation for $N(2)$, $N(5)$, and $N(16)$. Craddock and Maxwell (1993) interpreted these trends to suggest that the resurfacing of the $N(16)$ crater population (this population

increased with increasing elevation for the degraded population but not the fresh population) ceased in the late Noachian. The smaller crater diameter populations (i.e. N(2) and N(5)), however, were progressively affected by resurfacing with time. On the basis of these observations, they concluded that degradation ceased at high elevations first, and then continued at progressively lower elevations through the early Hesperian. Irwin et al. (2013) used crater counts and geologic maps (Fig. 11) to divide the Noachian terrains into Late Noachian (3.57 to 3.85 Ga; Hartmann, 2005; Werner and Tanaka, 2011), Middle (3.85 to 3.96 Ga; Hartmann, 2005; Werner and Tanaka, 2011), and Early Noachian (>3.96 Ga; Hartmann, 2005; Werner and Tanaka, 2011). They found that in general, Early Noachian terrains are present at higher elevations. Irwin et al. (2013) further observed that craters present on the Middle Noachian terrain (lower elevation) exhibit shallower depths than their (older) Early Noachian terrain (higher elevation) counterparts, which they attributed to greater degrees of infill from gravity-driven (e.g., fluvial) processes, as opposed to airfall mantling. Craddock and Maxwell (1993) and Irwin et al. (2013) both interpreted their respective observations to indicate that lower elevation Noachian terrains were resurfaced more recently. Craddock and Maxwell (1993) explained these observations in the context of a warm and wet early Mars, wherein liquid water precipitation and consequent fluvial erosional processes transitioned to lower elevations as the atmosphere thinned through time by escape into space.

Could these observations also be explained in the context of a cold and icy early Mars? In the LNIH scenario, regional snow and ice deposits would be located at high elevations above the equilibrium line altitude of ~1 km (Fig. 16A). Any top-down melting of these ice deposits would produce runoff on top of the ice sheet and drainage

downhill to produce valley networks (Fig. 16B). As the ice sheet melted (predicted to be progressively from lower elevations to higher elevations; Fastook and Head, 2015) fluvial activity would be favored at and near the ice margins, and lower elevation areas would serve as sinks for fluvially transported material. Collectively, these processes would result in preferential resurfacing of lower elevation terrains (e.g., Irwin et al., 2013). As the LNIH ice sheet receded the high areas would remain as sources, while the low areas would be sediment sinks, and eradication and burial would be concentrated there. Even prior to melting, due to ice cover, the LNIH scenario could result in contributions to the trends observed by Craddock and Maxwell (1993) and Irwin et al. (2013) (Fig. 16). For example, thicker ice at low elevations would result in large numbers of pedestal craters in the ≤ 5 km diameter range, and these craters would not survive the LNIH ice loss, potentially producing the observed paucity of small craters (N(5) and N(2)) in the ~ 1 -3 km range. The emplacement of Hesperian ridged plains in low-lying areas with low slopes may also contribute to the paucity of small craters at these elevations (Irwin et al., 2013). In summary, the LNIH model could plausibly account for several of the trends in the observed crater population noted by previous workers.

7. Discussion of Candidate Noachian Crater Degradation Processes

What are the candidate processes for the degradation state of Late Noachian highlands craters (Table 3)? Aeolian mantling (Hartmann, 1971; Wilhelms and Baldwin, 1989; Grizzaffi and Schultz, 1989; Grant and Schultz, 1990; Moore, 1990; Barlow, 1995) is capable of eroding and depositing infilling crater floors. It appears insufficient, however, to explain the degraded nature of Noachian highland craters because crater

infilling was spatially variable in the highlands (Irwin et al., 2013). Furthermore, aeolian processes should have infilled the comparatively shallower valley networks, which does not appear to be the case (Craddock and Howard, 2002). Groundwater sapping has been suggested to account for the degradation state of Noachian highland craters (Gurnis, 1981; Pieri, 1980). It is unclear, however, how groundwater would be recharged within the crater rim crest that is required to produce substantial rim and ejecta erosion. Another degradation processes is impact-induced seismic liquefaction (Clifford, 1997). This process would also be likely to eliminate the smaller valley networks, which is not observed, and thus this mechanism is unlikely to exclusively explain highland crater modification (Craddock and Howard, 2002).

An attractive explanation for Noachian crater degradation is rainfall precipitation and associated fluvial erosion (Craddock and Maxwell, 1993; Craddock et al., 1997; Hynek and Phillips, 2001; Craddock and Howard, 2002; Howard et al., 2005) (Table 3). Recent climate models (Wordsworth et al., 2013), however, predict that widespread liquid water precipitation (rainfall) is unlikely to occur on early Mars, and that instead Mars was more likely to have been characterized by an icy highlands. Wordsworth et al. (2015) modeled the predicted geographic distribution of rainfall in a warm and wet early Mars scenario and found that rainfall patterns were not consistent with the observed geographic distribution of valley networks (Hynek et al., 2010). Another option for crater degradation processes is precipitation as snow (the LNIH model) and subsequent melting by anomalous global heating processes (volcanism, impact). On the basis of our observations, we find that crater formation and modification in a cold and icy early Mars is a plausible mechanism in explaining many of the key aspects of their degraded state

(Fig. 1). For example, upon impact, a crater forming in snow and ice will have a smaller cavity in the target rock and will be characterized by a structurally uplifted rim crest composed partially of ice (Fig. 6C); these factors will contribute to a subdued rim and shallower floor following ice removal, and the observed crater size-frequency distribution (Fig. 14B). Subsequent to crater formation, backwasting of the rim material is predicted to contribute to subdued rims and infilled floors, and any top-down melting or conduction-induced melting from hot ejecta (Morgan and Head, 2009; Kite et al., 2011; Mangold, 2012) of near-rim ice would contribute to subdued rims, infilled floors, and form superposing channels. Deposition of ejecta on top of snow and ice deposits during the impact event could significantly raise basal temperatures, facilitating basal melting (Fig. 9). This process could melt underlying ice, thereby eroding ejecta and generating channels superposing the rim crest, where the ejecta facies is thickest. Widespread melting of surface snow and ice deposits in a potentially volcanically-induced atmospheric thermal pulse (for example, generated by the emplacement of the Hesperian ridged plains; Halevy and Head, 2014) could potentially result in widespread fluvial erosion of craters and the complete removal of crater cavities that do not penetrate the ice (i.e. pedestal craters). This process could contribute significantly to the observed paucity of smaller Noachian craters (Figs. 10 and 14).

8. Conclusions

Compared with young, fresh craters on Mars, Noachian-aged highland craters (Fig. 1) appear substantially more degraded, exhibiting shallow floors, subdued rims, superposed near-rim channels, an absence of ejecta, and a paucity of small craters (Masursky et al.,

1977; Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Mangold et al., 2012). The degraded state of Noachian highland craters has been attributed previously to fluvial runoff in a warm and wet early Mars (Craddock and Maxwell, 1993; Craddock et al., 1997; Hynek and Phillips, 2001; Craddock and Howard, 2002). Climate models, however, have shown that early Mars may have been too cold to support liquid water (rainfall) precipitation (e.g., Kasting, 1991; Haberle et al., 1993; Mischna et al., 2013; Wordsworth et al., 2013; von Paris, 2014).

We develop further criteria to help test whether the Late Noachian craters could have been degraded in a cold and icy early Mars climate environment. In this scenario (Forget et al., 2013; Wordsworth et al., 2013), the thicker early martian atmosphere becomes thermally coupled to the surface, producing an adiabatic cooling effect that results in the concentration of several hundred meter thick snow and ice deposits in the Noachian southern highlands (Late Noachian Icy Highlands (LNIH) model; Wordsworth et al., 2013). We find that impact into Late Noachian hectometers-thick snow and ice deposits could potentially yield craters analogous to Amazonian double-layered ejecta and pedestal craters. Craters forming in hectometers-thick surface snow and ice are predicted to have shallower cavities in the underlying target rock, and lower rims subsequent to removal of the ice. Further modification (Table 4) by (1) backwasting of rim-crest material; (2) solar-insolation induced top-down melting of proximal rim-crest surface ice; (3) melting of ice and fluvial erosion due to contact with hot ejecta (Morgan and Head, 2009; Kite et al., 2011; Mangold, 2012); (4) basal melting of ice (enhanced by overlying low thermal conductivity ejecta); and (5) volcanic- or impact-induced atmospheric warming pulses, could infill crater cavities, erode crater rims and ejecta, and produce

superposing channels. Removal of the surface snow and ice (through melting or sublimation) in a subsequent climate regime could preferentially eliminate smaller craters that formed exclusively within the surface ice deposits (i.e., pedestal craters), and could drastically modify the observed crater size-frequency distribution by reducing the apparent diameters of larger craters.

In summary, we find that snow and ice deposits present in the highlands during the Late Noachian provide a plausible alternative hypothesis to explain several aspects of the degraded state of Noachian highland craters. The degradation state of Noachian highland craters represents one of many distinctive fingerprints of processes operating in the early martian climate regime. In order to assess further the plausibility of these processes in the Noachian the following studies and tests could be undertaken: 1) Is basal ice-melting near the rim of Noachian-age craters capable of substantial erosion? A more detailed thermal/basal melting model could be conducted in conjunction with detailed morphological analyses of degraded highland craters to assess this. 2) What types of fluvial features might be predicted in an ejecta-on-ice basal melting scenario? Special attention should be given to testing observed fluvial feature morphologies against recorded subglacial channel morphologies (e.g., Hobbie et al., 2014) and drainage patterns (e.g., Booth and Hallet, 1993), in addition to the potential for fluvial channels which may propagate uphill due to hydraulic head from the overlying ice sheet (e.g., over crater rim-crests; Irwin et al., 2005a). 3) Fluvial channels have been observed on the ejecta facies of Amazonian impact craters hypothesized to form in icy surface/near-surface layers (Morgan and Head, 2009; Mangold, 2012). Future work is required to assess the potential for such fluvial activity in a cold and icy early Mars, where an

elevated geothermal heat flux could provide more meltwater following the impact. 4) Crater size-frequency simulations should be further evaluated under a wider parameter range of starting surface and resurfacing ages to assess the conditions under which the observed crater size-frequency distribution can be reproduced by the removal of surface ice sheets. 5) Future tests of other Late Noachian features are required to assess the strengths and weaknesses of the LNIH model. These include: the thermal/melting response of LNIH ice sheets (Cassanelli and Head, 2015; Fastook and Head, 2015), the nature of the Late Noachian atmospheric pressure (e.g., Scanlon et al., 2013, Kite et al., 2014), the predicted Late Noachian ice supply (Andrews-Hanna and Lewis, 2011; Carr and Head, 2014), valley network formation in the LNIH scenario (e.g., Cassanelli and Head, 2014; Wordsworth et al., 2015), and the distribution of ice and its behavior in the LNIH model (Fastook and Head, 2015; Wordsworth et al., 2015).

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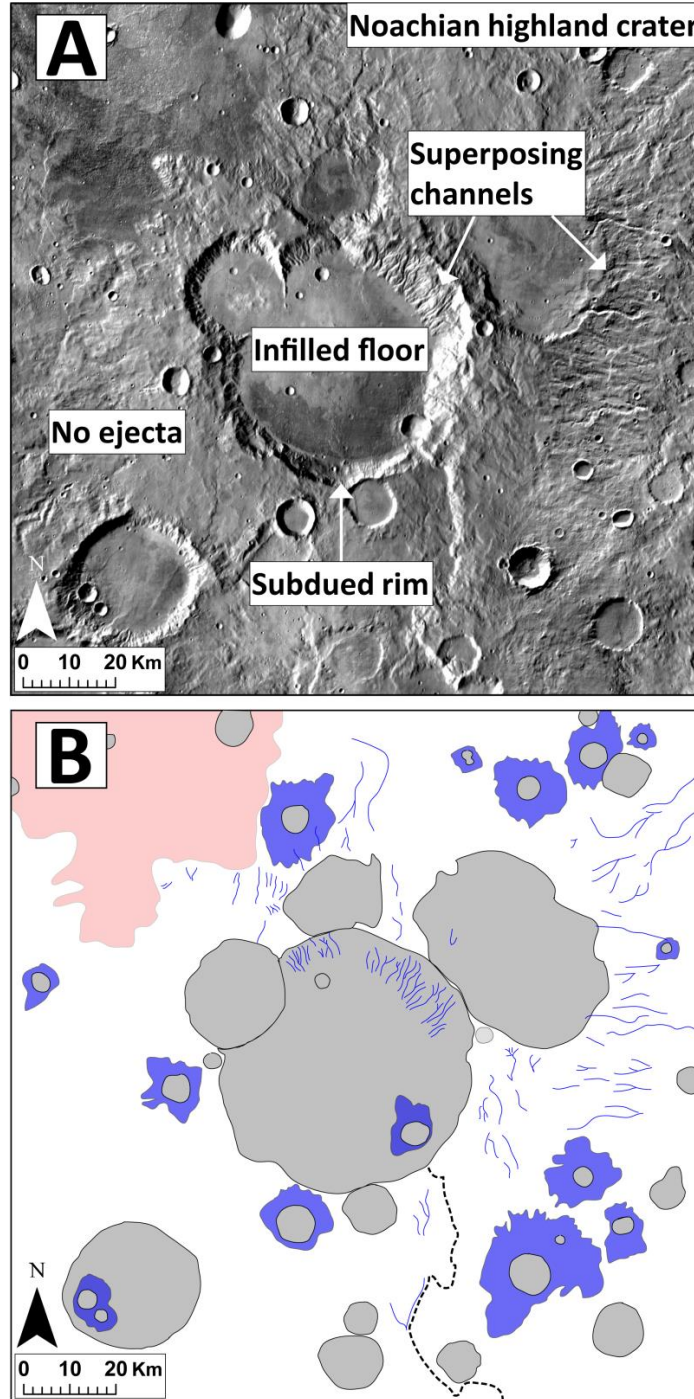


Figure 1. Map views illustrating typical degraded Noachian highland crater characteristics. (A) Map view of a 55 km diameter Noachian highland crater and characteristics (149°E, 36°S, THEMIS daytime mosaic). (B) Sketch map of (A) (Key: craters (grey), ejecta (blue), fluvial channels (blue lines), and plateau contact (dashed black line), low-lying region (red)).

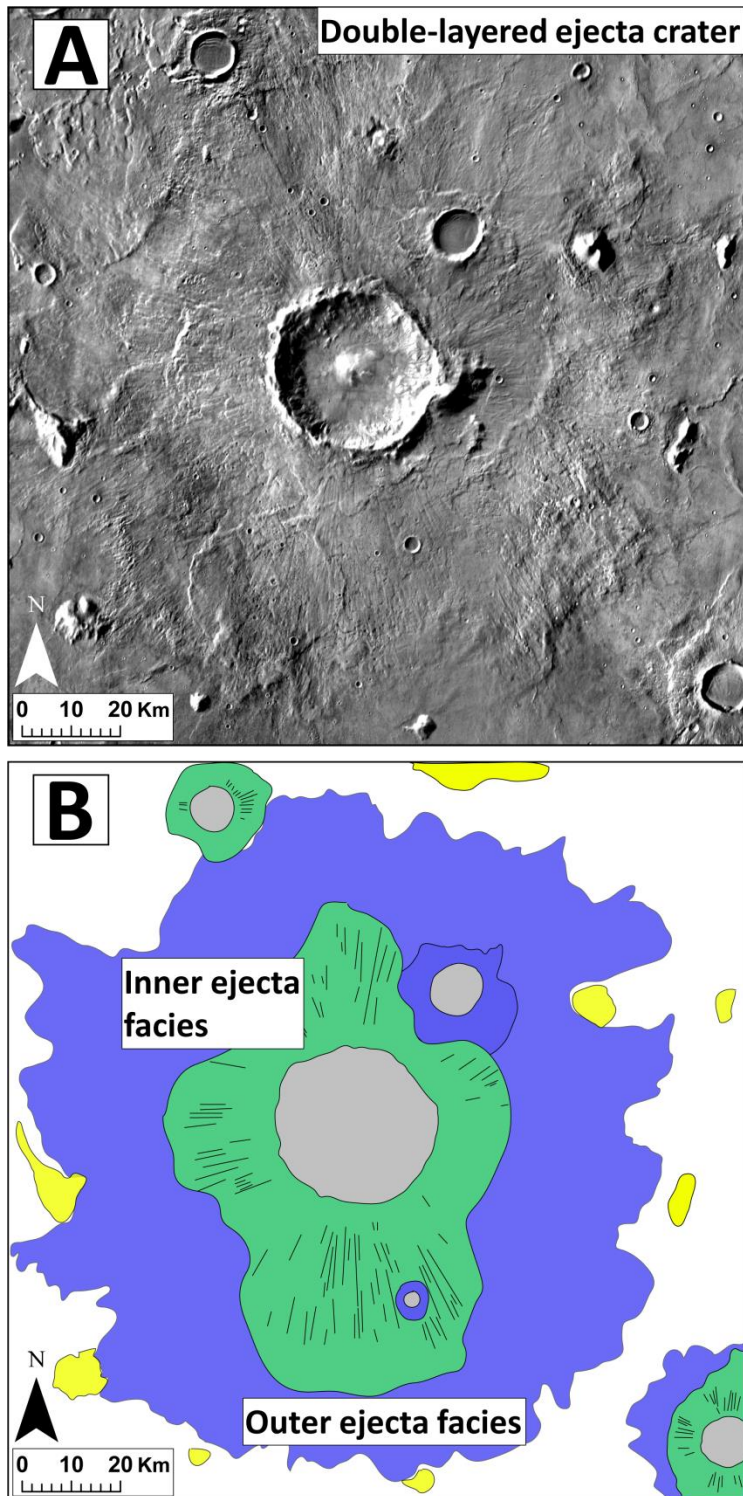


Figure 2. Map views illustrating typical double-layered ejecta (DLE) craters characteristics. (A) Example of a 33 km diameter double-layered ejecta crater (115°E, 35.6°S, THEMIS daytime mosaic). (B) Sketch map of (A) (Key: craters (grey), ejecta (blue), inner ejecta facies (green), longitudinal grooves (black lines), and high-standing topography (yellow)).

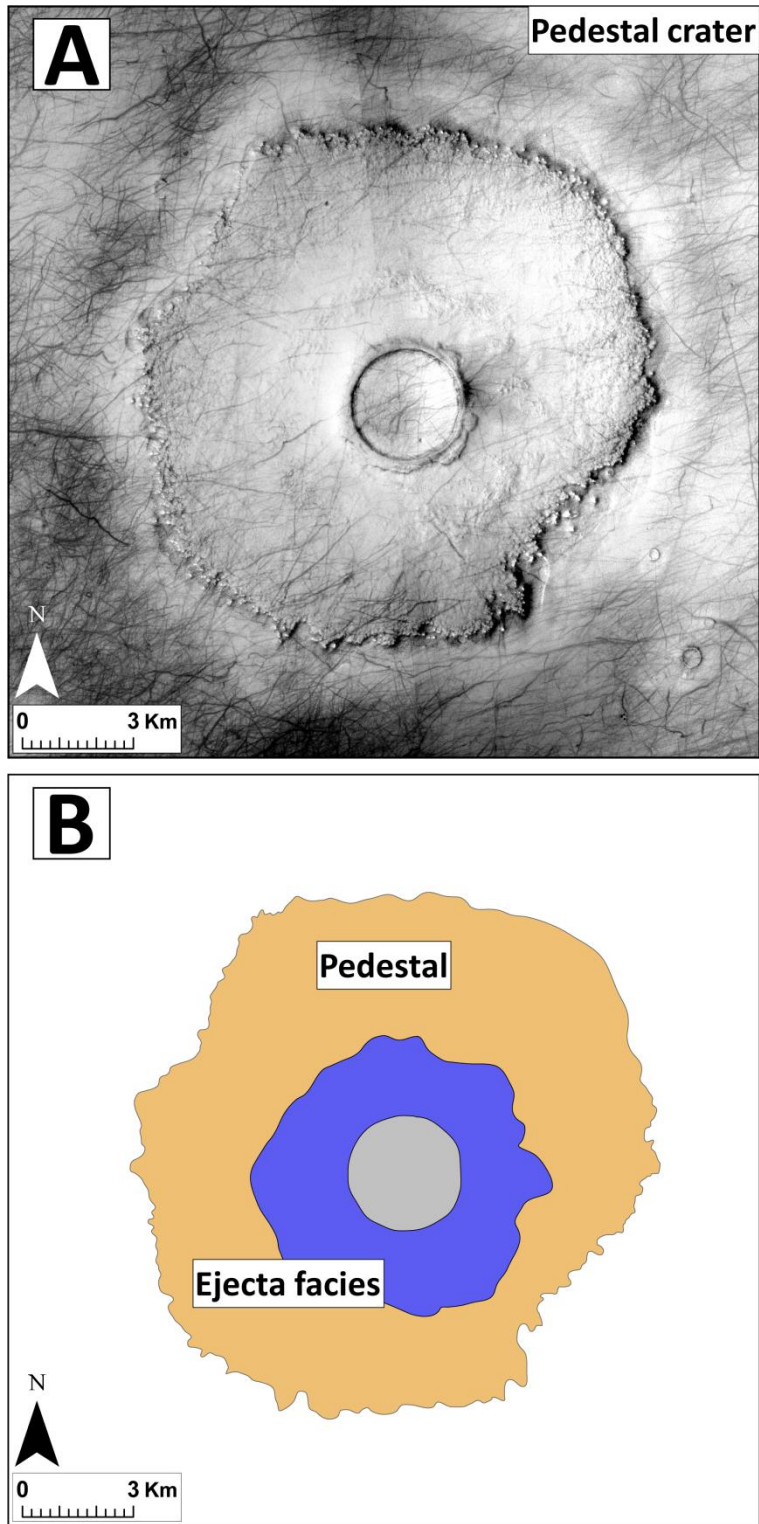


Figure 3. Map views illustrating typical pedestal craters characteristics. (A) Example of a 3 km diameter fresh (Amazonian) pedestal crater with a ~72 m thick pedestal (Kadish et al., 2010) (91.8°E, 55.3°N, CTX images G23_027110_2354 and G21_026477_2355). (B) Sketch map of (A) (Key: crater (grey), ejecta (blue), and pedestal surface (orange)).

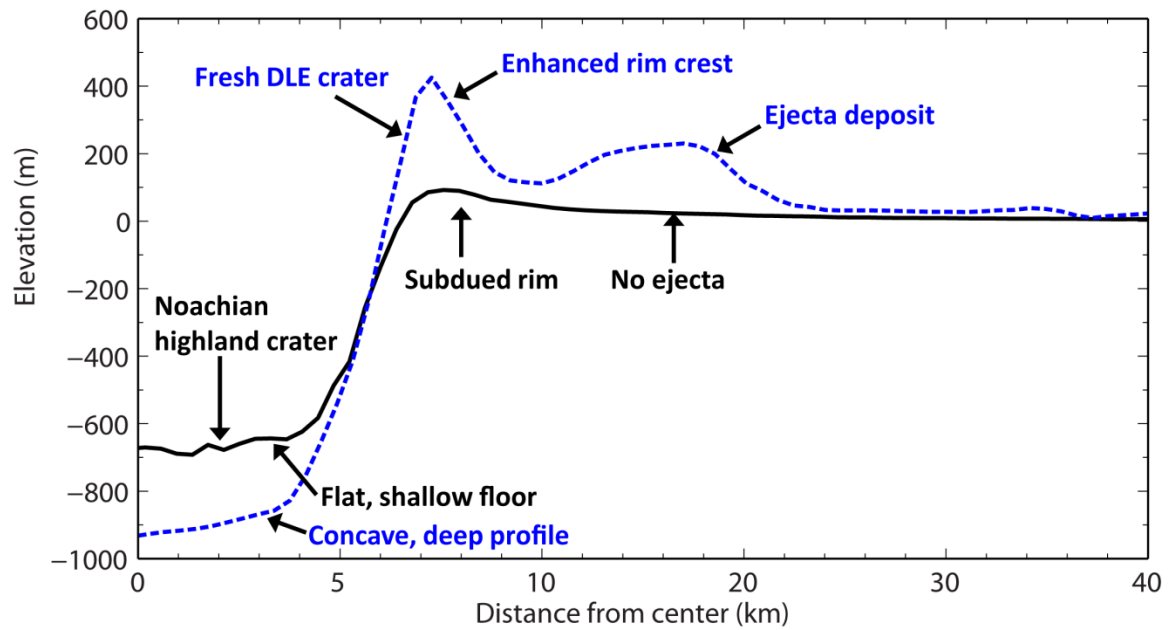


Figure 4. Altimetric profile of a typical 14 km diameter fresh double-layered ejecta (DLE) crater (dashed blue line) and a similarly sized typical Noachian highland crater (black line). MOLA altimetry data (463 m/pixel).

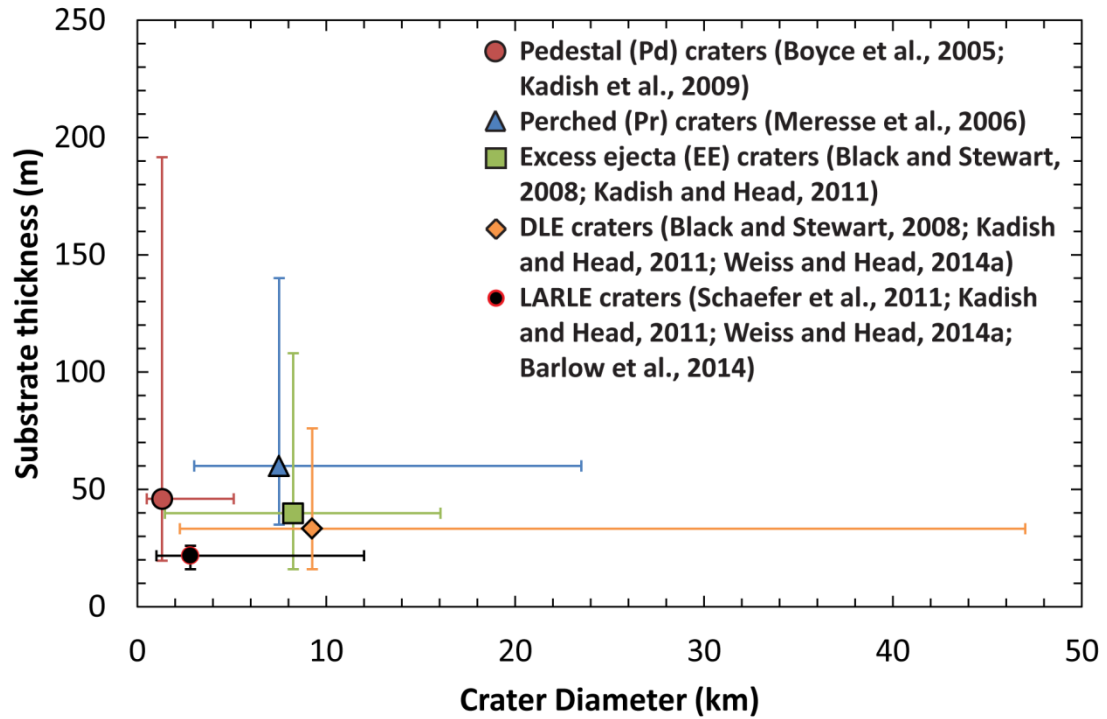


Figure 5. Relationships between crater diameter thickness of ice substrate layer for different crater types: pedestal craters (Kadish et al., 2009), perched craters (Meresse et al., 2006), excess ejecta craters (classified as DLE) (Black and Stewart, 2008; Kadish and Head, 2011), double-layered ejecta craters (excluding LARLE) (Black and Stewart, 2008; Kadish and Head, 2011; Weiss and Head, 2014), and low-aspect-ratio layered ejecta (LARLE) craters (Schaefer et al., 2011; Kadish and Head, 2011; Weiss and Head, 2014; Barlow et al., 2014). Average values for each crater population are plotted (colored markers); horizontal and vertical lines correspond to the range of values (median values are similar). Adapted from Kadish and Head (2011) and Weiss and Head (2014).

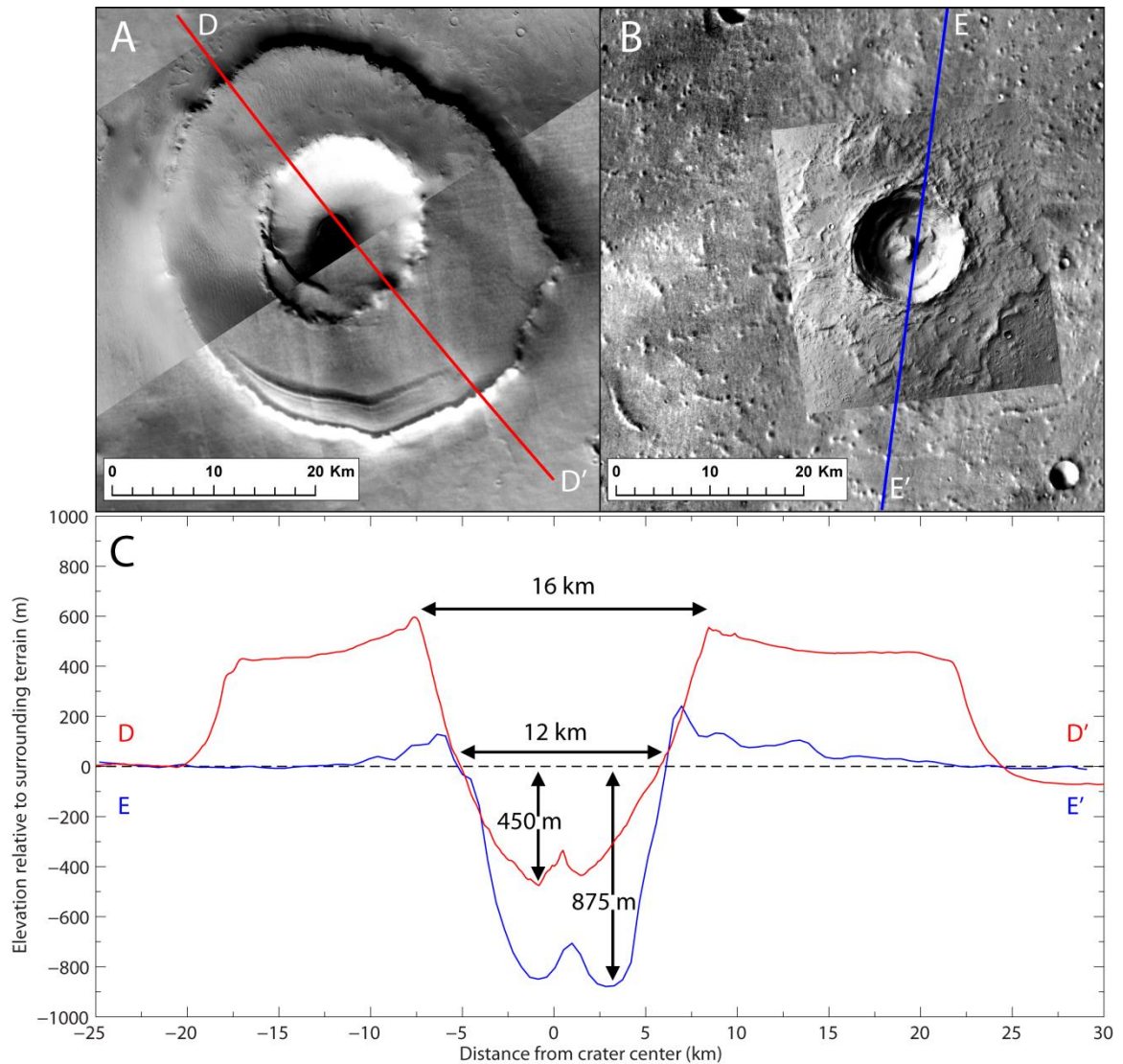


Figure 6. Comparison between pedestal craters and normal craters. (A) An unusually large pedestal crater (16 km in diameter) located in the Dorsa Argentea Formation (-111.47°E , -76.05°S); the pedestal is ~ 500 m thick. Unlike most pedestal craters, the crater cavity clearly extends below the surrounding terrain, as shown topographic profile (red line) in (C). (B) 12 km diameter single-layered ejecta crater (82.97°E , 13.27°N). (C) Topographic profiles comparing the pedestal and single layered ejecta craters from MOLA data from A (D-D'; red line) and B (E-E'; blue line). If the pedestal (red line) were removed, a ~ 12 km diameter depression would remain, but is likely to be infilled more easily relative to typical impact craters of similar size due to its shallower depth (blue line) and lack of rim. CTX images G13_023136_1029 and P22_009797_1934 overlain on THEMIS daytime mosaic.

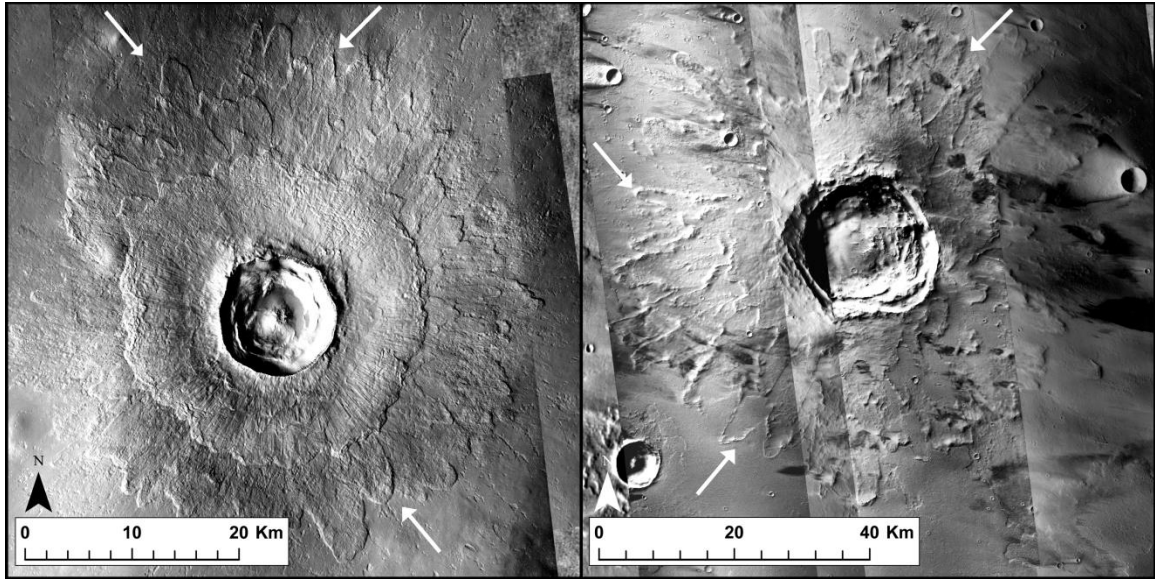


Figure 7. (A) Double-layered ejecta (DLE) crater Steinheim (169.35°W , 54.57°N). (B) Multiple-layered ejecta (MLE crater; 141.8°W , 9.52°S). The outer facies of DLE craters exhibits a similar lobate morphology (arrows) to that observed for the ejecta facies around MLE craters. Larger craters forming in surface ice sheets which do not exhibit inner ejecta facies may thus resemble MLE craters. CTX images P21_009160_2348, P17_007802_2349, D01_027647_2349, G21_026302_2344, P18_008210_1704, B20_017427, P04_002738_1704, P06_003305_1713.

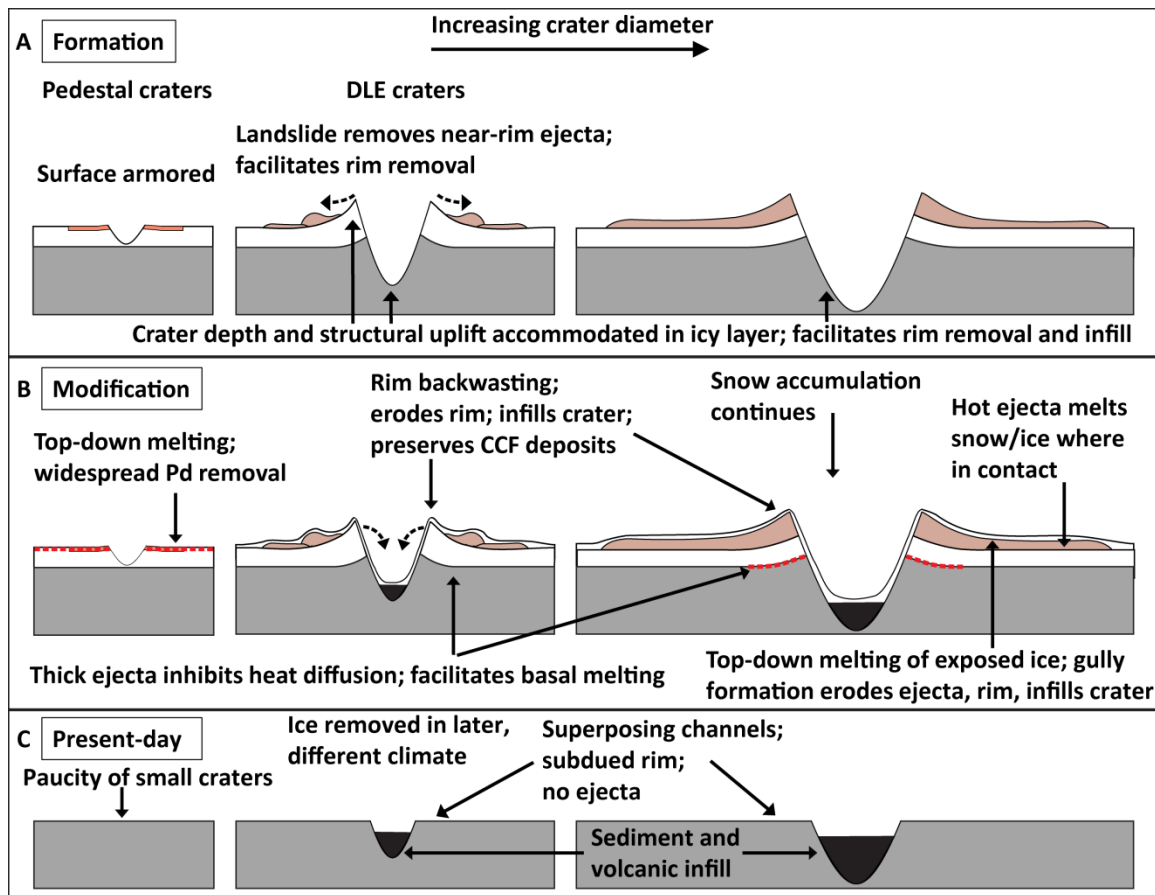


Figure 8. Predicted degradation processes for Noachian highland craters formed with Amazonian morphologies in a LNIH climate scenario. Pedestal craters (left), double-layered ejecta (DLE) craters (middle), and larger craters (right), all formed in ice. (A) Initial state. (B) Modification and degradation during evolution and melting of LNIH layer. (C) Present-day observed morphology.

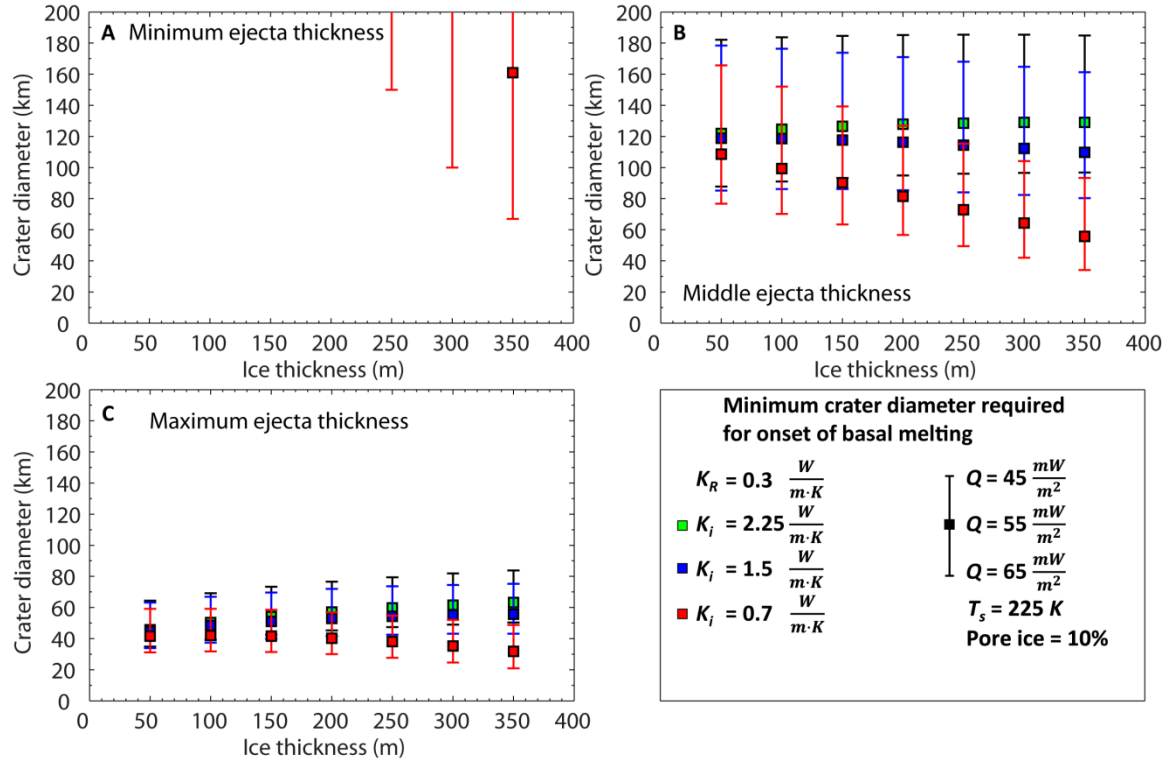


Figure 9. Crater diameters predicted to exhibit basal melting of near the rim. Shown is the minimum crater diameter which generates basal melting (i.e., which produce basal ice temperatures that exceed 273 K) as a function of ice thickness. The different panels represent different ejecta thicknesses: (A) Minimum crater diameters which exhibit basal melting for the minimum ejecta thickness values, (B) middle height values, and (C) maximum height values. The Square markers correspond to the predicted average Late Noachian heat flux ($Q=55 \text{ mW/m}^2$; McGovern et al., 2004), while the error bars correspond to the range in heat fluxes (The bottom error bar is $Q=65 \text{ mW/m}^2$, and the top error bar is $Q=45 \text{ mW/m}^2$). Red marker represent impact into a firm layer ($\kappa_i=0.7 \text{ W/m} \cdot \text{K}$), blue marker corresponds to impact into a firm and ice layer ($\kappa_i=1.5 \text{ W/m} \cdot \text{K}$), and green marker corresponds to impact into a pure ice layer ($\kappa_i=2.25 \text{ W/m} \cdot \text{K}$).

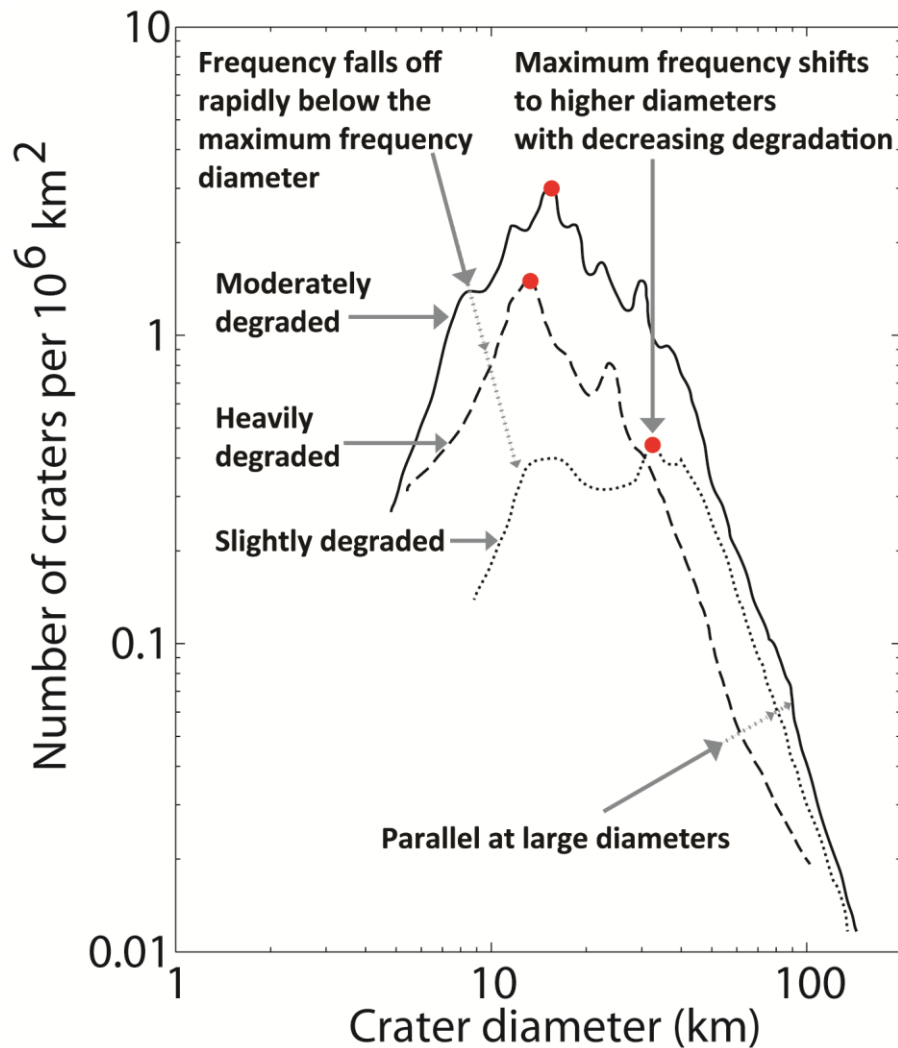


Figure 10. The incremental crater size-frequency distribution of martian impact craters (from Jones, 1974) shows the paucity of small degraded craters in a study area. The study area is south of Arabia Terra (approximately a rectangle extending from 30°W, 0° to 60°E, 20°N). The data were compiled using Mariner 9 data (average 600-650 m/pixel for latitudes 0°-20°N). The Jones (1974) study divided craters into slightly degraded (dotted line), moderately (dashed line), and heavily degraded (solid line). Minor bumps near maximum frequency (red circle) are an artifact of the lower resolution Mariner 9 images. This plot shows the paucity of small degraded Noachian craters relative to their less degraded counterparts.

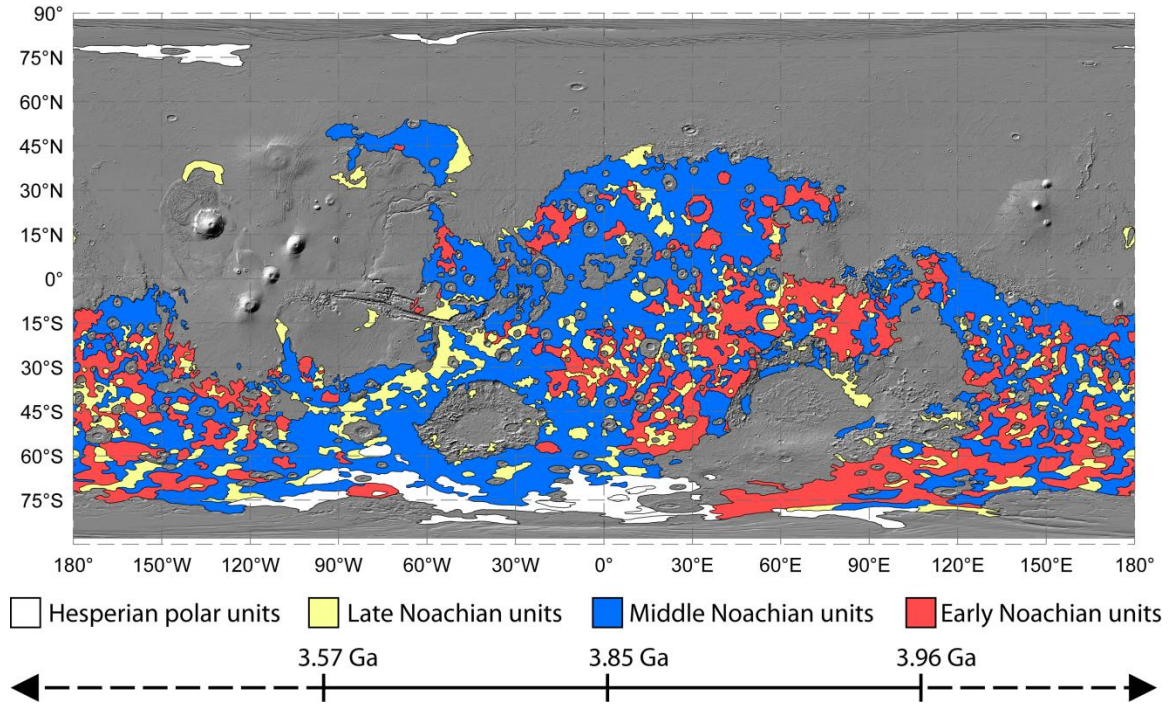


Figure 11. Distribution of Noachian units from Tanaka et al. (2014) overlain on MOLA shaded relief map (Smith et al., 1999). Shown are the Late (yellow), Middle (blue), and Early (red) Noachian units, as well as the Hesperian polar units (white), which comprise the southern circumpolar Dorsa Argentea Formation. Quantitative age boundaries shown in bottom line are from Werner and Tanaka (2011) fit with the Hartmann (2005) production function. Late Noachian and Middle Noachian-aged surfaces are generally at lower elevations than Early Noachian surfaces (Irwin et al., 2013).

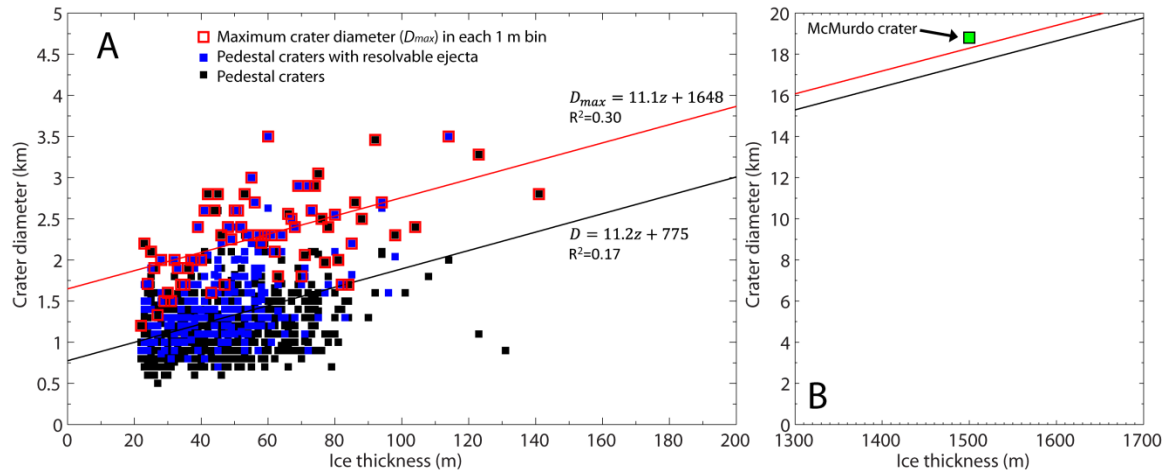


Figure 12. Predictions for the pedestal crater diameters which may form in a given ice sheet thickness. (A) Pedestal crater diameter as a function of ice thickness (derived from pedestal thickness) for a subset (~47%) of the pedestal craters mapped by Kadish et al. (2010). Shown are 1083 craters out of the total 2287 present in the Kadish et al. (2010) database. The population is shown in black squares with its corresponding linear fit (black line; all units in meters). Pedestal craters with resolvable ejecta are shown in blue squares. The maximum pedestal crater in each 1 m diameter bin are shown as red squares with its corresponding linear fit (red line; all units in meters). In order to remove the bias associated with smaller pedestal craters forming in thicker (≥ 60 m) ice which encompassed their own bin, we focused on craters above the linear fit of this sample population (black line) when binning these data. The maximum pedestal crater diameter in each bin (D_{max}) appears to increase as a function of ice thickness. (B) Co-plotted is the 23 km diameter McMurdo crater (Fig. 13) which impacted into the ~1.5 km thick ice/dust south-polar layered deposits, confirming that the D_{max} fit is valid for larger crater diameters. This suggests that the thicker ice sheet hypothesized to be present in the southern highlands in the LNIH scenario would host larger pedestal craters relative to the Amazonian-aged pedestal craters observed today.

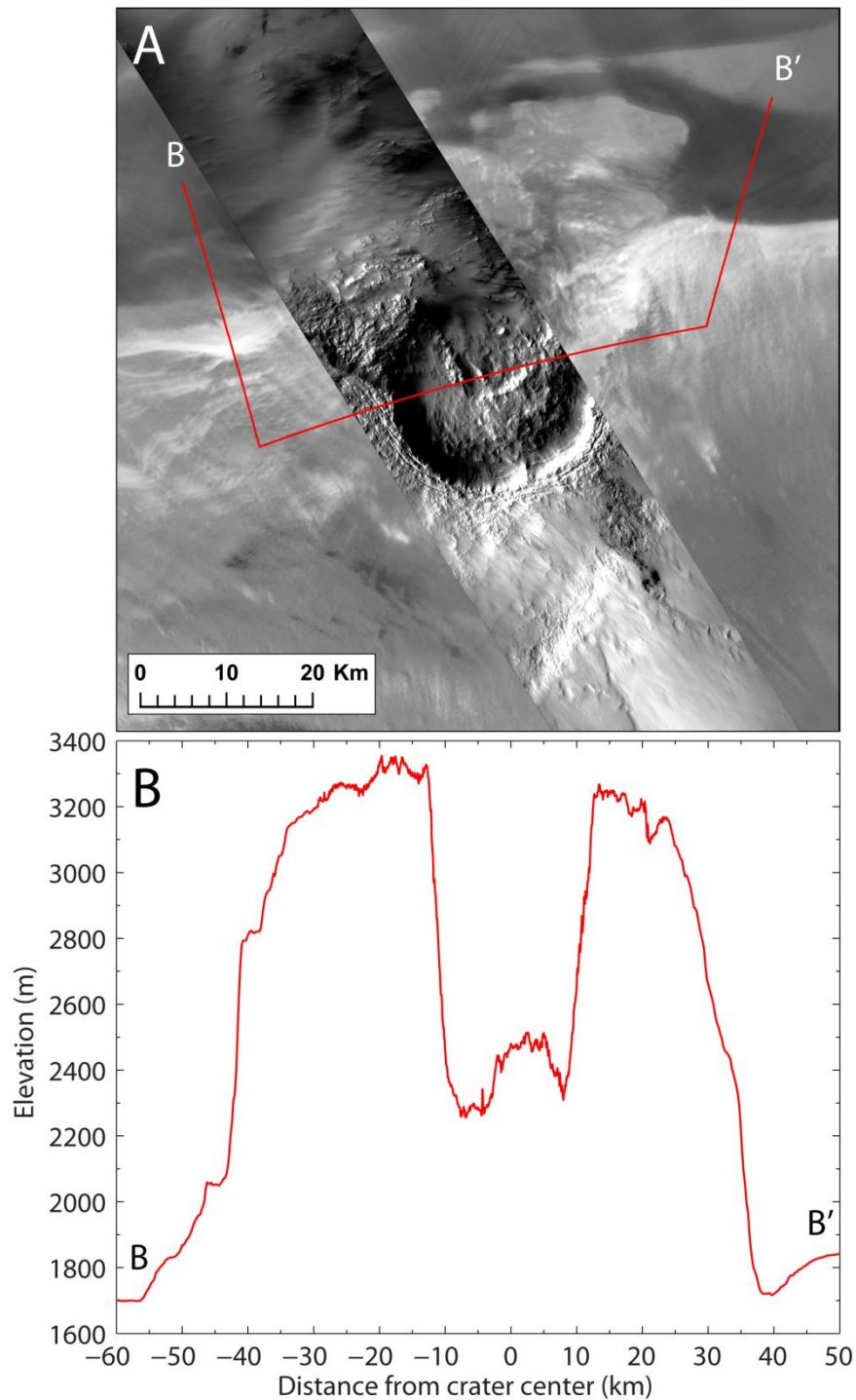


Figure 13. The 23 km diameter McMurdo crater, which formed in ~ 1.5 km of ice/dust south-polar layered deposits (Tanaka et al., 2000; Schaller et al., 2005). (A) McMurdo crater (0.54°E , 84.49°S), CTX image F08_038904_0955 overlain on THEMIS daytime mosaic. (B) Topographic profile B-B' in (A). McMurdo crater might represent a common pedestal crater morphology observed in a hypothesized thick Late Noachian Icy Highlands ice sheet, and would be entirely erased if the underlying ice/dust deposits were removed.

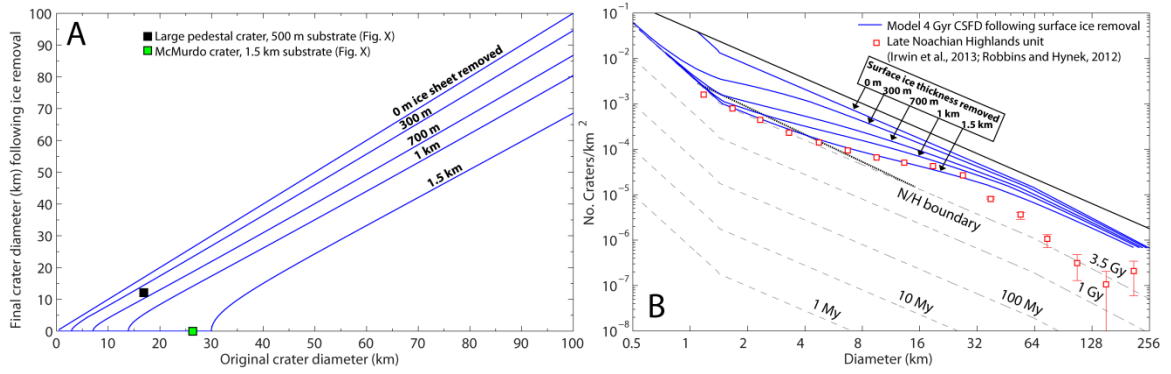


Figure 14. The effects of removing a surface ice layer on the crater size-frequency distribution of craters remaining in the underlying bedrock. (A) The final crater diameter expected following removal of ice sheets between 300 m and 1.5 km thick by equating an impact crater cavity as a parabola (eq. 5). The diameters and icy substrate thicknesses of the large pedestal crater (Fig. 6A) and McMurdo crater (Fig. 13) are consistent with these results. (B) The modeled crater size frequency distribution of a 4 Ga starting surface, where accumulating craters form in surface ice sheets until 3.5 Ga, whereupon the surface ice is removed. The crater size-frequency distribution has been modified (e.g., from (A) to account for the removal of a surface ice sheet ranging between 300 m and 1.5 km thick (blue lines) at 3.5 Ga. The Noachian/Hesperian boundary is shown as a dotted black line, and the Late Noachian Highlands unit (Irwin et al., 2013; Robbins and Hynek, 2012) include both fresh and degraded craters, and are shown as red squares. Isochrons are from Hartmann (2005). The model shows that crater formation in an ice sheet (which is removed at 3.5 Ga) allows the currently observed crater size-frequency distribution to shallow substantially for craters less than 32 km in diameter (blue lines), allowing it to better match the observed Late Noachian Highlands unit (red squares).

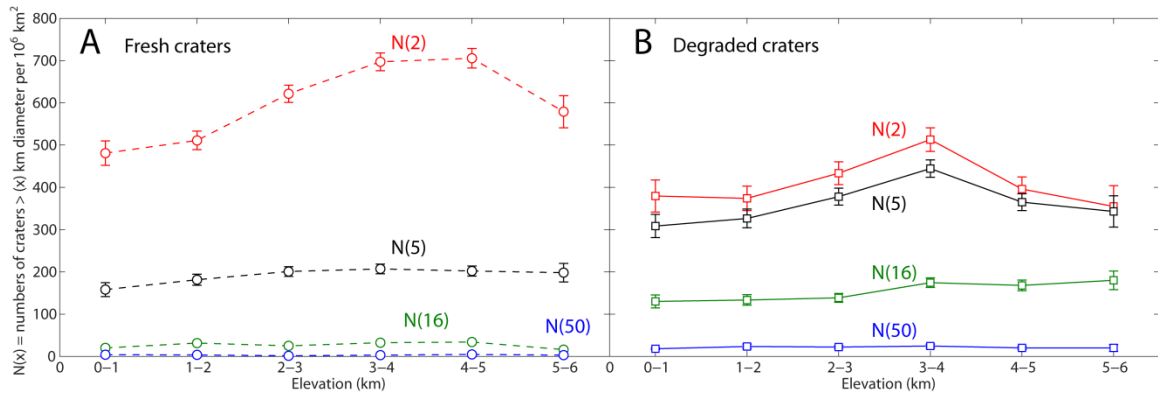


Figure 15. Relative N(x) ages of craters within the combined cratered (Npl1) and dissected (Npld) Noachian terrains (defined from Scott and Tanaka, 1986; Greeley and Guest, 1987); data from Craddock and Maxwell (1993). (A) Fresh crater counts for N(2), N(5), N(16), N(50), where N(X) = number of craters $\geq$ (X) diameter per km diameter per 10⁶ km² (a proxy for relative age). (B) Degraded crater counts. 5-6 km elevation bin corresponds only to craters within the dissected terrain. The fresh craters (A) exhibit older ages at higher elevations for N(2) and N(5), whereas the degraded craters (B) exhibit older ages at higher elevations for N(2), N(5), and N(16). Craddock and Maxwell (1993) interpreted this data to indicate that resurfacing affected younger surfaces with increasing time and decreasing elevation.

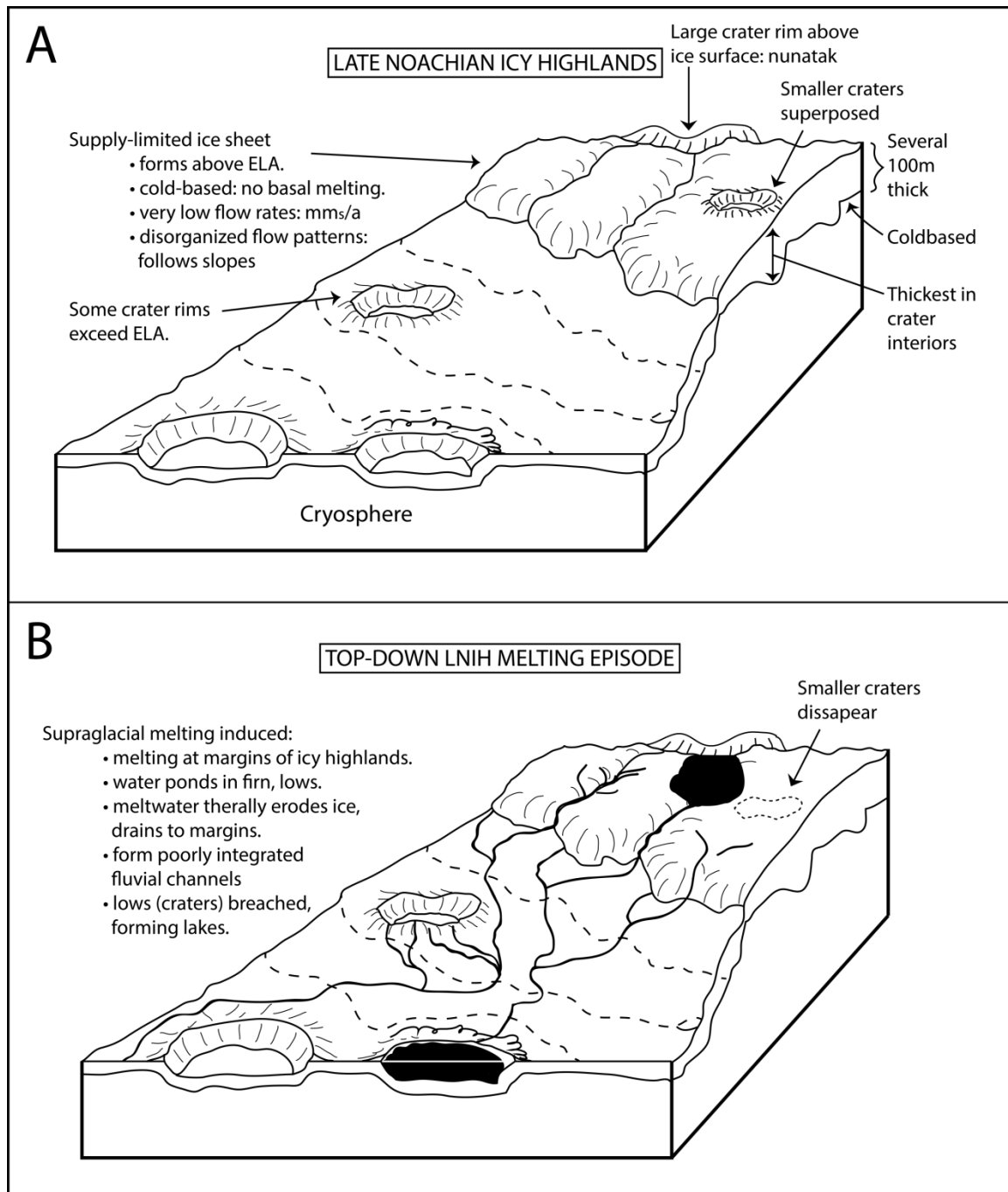


Figure 16. Block diagrams illustrating the concept of the Late Noachian Icy Highlands (LNIH) climate model (Wordsworth et al., 2013) and some implications for impact crater formation and preservation (from Fastook and Head, 2015). (A) Several hundreds of meters of ice cover the Late Noachian icy highlands above an Equilibrium Line Altitude (ELA) of about 1 km. Predicted coldbased glacial behavior (Fastook and Head, 2015) means that modification of the substrate is minimal. During this time, degraded craters from pre-LNIH periods would be preferentially preserved at higher elevations beneath the ice sheet. (B) Top-down melting and retreat of the LNIH ice sheet. As the margins of

the ice sheet melt and transport the meltwater to lower elevations, small to intermediate sized craters are preferentially eroded and buried. Eolian activity redistributes the sediment, further altering craters at lower elevations, and Hesperian ridged plains volcanism fills and buries others. These LNIH observations provide an alternate explanation for: 1) the trend of the increase in the relatively fresh portion of the small-diameter part of the crater population N (2,5) with increasing elevation (older smaller fresh craters could be preferentially preserved under the cold-based ice cover), 2) the lack of a similar trend for larger craters (N (16, 50)) (ice could be insufficiently thick to protect larger craters for some degradation), and 3) the trend of an increased degraded crater population with increasing elevation for N (2,5,16) (preferential preservation of an older degraded crater population under the cold-based ice sheet). Retreat of the ice sheet and delivery of meltwater and sediments to the lowlands could also contribute to the degradation of small craters in the lower elevations.

Morphological characteristics of Noachian highland craters	Reference
Subdued rims	(Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002)
Shallow, flat floors	(Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002)
Superposing channels on crater wall interior and within 2R from the rim.	(Masursky et al., 1977; Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Mangold et al., 2012)
Absence of ejecta deposits and secondary craters	(Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Mangold et al., 2012)
Paucity of small craters (<~10-20 km diameter)	(McGill and Wise, 1972; Jones, 1974; Craddock and Maxwell, 1990)

Table 1. Characteristics of Noachian highland craters.

Ice thickness (m)	Predicted pedestal crater D_{max} (km)	Amazonian ice supply
50	2.2	1x (mid-latitude)
300	4.95	~1x (LNIH)
700	9.42	~5x (LNIH)
1000	12.65	~10x (LNIH)

Table 2. Estimates of the maximum pedestal crater diameter (D_{max}) as a function of ice thickness from Fig. 12. Ice thicknesses correspond to potential volumes of ice (represented as a multiple of the present-day Amazonian ice volume) hypothesized to be present in the LNIH (from Fastook and Head, 2015).

Geologic observation	LNIH	Rainfall and fluvial erosion	Model		
			Airfall mantling	Groundwater sapping	Impact-induced seismic liquefaction
Noachian crater characteristics					
Degraded rim	x	x	x	Unclear how crater rim-crest is recharged with groundwater	?
No ejecta	x	x	x	Unclear how ejecta is eroded away	x
Superposing channels	x	x	Requires additional explanation for fluvial features	x	x
Infilled crater cavity	x	x	x	x	x
Paucity of small craters	x	x	x	x	x
Depth/diameter measurements	x	x	Inconsistent with elevation differences	x	Inconsistent with elevation differences
Valley networks	Requires separate mechanism	x	Inconsistent with presence of small valley networks	Morphology of valley networks inconsistent with formation by groundwater, mechanically difficult	Inconsistent with presence of small valley networks
Climate models	Requires mechanism for melting	Recent climate models unable to produce requisite rainfall pattern and warmer surface temperatures	?	Climate models unable to produce requisite warmer surface temperatures to eliminate ice-cemented cryosphere	Climate models unable to produce requisite warmer surface temperatures to eliminate ice-cemented cryosphere

Table 3. Summary of some Late Noachian characteristics and consistency with different models for the Late Noachian climate. Consistency is denoted with x. Noachian crater characteristics, the presence of valley networks, and the global climate model results are compared for consistency against the Late Noachian Icy Highlands (LNIH) model (Wordsworth et al., 2013), the warm and wet (rainfall and fluvial erosion) scenario (Craddock and Maxwell, 1993; Craddock et al., 1997; Hynek and Phillips, 2001; Craddock and Howard, 2002; Irwin et al., 2005a, b; Howard et al., 2005; Hoke and Hynek, 2009; Hynek et al., 2010; Hoke et al., 2011), and degradation through airfall mantling (Hartmann, 1971; Wilhelms and Baldwin, 1989; Grizzaffi and Schultz, 1989; Grant and Schultz, 1990; Moore, 1990; Barlow, 1995), groundwater sapping (Gurnis, 1981; Pieri, 1980), and impact-induced seismic liquefaction (Clifford, 1997).

Noachian crater characteristics	Mechanism in the LNIH scenario	
	Formation	Modification
No rim	☐ Impact into surface ice: rim uplift in underlying rock is subdued.	☐ Solar-insolation-induced top-down melting of surface snow/ice.
No ejecta	☐ Smaller volume of excavated rocky ejecta to be eroded.	☐ Melting of snow/ice by hot ejecta.
Superposing channels		☐ Basal melting of near-rim ice.
		☐ Volcanically-induced warming pulses enhances fluvial processes.
Infilled crater cavity	☐ Impact into surface ice: shallower cavity in underlying rock.	☐ Backwasting of rim-crest.
		☐ Fluvial transport of material into crater.
		☐ Volcanic infill.
Paucity of small craters		☐ Removal of surface snow/ice in different, later climate regime removes craters formed exclusively in surface ice deposits.

Table. 4. Summary of crater degradation mechanisms in a cold and icy early Mars scenario.

Chapter 3:
Impact ejecta-induced melting of surface ice deposits on Mars

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Abstract

Fluvial features present around impact craters on Mars can offer insight into the ancient martian climate and its relationship to the impact cratering process. The widespread spatial and temporal distribution of surface ice on Mars suggests that the interaction between impact cratering and surface ice could have been a relatively frequent occurrence. We explore the thermal and melting effects on regional surface ice sheets in this case, where an impact event occurs in regional surface ice deposits overlying a regolith/bedrock target. We provide an estimate for the post-impact temperature of martian ejecta as a function of crater diameter, and conduct thermal modeling to assess the degree to which contact melting of hot ejecta superposed on surface ice deposits can produce meltwater and carve fluvial features. We also evaluate whether fluvial features could form as a result of basal melting of the ice deposits in response to the thermal insulation provided by the overlying impact ejecta. *Contact melting* is predicted to occur immediately following ejecta emplacement over the course of hundreds of years to tens of kyrs. *Basal melting* initiates when the 273 K isotherm rises through the crust and reaches the base of the ice sheet ~ 0.1 to ~ 1 Myrs following the impact. We assess the range of crater diameters predicted to produce contact and basal melting of surface ice sheets, as well as the melt fluxes, volumes, timescales, predicted locations of melting (relative to the crater), and the associated hydraulic and hydrologic consequences. We find that the heat flux and surface temperature conditions required to produce contact melting are met throughout martian history, whereas the heat flux and surface temperature conditions to produce basal melting are met only under currently understood ancient martian thermal conditions. For an impact into a regional ice sheet, the contact and basal melting mechanisms are predicted to generate melt volumes between $\sim 10^{-1}$ and

10^5 km^3 , depending on crater diameter, ice thickness, surface temperature, and geothermal heat flux. Contact melting is predicted to produce fluvial features on the surface of ejecta and the interior crater walls, whereas basal melting is predicted to produce fluvial features only on the interior crater walls. Before basal melting initiates, the ice-cemented cryosphere underlying the crater ejecta is predicted to melt and drain downwards through the substratum, generating a source of water for chemical alteration and possibly subsurface clay formation. These candidate melting processes are predicted to occur under a wide range of parameters, and provides a basis for further morphologic investigation.

1. Introduction

The wide array of fluvial features present on the surface of Mars, despite its current below-freezing surface temperatures, has raised many questions regarding the climate of Mars throughout its early history. Fluvial channels incised onto impact crater walls, rims, and ejecta (e.g., Craddock and Maxwell, 1993; Craddock and Howard, 2002; Howard, 2007; Morgan and Head, 2009; Mangold, 2012; Mangold et al., 2012b; Hobbs et al., 2016) are particularly interesting because these features may offer insight into the conditions of the ancient martian climate, and its relationship to the impact cratering process. Fluvial channels have been reported around the rims and ejecta facies of impact craters spanning the entire martian geologic record. These include (1) Amazonian- and Hesperian-aged craters which exhibit fluvial channels superposing the crater walls and ejecta facies (Fig. 1A) (Morgan and Head, 2009; Howard and Moore, 2011; Mangold, 2012; Mangold et al., 2012a; Schon and Head, 2012; Hobbs et al., 2016). These channels

are generally isolated with poor connectivity, and range from locally sinuous to wide and braided (Mangold et al., 2012a). (2) Relatively older early Amazonian- and Hesperian-aged closed-basin-lakes (CBLs) (Cabrol and Grin, 1999), which are craters that exhibit inlet channels superposed on the rim-crest (Fig. 1A). These inlet channels typically exhibit an amphitheater-shaped headwall with one short (~ 10 km long) channel that drains into the crater interior (Fig. 1A) (Goudge et al., 2015). (3) The ancient and highly degraded Late Noachian highland craters, which exhibit fluvial features superposing highly subdued rims (Fig. 1B) (Masursky et al., 1977; Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Mangold et al., 2012b; Grant et al., 2015; Hobbs et al., 2016). With the exception of degraded craters in close proximity to, or in contact with valley networks (e.g., Fassett and Head, 2008; Fig. 3 in Hoke and Hynek, 2009), the fluvial features associated with the degraded Late Noachian highland craters (including some CBLs with longer, branching tributaries (Fig. 1B) (Goudge et al., 2015) appear to drain into the crater interior (Mangold et al., 2012b).

These different fluvial features have been variously explained through several processes. For example, the relatively younger Amazonian- and Hesperian-aged fluvial features associated with impact crater ejecta have generally been attributed to contact melting, wherein hot ejecta deposited on surface or near-surface icy deposits generates melting (Morgan and Head, 2009; Mangold, 2012; Mangold et al., 2012a; Schon and Head, 2012). The fluvial features present on the rims of Amazonian- and Hesperian-aged CBLs remain of uncertain origin, but have been proposed to form from overland flow generated by regional flooding events (Goudge et al., 2015). The older Late Noachian

fluvial features associated with impact craters have typically been attributed to rainfall and fluvial erosion in a warmer and wetter early martian climate (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004). Noting that a warm early martian climate (and sustained rainfall) is not predicted by recent 3D global climate models (Forget et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2015), these fluvial features have been alternatively explained by snowmelt and fluvial erosion (Hobbs et al., 2016), from snow deposition on hot ejecta (Kite et al., 2011), or top-down melting during peak seasonal or daytime temperatures (Head and Marchant, 2014) in a cold and icy Late Noachian climate. Recent work (Weiss and Head, 2015) has proposed that impact-induced contact and basal melting of surface ice are candidate processes which may contribute to some of the ancient impact-related fluvial features in such a cold and icy ancient martian climate.

Here we investigate the quantitative characteristics of impact ejecta-induced melting of surface ice deposits. We examine the mechanisms of contact melting (Mangold, 2012) and basal melting (Weiss and Head, 2015) of surface ice deposits in order to assess their roles in the formation of impact-associated fluvial channels. In the contact melting scenario (Fig. 2), ejecta is at elevated temperatures due to a combination of pre-impact geothermal heating at the depth from which it is excavated, and shock heating during the impact. When the ejecta is emplaced on the surface snow and ice deposits (Fig. 2B), the hot ejecta radiates heat outwards and conducts heat downwards into the icy deposits, thereby generating meltwater. In the basal melting scenario (Fig. 2), ejecta deposition on top of regional surface snow and ice deposits inhibits geothermal heat diffusion through the ice. As a result, following the impact event the 273 K ice melting isotherm within the

shallow crust is predicted to rise to the base of the ice sheet given sufficient ejecta thicknesses (Fig. 2D, E). This causes the ice sheet to melt from the bottom-up, supplying a potential source of liquid water for fluvial erosion proximal to the impact crater. These mechanisms appear attractive within the constraints of the current 3D climate models for the Late Noachian (e.g., Wordsworth et al., 2013) because they do not require warm atmospheric temperatures (e.g., the rainfall hypothesis for Late Noachian craters (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002). For example, contact and basal melting could operate as a background landscape/crater degradation processes in a cold and icy early Mars (Weiss and Head, 2015) even in the absence of punctuated warming events (e.g., Halevy and Head, 2014; Wordsworth et al., 2015).

The source of water (surface snow and ice) in the contact and basal melting hypotheses has been shown to be readily available throughout martian history. This includes, for example:

(1) The Amazonian-aged ~10 m thick latitude-dependent mantle (LDM) (Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003), up to ~1 km thick lobate debris aprons (LDAs) (e.g., Peirce and Crown, 2003; Chuang and Crown, 2005; Head et al., 2006a; Plaut et al., 2009; Baker et al., 2010; Fastook et al., 2014), lineated valley fill (LVF) (Head et al., 2006b; Holt et al., 2008; Kress and Head, 2008; Morgan et al., 2009; Baker et al., 2010), concentric crater fill (CCF) (Levy et al., 2010; Dickson et al., 2010; Fastook and Head, 2014), and other buried ice deposits (e.g., Viola et al., 2015; Bramson et al., 2015) interpreted to be debris-covered glacial deposits that are remnants of regional ice sheets (Fastook et al., 2014; Fastook and Head, 2014) formed in the mid-high

latitudes during periods of higher martian obliquity (Madeleine et al., 2009).

(2) Amazonian-aged pedestal craters (Barlow, 2006; Wrobel et al., 2006; Kadish et al., 2008, 2010) and double-layered ejecta (DLE) craters (Weiss and Head, 2013; 2014) hypothesized to form in ~20-200 m thick regional surface ice sheets in the mid-high latitudes during periods of higher martian obliquity.

(3) Tropical mountain glacier deposits (Head and Marchant, 2003; Shean et al., 2005; Kadish et al., 2014; Head and Weiss, 2014; Scanlon et al., 2015).

(4) Evidence for a Late-Noachian-Early Hesperian-aged expanded south-polar cap (Kargel and Strom, 1992; Head and Pratt, 2001; Ghatan and Head, 2002; Fastook et al., 2012; Kress and Head, 2014; Scanlon and Head, 2014; Scanlon et al., 2016).

(5) The potential presence of hectometers-thick regional surface snow and ice deposits in the southern highlands during the Late Noachian period (Head and Marchant, 2014; Fastook and Head, 2015), proposed on the basis of recent 3D global climate models (Forget et al., 2013; Wordsworth et al., 2013, 2015).

Although sources of surface snow and ice in the contact and basal melting scenarios are not lacking, the degree to which both contact and basal melting of surface ice may contribute to fluvial erosion is not yet clear from a physical standpoint. Previous work which assessed contact melting of near-surface icy deposits by conduction from hot ejecta (Mangold, 2012; Mangold et al., 2012a) did so under a wide range of ejecta temperatures. Here, we attempt to provide more precise estimates for ejecta temperature based on established shock physics principles. In this contribution, we provide a quantitative treatment of both the contact and basal melting mechanisms in order to assess whether impact ejecta-induced melting of surface ice deposits could have played a

role in forming impact crater-associated fluvial channels during the history of Mars.

2. Heat flow modeling

Could an impact event into any of the various martian surface ice deposits discussed above generate melting and contribute to the observed fluvial erosion around the rim and ejecta of some martian impact craters? At what crater diameters might this process occur? What surface temperatures, geothermal heat fluxes, and ice thicknesses are required to generate basal melting? How long after impact would contact and basal melting of the surrounding surface ice occur, and over what period of time would it continue? What volume of melt is expected, and what are the predicted melting rates? During which martian periods could this process have operated?

In order to address these questions, we implement thermal models to test whether the presence of ejecta on top of surface ice can produce substantial contact melting at the ice sheet surface, or raise the geotherm sufficiently to induce melting at the base of an ice sheet (Fig. 2).

2.1. Contact melting model

To determine the amount of heat transferred from the ejecta into the underlying ice in the contact melting scenario, we solve the one-dimensional heat conduction equation

$(\frac{\partial T}{\partial t} = \frac{k \partial^2 T}{\partial z^2})$ following Wilson and Head (2007) and Cassanelli and Head (2016). We consider heat transfer only in the vertical direction. We hold the temperature at the top of the ejecta constant at a given surface temperature (T_s), and the temperature at the base of the ejecta (the ejecta-ice interface) (T_B) at the melting point of ice (273 K). Under these

boundary conditions, solving the heat equation over the thickness of the ejecta (E_T) yields the following analytic solution:

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

$$A_j = T_E - T_S + \left(T_S + \frac{(T_B - T_S)(1 - z)}{L} \right) \sin(j\pi z/E_T) \quad (2)$$

where $T(z, t)$ is the temperature (in K) at depth (z) within the ejecta at time (t), k is the thermal diffusivity, A_j is the Fourier coefficient, which sets the initial temperature distribution, and T_E is the initial ejecta temperature. We solve eq. 1 for $n=20$, which ensured solution convergence at small values of time.

In order to determine the thickness of ice melted through time (the melting rate), we must first determine the heat flux from the ejecta into the underlying ice, which is:

$$Q_E = \frac{\kappa dT}{dz}, \text{ where } \kappa \text{ is the thermal conductivity of ejecta, and } \frac{dT}{dz} \text{ at the ejecta-ice interface}$$

is found by extrapolating the temperature gradient from $0.98z$ to $0.999z$. The melting rate (in m/s of ice melted per m^2) is thus: $R_c = Q_E / (\rho_i L + \rho_i C_{Pi} \Delta T)$, where ρ_i is the density of the ice (917 kg/m^3), L is the latent heat of fusion of ice ($3.34 \times 10^5 \text{ J kg}^{-1}$), ΔT is the temperature difference between T_B and the ambient ice temperature ($T_{(z)}$ from eq. 3), and C_{Pi} is the specific heat capacity of ice; we use the temperature-dependent relationship for C_{Pi} from Giauque and Stout (1936). We assume that no heat energy is lost towards warming the meltwater because it is expected to drain out of the ejecta. The thermal diffusivity is found as: $k = \frac{\kappa}{\rho_e C_P}$, where ρ_e is the density of the ejecta (2500 kg/m^3 ; in the range of lunar impact breccias; Kiefer et al., 2015), and C_P is the specific heat capacity of

the ejecta (800 J/kg K; in the range of lunar impact breccias and basalts; Hemingway et al., 1973)

2.2. Ejecta thermal conductivity

Our models differ from previous work on ejecta thermal profiles (Mangold, 2012; Mangold et al., 2012a) which used an ejecta κ similar to the uncomminuted target rock (~ 1.7 - 2.1 W/m K). We instead adopt an ejecta thermal conductivity value of an impact breccia, since impact ejecta is expected to be largely composed of comminuted material. Previous investigators have found that the κ of lunar breccia is ~ 0.3 W/m K over a wide range of temperatures and atmospheric pressures (Horai and Winkler, 1980; Warren and Rasmussen, 1987; Warren, 2011), and so we use this value for dry martian ejecta.

The ejecta of martian craters is expected to host pore-ice from a global ice-cemented cryosphere (e.g., Clifford, 1993; Clifford et al., 2010), and so we calculate κ of the ejecta facies as a linear mixture of the volume of rocky ejecta and pore-ice, and ejected surface ice. In our models, the excavated rocky ejecta hosts 15% pore ice by volume within an ice-cemented cryosphere, the thickness of which is the depth of the 273 K isotherm determined from eq. 3. We use a temperature-dependent κ of the pore-ice from eq. 4 where $T_E < 273$ K. When $T_E > 273$ K, the κ of water was adopted for the pore H₂O as a constant 0.569 W/m K (the temperature dependence of water κ over the temperatures in our analysis is minimal) (Ramires et al., 1995). The ejecta κ is calculated from this mixture of breccia and pore ice by approximating the volume of the excavation cavity as a paraboloid with a diameter equal to that of the transient cavity,

$$D_T = D_{SC}^{0.15 \pm 0.04} D^{0.85 \pm 0.04} \text{ (Croft, 1985), and a depth equal to } 0.1 D_T \text{ (Croft, 1980;}$$

Melosh, 1989), where D_{SC} , the average martian simple-complex crater transition diameter is 6 km (Robbins and Hynek, 2012). Although the ejecta is predicted to mix with the target material during ballistic sedimentation (Oberbeck, 1975; Senft and Stewart, 2008), it remains quantitatively unclear how this process will affect the thermal conductivity of the ejecta, and so we ignore this process in determining the ejecta κ .

2.3. Ejecta temperature

In order to solve eq. 1 under a realistic range of temperatures, we first evaluate the average temperature within the ejecta as a function of crater diameter. Previous work has assessed contact melting of near-surface icy deposits by conduction from hot ejecta under a range of ejecta temperatures (Mangold, 2012; Mangold et al., 2012a) from 473 K to 1073 K. This temperature range was applied on the basis of the high temperatures and shock pressures within the outer suevite deposit (up to ~ 1020 K, up to 80 GPa) that superposes the primary Bunte Breccia ejecta facies of the ~ 26 km diameter terrestrial Ries crater (Newsom et al., 1986; Engelhardt, 1995). The outer suevite deposit (typically ~ 5 -25 m in thickness; Stöffler et al., 2013), however, is not analogous to the primary ejecta facies (e.g., Stöffler et al., 2013; Artemieva et al., 2013). The primary ejecta facies (the Bunte Breccia) is instead interpreted to have exhibited ambient temperatures upon emplacement (Stöffler et al., 2013). We apply a different approach, and calculate ejecta temperature based on thermal parameters of the target and shock physics principles. The average ejecta temperature, T_E , is evaluated as the sum of: (1) the pre-impact temperature of the target material within the excavation cavity; and (2) the post-shock temperatures of the ejecta following shock heating by the impact event. We note that ejecta temperature

should be generally independent of radial distance from the crater because excavation streamtubes cross shock pressure isobars (e.g., Melosh, 1989), which results in ejecta of a variety of shock pressures present at all radial distances; this assumption is supported by more recent shock physics modeling (Collins et al., 2008). We find the pre-impact temperature of the ejecta within the excavation zone using the one-dimensional steady state heat equation:

$$T_{(z)} = T_{(z-1)} + \frac{Q\Delta z}{\kappa_{(z)}} \quad (3)$$

where $T_s = T_{(z=0)}$ and Q is the geothermal heat flux (in W/m^2). Previous work has shown that over the range of temperatures involved in this analysis, the thermal conductivity of basalt given by this relationship is generally equivalent to that of water ice (Clifford, 1993; Clifford et al., 2010):

$$\kappa = \frac{488.19}{T} + 0.4685 \quad (4)$$

Therefore, a basaltic mega-regolith saturated with pore ice is also predicted to share this thermal conductivity (Clifford, 1993; Clifford et al., 2010), and so we use this relationship for the thermal conductivity of the target. We also use this relationship for the thermal conductivity of the surface snow/ice deposits, which is representative of the fully densified portion of an ice sheet (Cassanelli and Head, 2015). The geometry of the excavated zone of the crater is estimated using the Maxwell Z model (Maxwell, 1977; Croft, 1980). The temperature distribution within the excavation cavity of a 50 km

diameter crater is shown in Fig. 3A ($T_s=215$ K, $Q=60$ mW/m²). In this example, the pre-impact material within the excavation zone of the crater is predicted to have a volumetric average temperature of 249 K.

We then find the post-shock change in ejecta temperature (ΔT_E) based on experimentally determined equations of state for mafic rocks (Stöffler, 1982; Trunin et al., 2001) following Artemieva and Ivanov (2004) and Fritz et al. (2005), where post-shock temperature is related to the residual energy after the shock event. The post-shock temperature of ejecta $\Delta T_{shock} = (E_H - E_R)/C_{pt}$, where E_H is the Hugoniot total energy ($E_H = \frac{U^2}{2}$; U is particle velocity), E_R is the energy released during decompression ($E_R = - \int P(V)dV$), and C_{pt} is the specific heat capacity of the target (1000 J/kg K) (Artemieva and Ivanov, 2004). Because shock pressure $\propto U$, peak shock pressure can then be related to post-shock temperature (e.g., Fig. 7 in Fritz et al., 2005). We find the peak shock pressures using the planar impact approximation (Melosh, 1989, p. 54; Melosh, 2012) and the semi-analytic Gamma model (Croft, 1982; Pierazzo et al., 1997) for a chondritic impact into basalt, where the impact velocity (10 km/s) is multiplied by a factor of $\sin(45^\circ)$ to better represent the shock pressure reduction expected from an oblique impact. Impactor diameter is found from π scaling (Holsapple and Schmidt, 1982). For simplicity, we do not account for the near-surface shock pressure reduction within the near-surface interference zone. Using an interference zone geometry from Melosh (1984) of constant 1 GPa shows that neglecting interference will overestimate our average ejecta temperature estimates by only a small amount: $<2^\circ\text{C}$ for $D=10$ km, $<7^\circ\text{C}$ for $D=50$ km, $<13^\circ\text{C}$ for $D=100$ km (D is crater diameter). The peak shock pressures within the excavated zone using this method are shown in Fig. 3A for a 50 km diameter

crater. We then convert these shock pressures into post-shock temperature following the methods of Fritz et al. (2005), and add these values to the pre-impact temperature from Fig. 3A (left panel) to find the post-impact ejecta temperature provenance within the excavation cavity (Fig. 3B). In this example, the material within the excavation zone of the crater is predicted to have a volumetric average ejecta temperature (T_E) of 309 K. Using these models, we find the volumetric average T_E over a range of crater diameters. We neglect cooling of the ejecta during its ballistic trajectory. Our model results showing the post-emplacement volumetric average ejecta temperature (T_E) (the sum of pre-impact temperature and post-shock temperature) are presented in Fig. 4 for a range of crater diameters, surface temperatures, and geothermal heat fluxes.

2.4. Ejecta thickness

In order to find the ejecta thickness (E_T), we use a simplified ejecta thickness function for martian craters (Garvin and Frawley, 1998; Garvin et al., 2000). We approximate the ejecta thickness as a function of radial distance (r) from the rim-crest as the total topography of the ejecta minus the structural uplift height function from Stewart and Valiant (2006), such that

$$E_T = h \left(\frac{r}{R} \right)^b - hC \left(\frac{r}{R} \right)^n \quad (5)$$

where R is crater radius and h is rim-crest height. Rim-crest heights and ejecta topography display exceptional variability (Barnouin-Jha et al., 2005), and so we examine both the global average fresh rim-height least-squares fit function ($h_{avg} =$

$0.025D^{0.820}$, in km) (Robbins and Hynek, 2012) and a maximum ($h_{max} = 0.13D^{0.573}$) (Craddock et al., 1997), both of which are valid for rim heights post-crater collapse. For the fraction of the rim-crest composed of the structural uplifted target (denoted by C), we use a value of 0.5 (e.g., Stewart and Valiant, 2006; Mouginis-Mark and Boyce, 2012; Mouginis-Mark, 2015) (the C value of ~ 0.2 for lunar craters, Sharpton, 2014, has not been demonstrated for Mars), and for the structural uplift decay constant, we use an n value of -5.5 (Stewart and Valiant, 2006; Black and Stewart, 2008). Garvin et al. (2000) found that the average ejecta decay constant, b , is -3.73 for polar craters, and -2.30 for nonpolar craters; we adopt an intermediate b value of -3. To assess our models under the wide range of observed crater rim-heights, we present the average rim-height (h_{avg}) models as generally representative of typical craters, and the maximum rim-height (h_{max}) models as an upper-limit, but not necessarily rare configuration (e.g., Craddock et al., 1997, Stewart and Valiant, 2006).

2.5. Basal melting model

In order to evaluate the basal melting mechanism, we find basal ice temperatures from the 1D steady-state heat equation (eq. 3) using the parameters and conditions outlined in section 2.1. Our preliminary analysis shows that this approach is appropriate because the cooling timescales of the warm ejecta are substantially lower than the time required for the 273 K isotherm to melt through the cryosphere and rise to the base of the ice sheet. We calculate total melting timescales (where melting begins at the base of the ice-cemented cryosphere and evolves upwards) by solving for the melt rate (in m/s of ice melted per m^2):

$$R_b = \left(Q - \frac{\int_0^Z \kappa(z) (T_m - T_s)}{z} \right) \frac{1}{\rho_i L} \quad (6)$$

where T_m is the ice-melting isotherm (273 K). We set the ice-cemented cryosphere thickness at the initial depth of the 273 K isotherm. The final stratigraphy used in our thermal models is shown in Fig. 2B.

We run the models under parameters illustrative of the entire course of martian geologic history: we use surface heat fluxes ranging from 20 mW/m² to 100 mW/m² (Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005; Ruiz et al., 2011; Plesa et al., 2015), ice thicknesses ranging from 10-1000 m (Kadish et al., 2010; Fastook and Head, 2014), and crater diameters up to 150 km. We implement surface temperatures of 215 K and 235 K, which are within the range of mean annual surface temperatures between the Amazonian and those predicted by Late Noachian global climate models with a thicker CO₂ atmosphere and 100% humidity (Forget et al., 2013; Wordsworth et al., 2013). We also explore a warmer Late Noachian scenario with a surface temperature of 255 K.

3. Discussion

In this section, we evaluate our contact and basal melting model results. Our models assume the presence of a continuous ice sheet, although we note that local, patchier ice deposits may also have occurred, and could also be subject to melting (albeit with lower melt volumes). The crater diameters shown represent the diameter upon impact rather than the reduced diameter following surface ice removal (Fig. 5). We first discuss the

general predictions made by the models (Section 3.1 and 3.2), and then assess the fate of the meltwater in chronological order: We first examine the fate of the meltwater produced at the top of the ice sheet by contact melting immediately following the impact (Section 3.3), and then examine the fate of the meltwater produced at the base of the cryosphere as the 273 K isotherm rises through the martian subsurface (Section 3.4). We then discuss the fate of the basal meltwater generated after the 273 K isotherm has reached the base of the ice sheet (Section 3.5). Finally, we examine the possibility for the icy surface layer to glacially flow in order to evaluate its role in changing the model geometry and geomorphological predictions (Section 3.6).

3.1. Contact melting model results

In our contact melting models we calculate (1) the volumetric average ejecta temperature for a given crater diameter (Fig. 4) (Section 2.3), (2) the thickness of ice melted as a function of radial distance from the rim crest (Fig. 6), (3) the total melt volumes produced (Fig. 7), and (4) the total contact melting duration (Fig. 7).

Our ejecta temperature model results (Fig. 4) show that average martian ejecta temperatures are substantially lower than assumed by previous work (473-1073K; Mangold, 2012). We find that the average ejecta temperatures typically range from ~220 K to ~500 K under typical martian conditions, and may range up to ~600 K for exceptionally high heat fluxes (100 mW/m^2). Our results show that the average ejecta temperature increases with increasing crater diameter due to the combination of increasing average peak shock pressures, and increasing the excavation depth of the target (thereby excavating deeper, hotter material). Increasing the surface temperature has

an almost linear effect of increasing the ejecta temperature (Fig. 4). For example, increasing the surface temperature to 235 K from 215 K increases the average ejecta temperature by a nearly constant 20 K over the entire crater diameter range. At larger crater diameters, the temperature increase is slightly greater due to the temperature-dependent κ of the subsurface (eq. 4) in concert with the increased excavation depth. Increasing the surface heat flux has a much larger effect, and can increase surface temperature by tens to hundreds of degrees (Fig. 4), an effect that is amplified at larger crater diameters due to the increased excavation depth.

Our contact melting models which show the ice thickness melted as a function radial distance from the crater rim-crest are shown in Fig. 6 for surface temperatures of 215 K (left panels; Fig. 6A, D), 235 K (middle panels; Fig. 6B, E), and 255 K (right panels; Fig. 6C, F), and for surface heat fluxes of 40 mW/m² (top panels; Fig. 6A-C) and 60 mW/m² (bottom panels; Fig. 6D-F). Here we do not place a limit on the ice sheet thickness, and thus show the maximum amount of ice that can be melted for a given crater diameter. Our models show that under martian conditions, hot ejecta is able to melt tens to hundreds of meters of ice depending on the surface temperature and heat flux. Although T_E is constant with radial distance from the rim-crest in our models, the thicker ejecta near the rim (eq. 5) is able to melt a substantially greater thickness of ice (Fig. 6). For a surface temperature of 215 K and heat flux of 40 mW/m², a 100 km diameter crater (orange line; Fig. 6A) can melt ~120 m of ice at 0.1R from the rim-crest, but only ~20 m of ice 1.2R from the rim-crest. Larger craters have hotter ejecta (Fig. 4), and so are able to melt surface ice out to greater distances relative to smaller craters (Fig. 6). Increasing the surface temperature and heat flux results in melting of modestly thicker amounts of

ice at greater distances from the rim-crest for any given crater diameter (Fig. 6).

Increasing the ejecta thickness also modestly increases the thickness of ice predicted to melt (see smaller panels for maximum-rim height case) (Fig. 6).

We also present the total volume of surface ice predicted to melt by conduction from hot ejecta in Fig. 7 for surface temperatures of 215 K (top panels; Fig. 7A, D), 235 K (middle panels; Fig. 7B, E), and 255 K (bottom panels; Fig. 7C, F). The contact melt volumes are shown for surface heat fluxes of 40, 60, and 100 mW/m². We also present contact melt volumes for a lower (Amazonian) heat flux of 20 mW/m² for $T_s=215\text{K}$ (Fig. 6A, D). The contact melting volumes increase substantially as crater diameter increases, but there is only a modest increase in melt volume as surface temperature increases. The most important variables controlling the melt volume are (1) thickness of surface ice available for melting (i.e., the supply limit; Fastook and Head, 2014); (2) the crater diameter; and (3) the surface heat flux.

In all cases, meltwater volumes are predicted to range between 10⁰ and 10⁴ km³, and melting timescales range from ~100 years to ~80 kyrs (Fig. 7). We find that contact melting meltwater fluxes range from 0.003 to 0.3 m/yr per m² ($\sim 10^{-11}$ to $\sim 10^{-9}$ m/s per m²). The large variation in meltwater volumes is largely due to crater diameter and ice thickness, and to a lesser extent, surface heat flux. At small crater diameters, the ejecta facies is colder and thinner, and so cools below 273 K before it can melt substantial volumes of ice. The thickness of the surface ice at the time of impact further acts as a supply limit for the volume of meltwater produced. For example, the melting of a 10 m thick ice sheet can produce up to $\sim 10^3$ km³ of meltwater (e.g., dashed white line in Fig. 7A), whereas the ejecta may be hot enough to melt substantially greater thicknesses of ice

(see solid white line in Fig. 7A for 100 m thick ice) and the underlying ice-cemented regolith. Geothermal heat flux also plays a large role in meltwater production because modest variations in the heat flux can generate large variations in the temperature distribution within the pre-impact target, and thus the post-impact ejecta (Fig. 3).

The major parameter that controls the wide range in melting timescales (from hundreds of years to ~80 kyrs) is ice thickness, because thicker ice sheets provide a larger volume available for melting. For example, for $T_s=215\text{K}$ and $Q=40\text{ mW/m}^2$, a 120 km diameter crater will exhibit contact melting of a 10 m thick ice sheet for hundreds of years (Fig. 6A). For the same thermal conditions, the 120 km diameter crater would exhibit contact melting of a 100 m thick ice sheet for a few tens of kyrs. Additionally, ejecta thickness affects the melting timescale. At large ice thicknesses, thicker ejecta facies increases the melting timescales because cooling is slower, and thus melting is prolonged (e.g., Fig. 7A, D). For thinner ice sheets, thicker ejecta instead reduces melting timescales because the heat flux into the surface ice is increased, and the entire thickness of the ice sheet can be fully melted (e.g., Fig. 7A, D).

In summary, contact melting of surface ice sheets overlain by impact ejecta is predicted to occur under a wide range of surface temperatures, heat fluxes, and ice thicknesses, and is thus predicted to operate on a large range ($\sim >10\text{-}30\text{ km}$ diameter) of martian impact craters that impacted into surface ice deposits on Mars. These model results are broadly consistent with our current knowledge of the size-distribution of impact craters that host fluvial channels on the ejecta (Morgan and Head, 2009; Mangold, 2012; Schon and Head, 2012) (discussed in Section 5), but now place quantitative limits on the ejecta temperatures and expected meltwater volumes. We discuss predictions for

contact melting through martian geologic history in Section 4.

3.2. Basal melting model results

In our basal melting models (Figs. 8-10), we calculate the total melt volumes produced (Fig. 8A), and because basal melting is driven entirely by the geothermal heat flux (which is low compared with the heat flux from contact melting), we also present the time-averaged melt fluxes (Fig. 8B). We calculate the maximum radial-extent of basal melting from the crater rim-crest (Fig. 8C), as well as the total basal melting durations (Fig. 11). Our basal melting models use the ice thickness outputs from the contact melting models to set the initial ice thickness as a function of radial distance that is available for basal melting.

In a manner similar to contact melting, in the case of basal melting, increasing crater diameter, ice thickness, and surface heat flux all lead to greater volumes of melt, and melting at larger radial distances. We now briefly present one illustrative example: consider a 50 km diameter crater formed during a period where the surface temperature is 235 K, the surface heat flux is 60 mW/m^2 , and a regional ice sheet is present that is 100 m thick (Fig. 9). In this scenario, a substantial amount of surface ice has already been melted through contact melting (Fig. 6E), and so the average rim-height case would not predict basal melting to occur, except for craters between 85 and 90 km in diameter (Fig. 9A). In the maximum rim-height models, however (Fig. 9D), the thermal insulation from the ejecta is sufficient to generate $\sim 45 \text{ km}^3$ of meltwater (black star in Fig. 9D). The average melt flux in this case is predicted to be $\sim 10^{-3} \text{ m}^3/\text{yr}$ (black star in Fig. 9E), and basal melting is predicted to extend out to $0.2R$ from the rim-crest (black star in Fig. 9F).

Basal melting of the surface ice begins ~500 kyr following the impact (black star in Fig. 11K), and continues for ~80 kyr (black star in Fig. 11E).

The models show that the most important factor in impact ejecta-related basal melting is how much surface ice remains following contact melting. Consequently, the initial ice thickness plays a large role in determining whether or not a particular crater will exhibit basal melting; for example, in Fig. 9D, craters above ~100 km in diameter are not predicted to exhibit basal melting for an ice thickness of 100 m ($Q=100 \text{ mW/m}^2$; dashed red line in Fig. 9D) because there is no ice left to be basally melted. For an ice thickness of 1000 m, however, basal melting is predicted to occur for all craters $\geq 5 \text{ km}$ in diameter (solid red line in Fig. 9D) because contact melting is not able to melt the entire 1 km thick surface ice.

The geothermal heat flux also plays a major role in the melt volume, and it directly controls the melt fluxes and timescales. For example, Fig. 9A and B show that an impact crater forming on a 1000 m surface ice layer (for the case of an average rim-height and $T_s=235 \text{ K}$) generates several orders of magnitude greater melt volumes and melt fluxes for the 100 mW/m^2 heat flux compared with the 40 or 60 mW/m^2 heat fluxes. Figure 11H shows how the 100 mW/m^2 heat flux causes basal melting to begin in ~100 kyrs, compared with ~400 kyrs for the 60 mW/m^2 heat flux and ~1 Myr for the 40 mW/m^2 heat flux. The geothermal heat flux also controls the basal melting duration: Figure 11E shows how the 100 mW/m^2 heat flux causes basal melting to continue for ~200 kyrs for a 1000 m thick ice sheet, compared with ~300 kyrs for the 60 mW/m^2 heat flux and ~600 kyrs for the 40 mW/m^2 heat flux.

Thicker ejecta, higher geothermal heat fluxes, and larger crater diameters all lead to

greater melting volumes (e.g., Fig. 10A), and melting at increasing distance from the rim-crest (e.g., Fig. 10C). The timescale (following the impact) required for initiation of basal melting of the surface ice decreases with increasing ejecta thickness due to the greater insulation. Consequently, basal melting is predicted to initiate closest to the rim-crest, and continue progressively further from the rim-crest with time.

We now discuss the chronology of ice melting and the predictions for the hydrologic fate of the meltwater.

3.3. Contact meltwater

Contact melting is predicted to initiate immediately following the deposition of hot ejecta, and may last for hundreds to tens of thousands of years (Fig. 7). Ice melting begins at the interface between the overlying ejecta and the surface ice. Consequently, the meltwater will follow the local topographic gradient of the surface ice, which, in the absence of complex surface topography, will be controlled by the outward-facing slope of the rim structural uplift (Fig. 12A). Thus, once generated, the meltwater is predicted to flow downhill, away from the crater, on the top of the ice layer, and could contribute to erosion of the ejecta through sapping (Fig. 12). Exceptions to this are for craters forming in $\sim 10^2$ m thick ice deposits (as thick as the structural uplift; Fig. 12C), where melting of the near-rim ice is predicted to reverse the topographic slope and lead to contact meltwater flowing into the crater (Fig. 12D). Importantly, fluvial channels interpreted to result from contact melting are largely found at the ejecta surface (Morgan and Head, 2009; Mangold, 2012; Schon and Head, 2012). As discussed in Mangold (2012), if the meltwater is over-pressurized, it may be able to breach the ejecta vertically to generate

bursts of fluvial activity. Alternatively, fluvial channels at the surface of the ejecta could form in an analogous mechanism to terrestrial springs. In this scenario, meltwater generated at the ice surface close to the rim will flow downslope. If the meltwater meets an impermeable layer, it will flow along this layer laterally until it breaches the ejecta surface (blue arrows in Fig. 12A) and can incise the ejecta, ultimately forming fluvial channels.

3.4. Cryospheric meltwater

Following chronologically from contact melting is cryosphere melting. We predict that the 273 K isotherm will rise through the cryosphere and melt the subsurface pore-ice due to the thermally insulating effects of the ejecta. Indeed, in order for basal melting of the surface ice to occur, the 273 K ice-melting isotherm must first rise through the ice-cemented cryosphere to the base of the ice sheet (Fig. 2D, E). Consequently, any impact crater predicted to exhibit basal melting of an ice sheet is also predicted to have melted through the entire thickness of the ice-cemented cryosphere (Fig. 2E). However, the inverse is not necessarily always true, that is, a crater can experience melting of the ice-cemented cryosphere without associated basal melting of an ice sheet. This can occur in cases where contact melting of ejecta has already melted the emplaced surface ice sheet, or in cases where the 273 K isotherm is not raised all the way to the surface. Figure 11G-L shows the crater diameter and thermal conditions under which the entire cryosphere is predicted to be melted through.

The meltwater generated by bottom-up melting of the pore-ice within the cryosphere (due to the thermally insulating ejecta) is predicted to drain downwards through the dry

permeable subsurface (Fig. 2E), and thus offers a source of water for groundwater recharge below impact craters. This is analogous to the “heat-pipe drain pipe” groundwater recharge mechanism produced by local volcano thermal effects on the cryosphere (Cassanelli et al., 2015). Furthermore, as the pore-ice within the cryosphere melts and drains downwards, the combination of percolating water and high subsurface temperatures in the subsurface could cause rock-water interactions that might plausibly chemically alter crustal rocks into clays (e.g., Tosca and Knoll, 2009).

Clay minerals requiring water-rock interactions have been detected near-globally on the surface of Mars (Poulet et al., 2005; Bibring et al., 2006; Mustard et al., 2008; Murchie et al., 2009; Ehlmann et al., 2009; Loizeau et al., 2012; Quantin et al., 2012; Carter et al., 2013, 2015; Sun and Milliken, 2015). Their identification primarily within crater central peaks, walls, and ejecta has led to the interpretation that these clay minerals were excavated from depth. These clays minerals are variously interpreted to reflect weathering at the surface and subsequent burial (Loizeau et al., 2012; Poulet et al., 2005; Carter et al., 2015), or formation in the subsurface by interaction with crustal fluids (Ehlmann et al., 2011; Loizeau et al., 2012; Carter et al., 2013; Sun and Milliken, 2015). Indeed, the observation of increasing chloritization with depth (between ~1-7 km) requires the presence of fluids at these depths within the martian crust (Sun and Milliken, 2015). Craters which excavate these clays are primarily Noachian and Hesperian in age, although some Amazonian examples are also present (Sun and Milliken, 2015), raising the possibility of younger, localized regions of subsurface clay formation. The aqueous environment which formed the clays has previously been proposed to relate to impact-induced hydrothermal activity (i.e., Abramov and Kring, 2005; Schwenzer and Kring,

2009, 2013; Schwenzer et al., 2012) or burial diagenesis/metamorphism in the subsurface (e.g., Ehlmann et al., 2011; Loizeau et al., 2012; Carter et al., 2013; Sun and Milliken, 2015), or the remnant of a buried ancient Noachian-aged surface which experienced warmer and wetter conditions (e.g., Loizeau et al., 2012; Poulet et al., 2005; Bibring et al., 2006; Carter et al., 2015).

Due to the widespread distribution of martian impact craters throughout both space and time, we suggest that the impact-induced basal melting scenario (Fig. 2E) provides another candidate mechanism for generating the subsurface liquid water needed to form clays in the martian subsurface. Full melt-through of the cryosphere through basal melting is predicted to have occurred at progressively smaller crater diameters at higher heat fluxes (and is thus predicted to be more common during the Noachian and Hesperian periods) (see Fig. 11G), but partial melting of the cryosphere is predicted in the Amazonian as well. This is generally consistent with the inferred ages associated with clay formation (Sun and Milliken, 2015). Importantly, surface ice is not required to be present on the pre-impact target in this scenario. Given sufficiently large impactors and the presence of a global ice-cemented cryosphere (Clifford, 1993; Clifford et al., 2010), this process is predicted to be virtually ubiquitous throughout the subsurface of Mars. After an impact event, the insulating effects of the ejecta are predicted to elevate the geotherm. As the 273 K ice melting isotherm rises through the ice-cemented cryosphere, pore-ice will melt and percolate downwards through the crust. This mechanism could plausibly contribute to subsurface clay formation, and warrants further quantitative investigation to determine if the subsurface temperatures, volumes of water, and timescales of water-rock interaction (e.g., Tosca and Knoll, 2009) predicted by this

mechanism are consistent with the observed clay maturity and mineralogy.

3.5. Basal meltwater

After the 273 K isotherm rises through the cryosphere and reaches the base of the ice sheet, basal melting of the ice is predicted to initiate. What is the fate of the meltwater generated at the base of the ice sheet? In which direction will the basal meltwater be transported, and is it predicted to produce surface runoff, or infiltrate and drain through the substrate?

The infiltration capacity (I in m/s) of the target material (i.e., the rocky substrate underlying the surface ice) can be calculated as (Hendriks, 2010):

$$I = K \left(\frac{x+S+h}{x} \right) \quad (7)$$

where x is the thickness of the porous medium, S is the wetting front soil suction head, h is the head of infiltrating ponded water (all in m), and K is the hydraulic conductivity (in m/s). In order to estimate a minimum infiltration rate, we conservatively assume that S and h are equal to zero (e.g., Cassanelli and Head, 2016). Consequently, S and h in eq. 7 reduce to zero and the infiltration rate in the basal melting scenario can simply be approximated by the hydraulic conductivity:

$$K = \frac{k\rho_w g}{\mu} \quad (8)$$

where k is the intrinsic permeability (in m^2), ρ_w is the density of water (1000 kg/m^3), g is

gravity (3.71 m/s^2), and μ is the dynamic viscosity of water ($1.793 \times 10^{-3} \text{ Pa}\cdot\text{s}$ at 273 K). In our calculations, basal melt fluxes range from 3×10^{-5} to $0.009 \text{ m/yr per m}^2$ ($\sim 10^{-12}$ to $\sim 10^{-10} \text{ m/s per m}^2$) (Fig. 13A). On the basis of these melt fluxes, substrate permeability (k from eq. 8) is required to be $\leq \sim 10^{-18}$ to 10^{-16} m^2 (i.e., crystalline or consolidated sedimentary bedrock; Fig. 13A) for melting rates to exceed infiltration rates. If the melt fluxes exceed substrate infiltration rates, the melt may be transported laterally and lead to runoff (and thus potentially erosion). Higher permeabilities (such as expected for regolith) enhance infiltration rates, and are predicted instead to allow for the meltwater to infiltrate and drain through the substratum (Fig. 13A).

In order to assess if the melt can be transported up the rim structural uplift, we calculate the hydraulic head of the basal melt (H_{melt}) generated by the overlying ejecta:

$$H_{melt} = \frac{z_i \rho_i + z_e \rho_e}{\rho_w} \quad (9)$$

where z_i and ρ_i are the ice thickness and density, and z_e is the ejecta thickness. In order to examine the hydraulic head predicted independent of surface ice thickness, we do not include any surface ice ($z_i=0$). Including surface ice would increase the hydraulic head, and so these estimates should be considered lower bounds. Figure 13B shows the predicted hydraulic head as a function of crater diameter and distance from the rim-crest. If the rim structural uplift height (black line in Fig. 13B) exceeds the hydraulic head, basal meltwater will stall below the surface ice and ultimately drain downwards into the substratum. Conversely, if the hydraulic head exceeds the rim structural uplift height, then the meltwater will be sufficiently pressurized to flow up the rim structural uplift and

drain into the crater interior (Fig. 13B). Assuming a continuous ice-sheet underlying the ejecta, which acts as an impermeable barrier to vertical ascent of meltwater (i.e., it prevents meltwater from diffusing into the overlying ejecta), Fig. 13B shows that all melt within $\sim 0.6R$ from the rim-crest is predicted to be transported up the rim structural uplift slope towards the crater interior due to the overburden pressure of the overlying ejecta. The pressurization and transport of meltwater up the slopes of the crater rim structural uplift is made possible by the decreasing ejecta thickness with radial distance from the crater rim-crest. The decreased thermal insulation provided by relatively thinner distal ejecta prevents the 273 K melting isotherm from rising to the base of the ice sheet (Fig. 2F). Consequently, the ice sheet is predicted to be frozen to the bed far from the rim-crest (generally ~ 0.1 - $2.5R$ from the rim-crest), which prevents meltwater drainage away from the crater. Exceptions to this may exist in the case of patchy rim-ice deposits, where the discontinuous extent of surface ice allows meltwater drain away from the crater through the porous ejecta. Meltwater ascent up the rim structural uplift is only possible as long as the substrate permeability is low enough such that melt rates exceed infiltration rates (Fig. 13A). If the rim structural uplift is highly fractured by the impact (which increases infiltration rates), or in the case where contact melting has melted the entire thickness of near-rim ice, the meltwater may instead infiltrate downwards through the substratum. Thus, thicker ice layers, which are less likely to be fully melted (Fig. 6) favor transport of melt up the rim and into the crater interior, whereas thin ice layers generally prevent melt transport into the crater interior. If a hydrothermal system (Abramov and Kring, 2005; Schwenzer and Kring, 2009) is still present within the crater (predicted to last tens to hundreds of kyrs; cf. Fig. 11G-L) and the cryosphere directly adjacent to the crater (Fig.

2E) is melted during basal melting, the meltwater may flow laterally into the crater interior (e.g., Schwenzer and Kring, 2009) independent of ice thickness.

Is the meltwater expected to form subglacial fluvial channels on the outer slopes of the structurally uplifted rim (the rocky rim below the surface ice; Fig. 2F)? An analogy may be drawn with other martian fluvial features interpreted to result from basal melting of surface ice deposits. For example, the relatively young valley networks present on several martian shield volcanoes (e.g., Hecates Tholus, Ceraunius Tholus, Alba Patera) are interpreted to result from basal melting of snowpack (Fassett and Head, 2006, 2007) during periods of high surface heat flux associated with magmatic intrusions. Fassett and Head (2006, 2007) show that these valley networks are present primarily on steep slopes, downslope of the predicted location of the snowpack. They thus concluded that surface slope (i.e., flow velocity) is a major controlling factor on the formation of basal melt-generated fluvial channels (Fassett and Head, 2006, 2007). We note that in the impact crater-induced basal melting scenario explored here, the velocity of the meltwater which flows up the rim structural uplift will be controlled solely by the rate of meltwater production. Because our models predict extremely low melt-fluxes ($\sim 10^{-4}$ to $\sim 10^{-3}$ m/yr per m^2), the velocity of the meltwater as it flows up the rim structural uplift is predicted to be comparably low; consequently, the meltwater is generally not predicted to incise subglacial fluvial channels into the rocky substrate during flow up the rim structural uplift. While the presence of fluvial channels that flow up topography would provide strong evidence for a subglacial origin, these channels are unlikely to be observed due to their formation beneath the ejecta. Fluvial incision is predicted instead to initiate when the meltwater begins to flow down the slopes of the interior crater walls (Fig. 2F). This

prediction may not hold in the case of impacts into relatively thick surface ice, where the rim structural-uplift may be partially accommodated by the surface ice, leading to relatively lower (or entirely absent) bedrock rim-slopes (Weiss and Head, 2015) that are more likely to form fluvial channels. In this case, the removal of a thick surface ice layer at a later time would serve to reduce the currently observed crater diameter (Fig. 5) (Weiss and Head, 2015) such that channels currently observed at large radial distances from the rim may in fact have been located proximal to the rim-crest.

3.6. *Ice flow*

Could the glacial flow of the surface ice layer modify the model geometry and morphological predictions? It is important to consider the effects of ice flow because the weight of the overlying ejecta will increase the shear stress within the icy layer. In the geometry explored in this study, ice is primarily predicted to flow down the slopes of the rim structural uplift (Fig. 2D); however, because ice flow follows pressure gradients, a small amount of ice may also extrude into the crater (Fig. 2D) We assess the speed at which the icy layer beneath the ejecta will flow, and examine how ice flow is predicted to effect the contact or basal melting process. We adopt a steady-state parallel-sided flow model (e.g., Paterson, 1981, p. 86) modified to include the effects of the overlying debris weight (Konrad and Humphrey, 2000) in order to explore the range of speeds the ice could flow downslope of the rim-crest. In the parallel-sided flow model, an ice sheet of infinite width rests on an inclined plane of constant slope (θ). The ice sheet has a thickness (z_i) and density (ρ_i) and is superposed by a debris layer of thickness (z_e) and density (ρ_e). In this case, the steady-state depth-averaged horizontal velocity of the ice

sheet (Konrad and Humphrey, 2000) is:

$$U_{xavg} = \frac{2A(\rho g \sin \theta)^n}{(n+1)(n+2)z_i} \left[z_i(n+2) \left(z_i + \frac{\rho_e}{\rho_i} z_e \right)^{n+1} + \left(\frac{\rho_e}{\rho_i} z_e \right)^{n+2} - \left(z_i + \frac{\rho_e}{\rho_i} z_e \right)^{n+2} \right] \quad (10)$$

where the flow law constant (n) is 3, and the constant rate factor, A , (in $\text{s}^{-1} \text{Pa}^{-1}$) varies as a function of temperature. Contact melting (and ejecta cooling) durations are predicted to be on the order of hundreds of years to a few tens of kyrs (Fig. 7) which is typically much lower than our initial glacial flow model timescales, and so we calculate the icy layer flow speeds following contact melting, and apply ambient temperature conditions for the ejecta temperature. We adopt the temperature-dependence of A values from Paterson (1981, p. 39), so that the depth-averaged A values within the ice sheet (using the temperature profile from eq. 3) ranges between 2.8×10^{-27} for $T_s=215$ K, and 1.3×10^{-24} for $T_s=255$ K. For the parameters θ (from eq. 5), Z_i , and Z_e in eq. 10, we adopt the values averaged between 0R and 0.6R from the rim crest (the area of interest for basal melting and the area of high slopes; Fig. 13B). As in the basal melting models, the ice thickness inputs are taken from the contact melting model outputs (Fig. 6). This allows us to address how ice flow could modify the initial basal melting conditions. In this case, the 273 K isotherm has not yet risen through the ice-cemented cryosphere; the ice is still frozen to the bed, and so we neglect basal sliding. Even in the case of a wet-based ice sheet during melting, we do not consider basal sliding to be a major contributor to the flow speed because the speed of the ice will be buffered by the ice downslope from the crater that remains frozen to the bed (Fig. 6F).

Figure 14 shows the ice flow speed model results for $T_s=215$ K (Fig. 14A), $T_s=235$ K

(Fig. 14B), and $T_s=255$ K (Fig. 14C). The ice layer within $0.6R$ exhibits a wide range in average flow speeds between $0.1 \mu\text{m/yr}$ and 200 m/yr depending on crater diameter, ice thickness, surface temperature, and the ejecta thickness (average vs. maximum rim-height functions). Glacial flow speed generally decreases with increasing crater diameter (Fig. 14) because larger craters melt thicker sequences of the ice layer (Fig. 6). Thicker ice layers (particularly 1000 m) are excluded from this trend because they retain a substantial thickness even after contact melting (Fig. 6). The thick blue lines in Fig. 14 show the speed required in order for the ice to flow $0.1R$ in 100 kyr and 1 Myr , which is the range in timescales required for the 273 K isotherm to rise from depth to the base of the ice sheet (Fig. 11). We find that ice flow speeds are generally well below these values (blue lines in Fig. 14), indicating that minimal ice flow ($\leq \sim 0.1R$ flow distance) is expected on the timescales of cryospheric and basal melting, except in the case of 1 km thick ice sheets (or 100 m ice sheets and $T_s=255 \text{ K}$ for the maximum rim-height case). We conclude from the results that ice flow is generally not a major process in the impact-induced ejecta melting scenario, and is not expected to modify the model geometry except in the case of thick ice sheets (on the order of $\sim 1 \text{ km}$) and high surface temperatures ($>235 \text{ K}$).

3.7. Summary of results

In summary, following the impact (Fig. 2), hot ejecta is predicted to melt tens to hundreds of meters of surface ice (Fig. 6) over the course of ~ 100 years to 80 kyr through contact melting, corresponding to 10^0 - 10^4 km^3 of meltwater (Fig. 7). Larger craters and higher surface heat fluxes generate substantially greater volumes of

meltwater. Contact melting initiates throughout the entire radial distance of ejecta after ejecta emplacement, but terminates earlier for the thinner, faster cooling distal ejecta. Meltwater is generally predicted to flow outwards (Fig. 12A), away from the crater, except in the case of ice sheets of comparable thickness to the rim structural-uplift (Fig. 12C). In this case, following substantial melting of the near rim-crest ice, meltwater may flow downwards into the crater (Fig. 12D). This process is predicted to generate fluvial channels on the ejecta by meltwater breaching the ejecta along an impermeable layer.

Following the cooling of ejecta, the overlying ejecta facies frequently provides sufficient thermal insulation to raise the 273 K melting isotherm through the ice-cemented cryosphere to the base of the ice sheet (Fig. 11). Melting of the pore-ice within the cryosphere will provide a large volume of water for groundwater recharge focused in an area in a similar manner to the “heat-pipe drain pipe” groundwater recharge mechanism of Cassanelli et al., (2015). These processes could contribute to subsurface chemical alteration to form clays in the shallow crust.

The ice-melting isotherm is predicted to reach the base of the ice sheet within ~100 kyrs to ~1 Myr after the impact (Fig. 11). If the ice sheet is thick and has not been previously removed by contact melting, this can lead to basal melting of the ice sheet that lasts for ~1 kyr to ~1 Myr. Larger crater diameters and higher geothermal heat fluxes lead to increased melt volumes ($\sim 10^0$ - 10^5 km³) and fluxes. Melting initiates closest to the rim and continues progressively outward, generally out to ~0.1 to 2.5R from the rim-crest. If the rocky substrate below the surface ice has a relatively low permeability (i.e., crystalline or consolidated sedimentary bedrock), the basal melt fluxes will exceed the infiltration rates. In this case, the pressurization from the overlying ejecta and ice may

transport the basal meltwater up the slopes of the crater rim structural uplift, whereupon the meltwater is predicted to drain into the crater interior and down the crater walls.

The ice layer is predicted to flow away from the crater, but the flow speeds are so low that this process is not likely to have a large effect on the basal melting geometry or melt volumes. With these guidelines in place, we now specifically examine the conditions under which basal melting may occur in each of the different martian geologic periods.

4. Ice melting throughout martian geologic history

In this section, we discuss the results of our thermal models in the context of the different martian geologic periods (ages from Werner and Tanaka (2011) using the Hartmann (2005) production function) and their potential surface temperature (Wordsworth et al., 2013, 2015) and heat flux ranges (from Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005; Ruiz et al., 2011; and Plesa et al., 2015) for crater diameters up to 150 km. These results are summarized in Fig. 15.

4.1. Amazonian period (0.0-3.0 Ga; $T_s=215$ K; $Q=\sim 20-40$ mW/m²)

Under Amazonian conditions, post-impact ejecta temperatures range between 220K and 360 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than ~30-40 km in diameter are predicted to exhibit contact melting (blue star in Fig. 15A), and produce melt volumes ranging from 10^0 to 10^3 km³ (pink and white lines in Fig. 7A, D).

For typical Amazonian parameters and the average rim-height case, basal melting is not predicted to occur in our models (Fig. 8A, D), except perhaps for craters larger than

150 km in diameter, which we do not explore in this study. This is due to contact melting following the impact, which removes any surface ice that would have undergone basal melting under typical Amazonian conditions (blue star in Fig. 15B).

For a surface heat flux of 40 mW/m^2 ($T_s=215 \text{ K}$) the 273 K isotherm depth is ~ 3.6 km, and the full thickness of the cryosphere is not predicted to fully melt through.

4.2. Hesperian period (3.0-3.6 Ga; $T_s=215 \text{ K}$; $Q=\sim 30\text{-}60 \text{ mW/m}^2$)

Under Hesperian conditions ($Q=40\text{-}60 \text{ mW/m}^2$), ejecta temperatures range between 220K and 430 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than ~ 30 km in diameter are predicted to exhibit contact melting (green star in Fig. 15A) and produce melt volumes ranging from 10^0 to 10^4 km^3 (Fig. 7A, D).

Basal melting is predicted to occur only for $Q \geq 60 \text{ mW/m}^2$, and only for surface ice thicknesses of ~ 1 km; smaller ice thicknesses are almost entirely removed by contact melting, and thus prevent ice from being preserved for basal melting. For a heat flux of 60 mW/m^2 , basal melting will initiate for craters between ~ 85 km in diameter (maximum rim-height) and 140 km in diameter (average rim-height) (Fig. 8A, B). Melt volumes range from 10^1 to 10^3 km^3 (Fig. 8A, B), and basal melting is predicted to extend from 0.1-0.4R from the rim-crest (Fig. 8C, F).

For a surface heat flux of 60 mW/m^2 ($T_s=215 \text{ K}$), the 273 K isotherm depth is ~ 2.4 km. Basal melting will begin at least ~ 550 kyrs after impact (Fig. 11J), and will last up to ~ 400 kyrs in order to entirely melt-through a 1000 m ice sheet (Fig. 11D).

4.3. Late Noachian period (3.6-3.8 Ga; $T_s=215\text{-}255 \text{ K}$; $Q=\sim 40\text{-}65 \text{ mW/m}^2$)

Under Late Noachian conditions ($T_s=215\text{-}255\text{ K}$, $Q=40\text{-}60\text{ mW/m}^2$), ejecta temperatures range between 220K and 490 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than $\sim 10\text{-}25\text{ km}$ in diameter are predicted to exhibit contact melting (yellow star in Fig. 15A), and produce melt volumes ranging from 10^0 to 10^4 km^3 (white and black lines in Fig. 7) for ice sheet thicknesses (Fastook and Head, 2015) between 10 and 1000 m.

Basal melting is not predicted in a cold Late Noachian period ($T_s=215\text{ K}$), except for heat fluxes $\geq \sim 60\text{ mW/m}^2$ and ice thicknesses of $\sim 1000\text{ m}$. Representative melting volumes, distances, and timescales are identical to those discussed above for the Hesperian case with a 60 mW/m^2 heat flux.

In the case of a slightly warmer Noachian period ($T_s=235\text{ K}$; e.g., Wordsworth et al., 2013) onset diameters for basal melting range from $\sim 30\text{-}130\text{ km}$ for heat fluxes between $40\text{-}60\text{ mW/m}^2$ and an ice thickness of 1000 m (yellow star in Fig. 15B; Fig. 9A, D). Melt volumes range from 10^0 to 10^4 km^3 (Fig. 9A, D), and basal melting may extend from 0.1 to 1.0R from the rim-crest (Fig. 9C, F). Basal melting is predicted for 100 m thick ice sheets only for the maximum rim-height case and 60 mW/m^2 heat flux for a restricted crater diameter range between $35\text{ and }85\text{ km}$ in diameter (Fig. 9D), which may produce melt volumes between ~ 4 and $\sim 65\text{ km}^3$. For this case, basal melting extends between 0.1 and 0.5R from the rim.

To simulate an even warmer Late Noachian period, we also examine the thermal models for a surface temperature of 255 K (Fig. 10). For the average rim-height case, onset diameters for basal melting are $\sim 35\text{-}55\text{ km}$ for a 100 m thick ice sheet, or 5 km for a 1000 m thick ice sheet (Fig. 10A). Craters larger than 125 km in diameter are not

predicted to exhibit basal melting in the case of 40 mW/m^2 with 100 m thick ice because contact melting has already removed the surface ice. Melt volumes range from 10^1 to 10^2 km^3 for a 100 m thick ice sheet, and from 10^0 to 10^5 km^3 for a 1000 m thick ice sheet (Fig. 10A). Basal melting may extend from 0.1-0.9R from the rim-crest for 100 m thick ice, and 0.4-2.5R for 1000 m thick ice (Fig. 10C).

For the maximum rim-height case, onset diameters for basal melting are $\sim 15\text{-}25 \text{ km}$ for a 100 m thick ice sheet, or 5 km for a 1000 m thick ice sheet (Fig. 10A). Craters larger than 125 km in diameter are not predicted to exhibit basal melting in the 40 mW/m^2 case with 100 m thick ice because contact melting has already removed the surface ice. Melt volumes range from 10^1 to 10^2 km^3 for a 100 m thick ice sheet, and from 10^1 to 10^5 km^3 for a 1000 m thick ice sheet (Fig. 10A). For a 10 m thick ice sheet, there is a restricted range of crater diameters between 10-15 km in diameter that exhibit basal melting, but melting volumes are negligible (0.2 km^3). Basal melting may extend from 0.2-1.2R from the rim-crest for 100 m thick ice, and 1.1-2.5R for 1000 m thick ice (Fig. 10F).

For a surface heat flux of 60 mW/m^2 ($T_s=235 \text{ K}$), the 273 K isotherm depth is $\sim 1.2 \text{ km}$. Basal melting will begin at least $\sim 325 \text{ kyrs}$ after impact (Fig. 11K), and will last $\sim 10\text{-}100 \text{ kyrs}$ for a 100 m ice sheet, and $\sim 200 \text{ kyrs}$ for a 1000 m ice sheet (Fig. 11E). For $T_s=255 \text{ K}$, basal melting will begin $\sim 150 \text{ kyrs}$ after impact (Fig. 11L), and continue for at least $\sim 5 \text{ kyrs}$ for 10 m thick ice, $\sim 30 \text{ kyrs}$ for 100 m thick ice, and $\sim 200 \text{ kyrs}$ for 1000 m thick ice (Fig. 11F).

4.4. Early-Mid Noachian period (3.8-4.5 Ga; $Q=\sim 45\text{-}100 \text{ mW/m}^2$)

Atmospheric temperatures during this period are not yet well constrained, and so we evaluate this period with a temperature range of 215-235K, noting that warmer or colder temperatures are possible based on warming (or lack thereof) from impacts and volcanism (e.g., Toon et al., 2010; Halevy and Head, 2014). This period exhibits higher geothermal heat fluxes, and is thus also applicable for younger periods during martian history in areas of anomalous heat flux, such as in volcanic terrains (e.g., Fassett and Head, 2006, 2007; Cassanelli et al., 2015). Results for heat fluxes of 60 mW/m² are discussed above, and so the following discussion is for a heat flux of 100 mW/m².

Under potential Early-Mid Noachian conditions ($T_s=235$ K, $Q=100$ mW/m²), ejecta temperatures range between 250K and 550 K for craters between 5 km and 150 km in diameter (Fig. 4). Craters larger than 15 km in diameter are predicted to exhibit contact melting (red star in Fig. 15A) and produce melt volumes ranging from 10⁰ to 10⁴ km³ (Fig. 7B, E).

For a surface temperature of 215 K and the average rim-height case, surface ice 1000 m thick is predicted to exhibit basal melting for crater diameters as low as 50 km in diameter, and 100 m thick ice is predicted to exhibit basal melting for a restricted crater diameter range between 80-85 km in diameter (Fig. 8A). Melt volumes range from 10¹ km³ for the 100 m thick ice sheet, and from 10⁰ to 10⁴ km³ for the 1000 m thick ice sheet (Fig. 8A, D), and basal melting may extend out to 0.1-0.8R from the rim-crest (Fig. 8C). For the maximum rim-height case, surface ice 1000 m thick is predicted to exhibit basal melting for crater diameters as low as 25 km in diameter, and 100 m thick ice is predicted to exhibit basal melting for a restricted crater diameter range between 35-90 km in diameter (Fig. 8D). Melt volumes range from 10¹ km³ for the 100 m thick ice sheet, and

from 10^0 to 10^4 km³ for the 1000 m thick ice sheet (Fig. 8D), and basal melting may extend out to 0.1-1.1R from the rim-crest (Fig. 8F).

In a slightly warmer environment ($T_s=235$ K), and the average rim-height case, a 10 m thick ice layer will still not exhibit basal melting because it has already been fully contact melted. Basal melting is predicted to occur for crater diameters between 45-105 km for a 100 m thick ice layer (red star in Fig. 15B; Fig. 9A). In the case of a 1 km thick surface ice sheet, basal melting will occur even in the absence of an impact crater and insulating ejecta (Fig. 9A). Basal melting may extend out to 0.1-0.6R from the rim-crest for 100 m thick ice, or 2.5R for 1000 m thick ice. For the maximum rim-height case, a 10 m thick ice layer will produce basal melting for crater diameters between 15-20 km, and a 100 m thick ice layer will produce basal melting for crater diameters between 20-100 km (red star in Fig. 15B; Fig. 9D). Basal melting may extend up to 0.2-0.9R from the rim-crest for 10-100 m thick ice sheets, and 2.5R for 1000 m thick ice sheets. Melt volumes range from 10^{-1} to 10^5 km³ (Fig. 9A, D).

In summary, contact melting is predicted to occur throughout martian history for all craters greater than ~40 km in diameter that impacted into surface ice sheets. In contrast, basal melting is not predicted in the Amazonian period, except perhaps for craters larger than 150 km in diameter which we did not explore in this study. Higher heat fluxes in the Hesperian period allow basal melting to occur under a relatively small parameter range, only for large craters >85-140 km in diameter for the upper-end predicted surface heat flux of 60 mW/m² and 1000 m thick ice sheets. Due to the potential for higher heat fluxes and higher surface temperatures in the Noachian period, basal melting is predicted to occur under a relatively broader range of crater diameters, potentially for craters as small

as ~5-~30 km depending on the surface temperature and ice thickness. Because contact melting removes a substantial portion of surface ice prior to the onset of basal melting, the ice thickness (Fastook and Head, 2015) plays a major role in limiting the crater diameters which are predicted to exhibit basal melting. Thus, ice sheet thicknesses in the 10 to 100 meter range are generally predicted to produce basal melting under a more restricted crater diameter range, even during the Noachian period.

On the basis of the predicted basal meltwater hydrologic geometry (Fig. 2F) and the restricted diameter range predicted to exhibit basal melting (compared with Noachian-aged craters which exhibit fluvial channels) (Fig. 15B), basal melting is not predicted to erode substantial volumes of ejecta, and is not considered a candidate process which can carve any valley network features. Contact melting, however, is predicted to occur under a broader parameter range in the Noachian period (Fig. 15A), and may contribute to the lack of observable ejecta associated with Noachian-aged craters (e.g., Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Mangold et al., 2012b).

5. Candidate craters for ejecta-induced ice melting

As discussed above, contact melting is predicted to occur under a wide range of impact and climatic conditions. In contrast, basal melting is theoretically plausible only under certain impact diameter and climatic conditions. The thermal models derived here can place important constraints on the crater diameters, target properties, and locations (relative to the crater) where contact and basal melting are predicted to occur. Higher surface temperatures ($\geq 225\text{ K}$) and/or heat fluxes, such as those expected during the

Noachian period, reduce the crater diameter and ejecta thickness conditions necessary to generate melting of surface ice. With these guidelines, we assess if there are any observed impact crater-associated fluvial channels that are consistent with either a contact or basal melting origin.

5.1. Candidate craters for contact melting

Previous investigators have identified fluvial channels superposing the crater walls, rim, and ejecta of Hesperian- and Amazonian-aged impact craters (e.g., Hobbs et al., 2016), and interpreted these to result from contact melting (Morgan and Head, 2009; Mangold, 2012; Mangold et al., 2012a; Schon and Head, 2012). Figure 1A shows an example of several fresh fluvial channels associated with the ejecta of two impact craters (26 km and a 12 km diameter) characterized by Mangold (2012). The larger 26 km diameter crater exhibits inlet channels on the rim (red and white arrows in Fig. 1A) and fluvial channels on the ejecta. The 12 km diameter crater in Fig. 1A (on the right) exhibits fluvial channels on the ejecta, but also exhibits fluvial channels within the surrounding terrain, draining away from the ejecta (middle blue and white arrows in Fig. 1A). Figure 16A and d are two more example of relatively young (Amazonian to Hesperian-aged) and small (19 and 26 km diameter) impact craters identified by Mangold (2012) which exhibit fluvial channels on their ejecta facies. The 26 km diameter crater in Fig. 16D also shows some fluvial channels in the surrounding terrain draining away from the ejecta (Fig. 16E). A larger example is shown in Fig. 17A; this 47 km diameter crater identified by Mangold (2012) exhibits fluvial channels superposing the ejecta facies (Fig. 17B, C). The 63 km diameter Sinton crater (Fig. 17D), characterized by Morgan and

Head (2009) is another larger impact crater which exhibits fluvial channels on the crater wall (draining into the crater interior) (Fig. 17E), and on the ejecta facies (Fig. 17F).

Recently, Hobbs et al. (2016) characterized the fluvial channels associated with two Hesperian-aged craters. Figure 18A shows a 74 km diameter crater characterized by Hobbs et al. (2016) which exhibits fluvial channels on the ejecta (Fig. 18B) and on the interior crater walls (Fig. 18C). Figure 18d shows the 35 km diameter Choyr crater characterized by Hobbs et al. (2016) which exhibits channels on the interior crater walls draining into the crater interior (Fig. 18E). These particular channels may be fluvial or glacial in origin (Hobbs et al., 2016). Interestingly, a nearby pre-existing ~10 km diameter appears to be infilled and exhibits a large fluvial channel draining out of the crater (Fig. 18F).

As noted by Mangold (2012), most fluvial features observed are on the ejecta facies, while larger craters may exhibit channels draining into the crater interior. Mangold (2012) also found that channels in the surrounding terrain are not observed as frequently, and that most of the fluvial landforms associated with impact ejecta exhibit amphitheater-shaped heads and are not located on topographic highs within the ejecta. Mangold (2012) interpreted these observations to suggest that the fluvial channels were consistent with “local water outbursts” (e.g., Fig. 12C) rather than precipitation. Mangold (2012) also noted that craters with fluvial channels on the ejecta are concentrated in the mid-latitudes (25-45°), which is consistent with crater formation in a regional surface ice sheet which was deposited in the mid-high latitudes during periods of higher martian obliquity (e.g., Head et al., 2006a, b; Madeleine et al., 2009; Fastook et al., 2011). Mangold (2012) concluded that these observations suggest that the fluvial channels were formed by

contact melting of hot ejecta on the surface icy deposits. We find the fluvial features around the 35 km diameter Choyr crater (Fig. 18D) to be particularly consistent with a contact melting origin. For example, if surface ice deposits were present at the time of impact and infilled the smaller crater (akin to Amazonian concentric crater fill deposits; Levy et al., 2010; Fastook and Head, 2014), hot ejecta from Choyr landing on the smaller ice-filled crater would be predicted to melt a substantial volume of ice, leading to a fluvial channel draining out of the crater (e.g., Fig. 19F).

Our modeling results support these interpretations. None of these craters are predicted to exhibit basal melting (Fig. 8A, B) due to their small size, and low surface temperature and heat flux. Our models predict that the onset diameter for contact melting in the Hesperian and Amazonian periods is ~30-40 km. Six of the craters discussed above (which are 26 km, 26 km, 35 km, 47 km, 63 km, and 74 km in diameter) are broadly consistent with these results. These craters are predicted to generate substantial amounts of meltwater: $\sim 10^0$ - $\sim 10^2$ km³ of meltwater (26 km), $\sim 10^1$ - $\sim 10^2$ km³ (35 and 47 km), $\sim 10^2$ - $\sim 10^3$ km³ (63 and 74 km). Mangold (2012) found that fluvial channels on the interior crater walls are only found on larger craters. We note that this is a prediction of the contact melting models, which show that larger craters are able to melt substantially greater thicknesses of ice near the rim (Fig. 6), which can generate the ice slope reversal necessary for sustained inflow of contact meltwater into the crater (Fig.12). The smaller craters (12 and 19 km in diameter) discussed above are smaller than the estimated onset diameters (~30 km) from our thermal models. As noted by Mangold (2012), however, the fluvial features associated with the 12 km diameter crater (Fig. 1A) may actually pre-date the impact. Nonetheless, of the 27 impact craters identified with fluvial landforms on the

ejecta by Mangold (2012), nine are smaller than 25 km in diameter. We attribute this difference to a combination of factors: (1) our shock pressures (and post-shock temperatures) neglect the contribution from pore ice, and so our shock heating may be slightly underestimated; and (2) the removal of thick surface ice sheets will decrease the observed crater diameter (Fig. 5). For example, a 30 km diameter crater that formed in a surface ice sheet is predicted to appear to be ~26 km in diameter following the removal of a 300 m ice sheet, or 22 km in diameter following the removal of a 700 m ice sheet (Fig. 5).

5.2. Candidates for basal melting: Noachian craters

The Amazonian and Hesperian-aged fluvial features associated with impact crater ejecta ($< \sim 85$ km in diameter and for ice sheets $< \sim 1$ km thick) can be attributed to contact melting (see Section 4). We now examine impact craters which are candidates for basal melting. Three examples of typical impact craters located in the Noachian-aged southern highlands are shown in Fig. 19. The two craters in Fig. 19A and B exhibit fluvial channels nearly circumferentially around the rim, and all three craters exhibit subdued rim-crests and a lack of observable ejecta. On the basis of their highly degraded state (e.g., Mangold et al., 2012b), these craters are interpreted to be Noachian in age. We first describe the model predictions for craters of this size and age, and then assess whether the basal melt fluxes predicted by the models are sufficient to form fluvial channels. Finally, we assess the unique geomorphologic characteristics that support a contact and/or basal melting origin for these two examples.

Figure 19A shows a ~180 km diameter crater of Noachian age. A crater of this

diameter is predicted to generate $\sim 10^4 \text{ km}^3$ of meltwater through contact melting. A wide range of heat fluxes are sufficient to generate basal melting for a crater of this size (Figs. 8-10) for T_s between 215 and 255 K. For a heat flux of 60 mW/m^2 and an ice sheet 1 km thick, predicted meltwater volumes range up to 400 km^3 for the average rim-height case ($T_s=215 \text{ K}$; Fig. 8A), and up to $20,000 \text{ km}^3$ for the maximum rim-height case ($T_s=235 \text{ K}$; Fig. 9D). The 10 and 100 m thick ice sheet scenarios will not provide any meltwater because contact melting is predicted to fully melt these in the locations where basal melting could occur (except for $T_s=255$, where an ice thickness of 100 m provides $\sim 10^2 \text{ km}^3$ of meltwater; Fig. 10A, D). Melt fluxes are predicted to be 3×10^{-4} to $4 \times 10^{-3} \text{ m/yr per m}^2$. The time-averaged melt flux values given in m/yr per m^2 can be converted to a peak melt flux into the crater by multiplying by the area of the annulus predicted to supply melt. The peak melt flux into the crater interior is estimated to be 14.5 to $1000 \text{ m}^3/\text{yr}$ over the course of ~ 40 - 250 kyr . This value represents the peak flux because it portrays the case where the entire annulus around the crater is undergoing basal melting (as discussed in Section 3, basal melting is predicted to initiate at the rim where the ejecta is thickest, and continue outwards). For comparison, the mean annual discharge of the Onyx River, Wright Valley, Antarctica between 1970 and 2003 was $\sim 3 \times 10^6 \text{ m}^3/\text{yr}$ (Gooseff et al., 2007)

Figure 19B shows a $\sim 50 \text{ km}$ diameter Noachian-aged crater. A crater of this diameter is predicted to generate $\sim 10^3 \text{ km}^3$ of meltwater through contact melting. For $T_s=235 \text{ K}$ and the maximum rim-height case (Fig. 9D), this crater may exhibit basal melting for a heat flux of 60 mW/m^2 for surface ice sheets between 100 m and 1000 m thick. Predicted meltwater volumes range from 45 km^3 (100 m ice sheet), up to 350 km^3 (1000 m ice

sheet). The 10 m thick ice sheet (and all $T_s \leq 235$ K with average rim-height) scenarios will not provide any meltwater because contact melting has already melted these deposits, and the 215 K scenarios do not predict sufficiently high basal temperatures. Melt fluxes are predicted to be 10^{-3} m/yr per m^2 (Fig. 9E). In these cases ($T_s = 235$ K, 60 mW/m^2), the maximum radial distance from the rim-crest that basal melting is predicted to occur ranges from $0.2R$ and $0.4R$ from the rim (Fig. 9F), yielding peak melt fluxes into the crater of $1.2 \text{ m}^3/\text{yr}$ to $2.3 \text{ m}^3/\text{yr}$ over the course of ~ 50 - 300 kyr.

Figure 19C shows a ~ 62 km diameter Noachian-aged crater characterized by Howard (2007). A crater of this diameter is predicted to generate $\sim 10^3 \text{ km}^3$ of meltwater through contact melting. For $T_s = 235$ K and the average rim-height case (Fig. 9A), this crater may exhibit basal melting for a heat flux of 60 mW/m^2 for surface ice sheets of ~ 1000 m thick, producing $\sim 30 \text{ km}^3$ of meltwater. For the maximum rim-height case (Fig. 9D), this crater may exhibit basal melting for a heat flux of 60 mW/m^2 for surface ice sheets between 100 m and 1000 m thick. Predicted meltwater volumes range from 55 km^3 (100 m ice sheet), up to 1000 km^3 (1000 m ice sheet). The 10 m thick ice sheet (and all $T_s \leq 235$ K with average rim-height below 1000 m thick ice) scenarios will not provide any meltwater because contact melting has already melted these deposits, and the 215 K scenarios do not predict sufficiently high basal temperatures. Melt fluxes are predicted to be 10^{-3} m/yr per m^2 (Fig. 9E). In these cases, the maximum radial distance from the rim-crest that basal melting is predicted to occur ranges from $0.3R$ and $0.5R$ from the rim (Fig. 9F), yielding peak melt fluxes into the crater of $3.7 \text{ m}^3/\text{yr}$ to $5.7 \text{ m}^3/\text{yr}$ over the course of ~ 40 - 300 kyr.

In order for a basal melting origin to be consistent with the fluvial features, the melt

fluxes must be sufficient to generate fluvial incision. Although the melt fluxes predicted for basal melting are low ($\sim 10^{-3}$ m/yr per m^2), it is important to note that the young valley networks present around some of the martian shield volcanos (Fassett and Head, 2006, 2007) are interpreted to form by basal melting of snow pack, and have comparable heat fluxes to the basal melting scenario considered here. For example, assuming peak heat fluxes between 130-220 mW/m^2 (Fassett and Head, 2007), a constant snow thermal conductivity of 1 K/ m K, and a snow thickness of 500 m, peak melt fluxes (from eq. 6) are predicted to be between $\sim 4 \times 10^{-4}$ and $\sim 9 \times 10^{-3}$ m/yr per m^2 for the valley networks considered by Fassett and Head (2006, 2007) (compared to $\sim 10^{-3}$ m/yr per m^2 for the impact craters considered here). In other words, similarly low melt fluxes appear to have plausibly generated fluvial channels on some martian shield volcanos, which suggests that the basal melt fluxes predicted in this study are also sufficient to form fluvial channels.

On the basis of the circumferential interior wall channel morphology observed (red and white arrows in Fig. 19) and model results, we propose that the fluvial channels on the interior crater walls (Fig. 19A, B) could have formed through basal melting of surface ice underlying the crater ejecta. If surface ice were present at the time of impact, basal melting of this deposit is predicted to lead to meltwater transported towards the crater interior (Fig. 2E). The meltwater would then be able to exit and flow down the slopes of the crater walls, where fluvial incision and/or liquid water-assisted debris flows could contribute to channel formation (Fig. 2F). Contact melting could also be responsible for the formation of the interior wall channels (Fig. 6D) in a manner similar to the Hesperian-aged 63 km diameter Sinton crater (Morgan and Head, 2009).

While the fluvial channels circumferentially arranged around the rims of these craters are consistent with either a contact or basal melting origin, their morphology does not exclusively require contact/basal melting to form. That is, other previously proposed formation mechanisms, such as rainfall and surface runoff (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Howard et al., 2005; Grant et al., 2015; Irwin et al., 2015) or surface snowmelt and runoff (e.g., Howard et al., 2005; Head and Marchant, 2014; Weiss and Head, 2015; Irwin et al., 2015; Hobbs et al., 2016) cannot be ruled out. We identified another characteristic of these typical craters that appears to be exclusively consistent with a contact/basal melting origin. The crater shown in Fig. 19A partially intersected two smaller pre-existing craters located in the south-eastern quadrant of the image (features D and E in Fig. 19A). Two anomalously large valleys associated with these impact craters (features D and E in Fig. 19A) appear to initiate at the crater center, and drain down into the larger crater. In a cold and icy Mars, impact craters are predicted to serve as topographic lows that accumulate large thicknesses of surface ice (akin to the Amazonian debris-covered glacial deposits known as concentric crater fill (e.g., Levy et al., 2010; Fastook and Head, 2014). Upon impact, the ejecta of the 180 km diameter crater will superpose both the surrounding icy surface and the pre-existing ice-filled impact craters (i.e., CCF). Contact and basal melting of any ice that infills the pre-existing craters (e.g., Fig. 19A) will then produce higher volumes of meltwater (relative to the surrounding terrain), leading to greater degrees of channel incision. This interpretation is identical to our interpretation discussed above for some of the fluvial channels associated with the younger, Hesperian-aged Choyr crater (Fig. 18F).

A similarly incised valley associated with an older (~15 km diameter) pre-existing impact crater is located in the north-west quadrant of the 50 km diameter crater (feature F in Fig. 19B). In a similar manner to that shown in Fig. 19A, if this topographically low ~15 km diameter impact crater (feature F in Fig. 19B) were filled with surface ice at the time of the impact event that formed the 50 km diameter crater, the ejecta from this event would have infilled the older crater (feature E in Fig. 19B) with hot ejecta, leading to contact and basal melting of the thick ice deposit (i.e., CCF) and enhanced fluvial incision. The 62 km diameter crater in Fig. 19C shows a similar relationship, in which the fluvial channels are concentrated in a pre-existing ~38 km diameter crater (feature G) that appears to be infilled by the ejecta of the 62 km diameter crater. In a manner similar to the previous examples, if the pre-existing crater were filled with ice prior to the impact of the 62 km diameter crater, the ejecta from this event would have infilled the older crater (feature G in Fig. 19C), leading to contact and basal melting of the thick ice deposit. Unlike the previous examples, feature G in Fig. 19C appears to exhibit a branching network of channels upstream to the main valley, and is thus distinct from the Amazonian examples (Fig 16-18) discussed above.

These observations (features D and E in Fig. 19A, feature F in Fig. 19B, and feature G in Fig. 19C) raise the possibility that surface ice deposits were present within the pre-existing craters at the time of impact (within the Noachian period), and that contact/basal melting of these deposits produced the associated channels. These fluvial features appear to be anomalously large when compared with the circumferential wall channels, but they are present on the low-sloped floors of the pre-existing craters as opposed to the steep crater walls. The features of the typical impact craters discussed above (Fig. 19) are

unique (intersecting pre-existing impact craters with larger superposing fluvial features on relatively flat topography), and point to a major contribution from contact and/or basal melting. We note that a contact and/or basal melting origin is also plausibly consistent with the other observed crater-associated fluvial features (i.e., circumferential fluvial channels). This then raises the possibility that contact and basal melting could be viable candidate degradation processes for all Noachian-aged impact craters $\geq \sim 15\text{-}30$ km in diameter (Figs. 7 and 9D) which exhibit fluvial channels circumferential to the rim (and not just those impact craters which intersect pre-existing low topography).

6. Conclusions

In summary, our calculations predict contact melting of surface ice sheets overlain by hot ejecta to occur over a range of crater diameters, surface temperatures, heat fluxes, and ice thicknesses. This process is predicted to occur for all craters larger than ~ 40 km in diameter which impacted into surface ice deposits, and may melt tens to hundreds of meters of surface ice (Fig. 6) over the course of ~ 100 years to 80 kyrs, corresponding to $10^0\text{-}10^4$ km³ of meltwater (Fig. 7). Meltwater is generally predicted to flow outwards (Fig. 12A), away from the crater. Fluvial channels on the ejecta may be formed by breaching of the ejecta, analogous to terrestrial springs (Fig. 12C). In the case of ice sheets of comparable thickness to the rim structural-uplift (Fig. 12C), meltwater may flow downwards into the crater (Fig. 12D).

Following the cooling of ejecta, the overlying ejecta facies is predicted to provide sufficient thermal insulation to raise the 273 K melting isotherm through the ice-cemented cryosphere to the base of the ice sheet (Fig. 11). Melting of the pore-ice within

the cryosphere will provide a large volume of water for groundwater recharge, and could contribute to subsurface chemical alteration to form clays in the shallow crust.

The ice-melting isotherm is predicted to reach the base of the ice sheet within ~100 kyrs to ~1 Myr after the impact (Fig. 11). If the ice sheet is sufficiently thick and has not been previously removed by contact melting, this can lead to basal melting of the ice sheet, which can last for ~1 kyr to ~1 Myr. Basal melting of surface ice sheets overlain by impact ejecta is predicted to occur under a more confined range of crater diameters, surface temperatures, heat fluxes, and ice thicknesses (relative to contact melting). Basal melting is predicted to occur only at near-rim ejecta thicknesses (generally within ~1R from the rim-crest). Overburden pressure from the overlying ejecta and surface ice is predicted to transport the meltwater up the structurally uplifted rim and toward the crater interior. Basal melting is generally not predicted to occur for typical Amazonian heat fluxes, but may occur for moderately large crater diameters ($\geq \sim 150$ km in diameter) in the Early Amazonian and Hesperian (Head et al., 2016), or in areas of anomalously high heat flux. For example, in the higher-end heat flux estimates in the Hesperian period, basal melting is possible for craters as small as ~85 km in diameter. The higher heat fluxes and potentially higher surface temperatures predicted for the Late Noachian period allow basal melting to occur for craters as small as ~5-30 km in diameter. We consider the basal melting process a candidate mechanism for fluvial channels located on the interior walls of ancient impact craters on Mars. The low melt fluxes could alternatively lead to low water:rock ratio flows (e.g., water-assisted debris flows). We conclude that contact and basal melting could operate as background crater degradation processes in a cold and icy early Mars even during periods which lack punctuated warming events (e.g.,

Halevy and Head, 2014; Head and Marchant, 2014; Weiss and Head, 2015; Wordsworth et al., 2015).

This study offers a theoretical treatment of the conditions required for contact and basal melting to occur as well as a brief case study of several candidate craters. Our models support previous interpretations that contact melting was responsible for the formation of fluvial channels on some Amazonian- and Hesperian-aged impact craters (e.g., Mangold, 2012). We also suggest that the fluvial features associated with three Noachian-aged impact craters are consistent with a contact and basal melting origin: The colocation of large valleys with ejecta-infilled pre-existing craters (features D and E in Fig. 19A, feature F in Fig. 19B, and feature G in Fig. 19C) raises the possibility that surface ice deposits were present within the pre-existing craters at the time of impact (during the Noachian period). In this scenario, contact/basal melting of these deposits could have produced the associated channels. Additional morphologic studies (e.g., Hobbs et al., 2016) should be conducted to evaluate whether additional evidence for the presence of impact ejecta-induced melting exists, particularly within ancient terrains.

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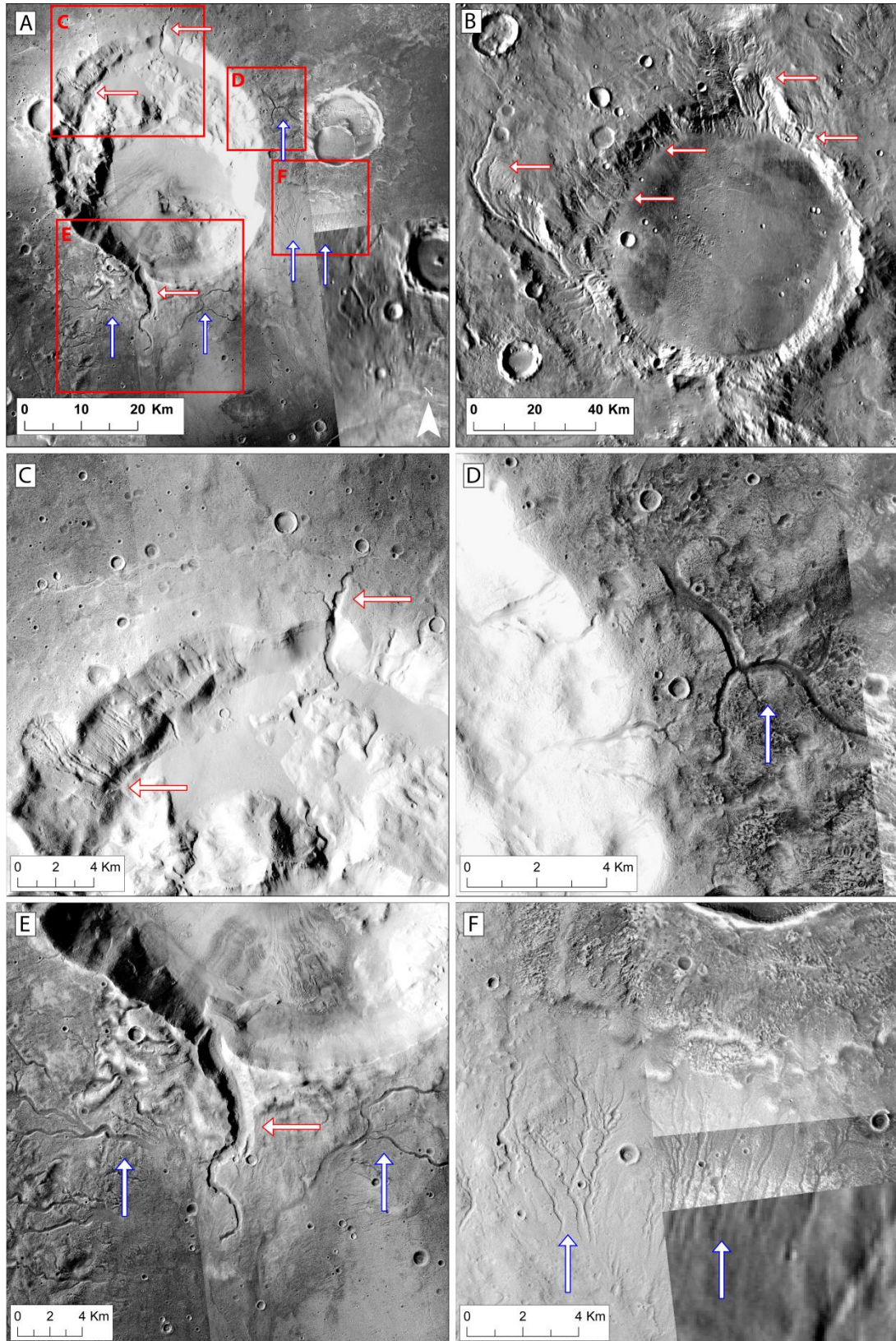


Figure 1. Examples of fluvial channels associated with impact craters on Mars. (A) A 26 km diameter closed basin lake (CBL) exhibiting inlet channels on the rim (red and white arrows) and fluvial channels superposing the ejecta (blue and white arrows), and the ejecta of a younger nearby 12 km diameter crater characterized by Mangold (2012) (6.6°E, 35.3°N). (B) A highly degraded Noachian-aged crater exhibiting numerous fluvial channels along the rim characterized by Mangold et al. (2012b) (59.1°E, 18.8°S). (C, D, E, F) Fluvial features from (A). A) CTX images B01_009892_2148, P21_009325_2160, D15_032968_2170, and D15_033047_2171 superposed on THEMIS IR global day. B) THEMIS IR global day.

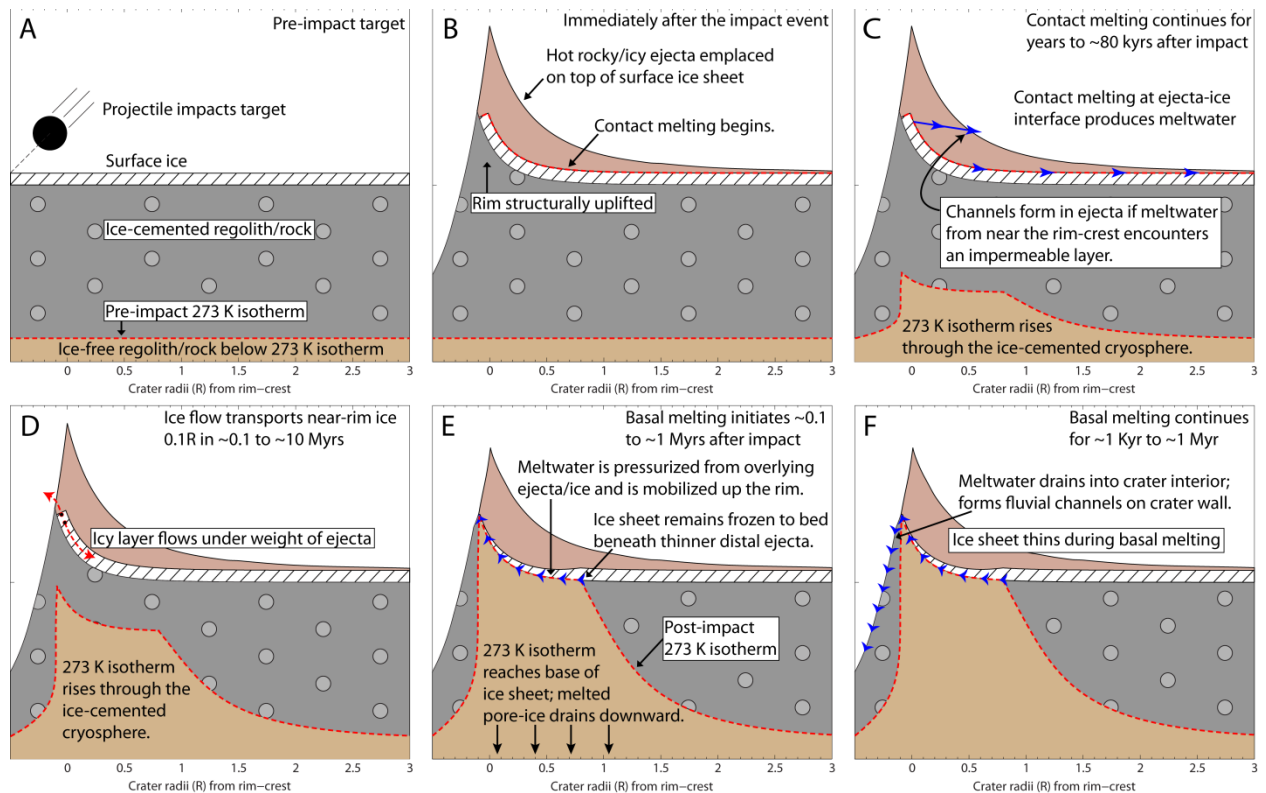


Figure 2. Post-impact melting configuration used in our models (with 55X vertical exaggeration). (A) The pre-impact target is composed of a surface ice layer overlying ice-cemented regolith/rock. The pre-impact ice-melting isotherm (273 K) (dashed red line) defines the base of the cryosphere (the zone cold enough for pore-ice stability). (B) The impact occurs, and hot ejecta is deposited on top of the surface ice; contact melting of the surface ice begins. (C) Contact melting continues and meltwater drains out of the ejecta; meltwater derived from near the topographically high rim-crest may form channels within the ejecta facies if the meltwater encounters an impermeable layer (e.g., a spring). (D) The surface ice sheet may flow, enhanced by the weight of the overlying ejecta. (E) The thermally insulating ejecta layer inhibits heat conduction, which raises the melting isotherm (273 K) (dashed red line) up through the cryosphere; the melted pore-ice then drains downward and is a source for groundwater recharge. The 273 K isotherm is raised up to the base of the ice sheet near the rim, where the ejecta is thickest. This allows for basal melting of the ice sheet; the meltwater is predicted to be transported up the crater rim (blue arrows) and towards the crater interior due to the pressurization from the overlying ejecta and ice. (F) The meltwater transported into the crater interior could form fluvial channels on the crater walls.

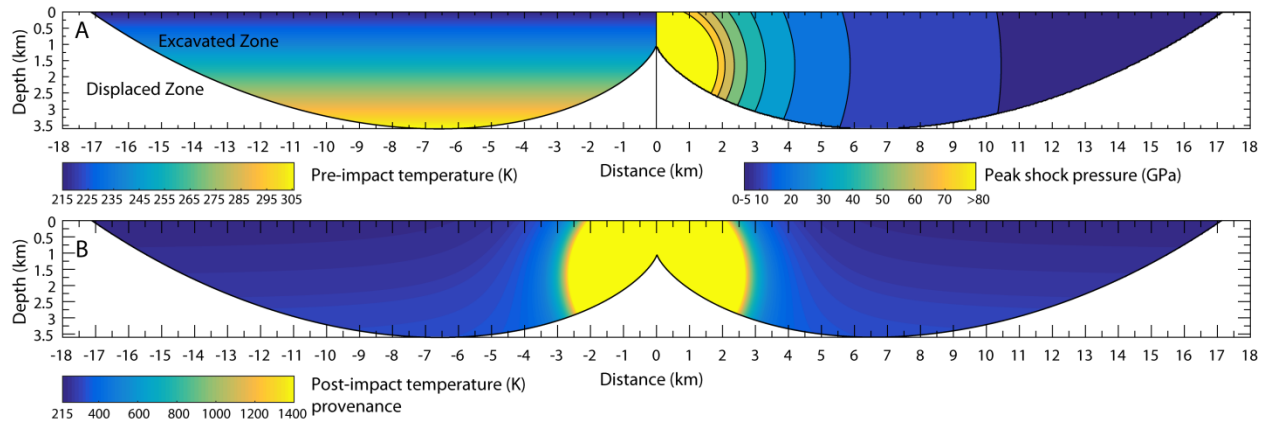


Figure 3. (A) Excavation cavity of a 50 km diameter crater showing the pre-impact ejecta temperature (left panel) from geothermal heating ($T_s=215$ K, $Q=60$ mW/m²) and the peak shock pressure within the excavation cavity (right panel) from the planar impact approximation (Melosh, 1989, p. 54; Melosh, 2012). Excavation cavity geometry is defined by the Maxwell Z model (Maxwell, 1977). (B) The post-impact ejecta temperature is found by volumetrically averaging the sum of the pre-impact ejecta temperature (A) with the post-shock temperatures derived from peak shock pressures (B).

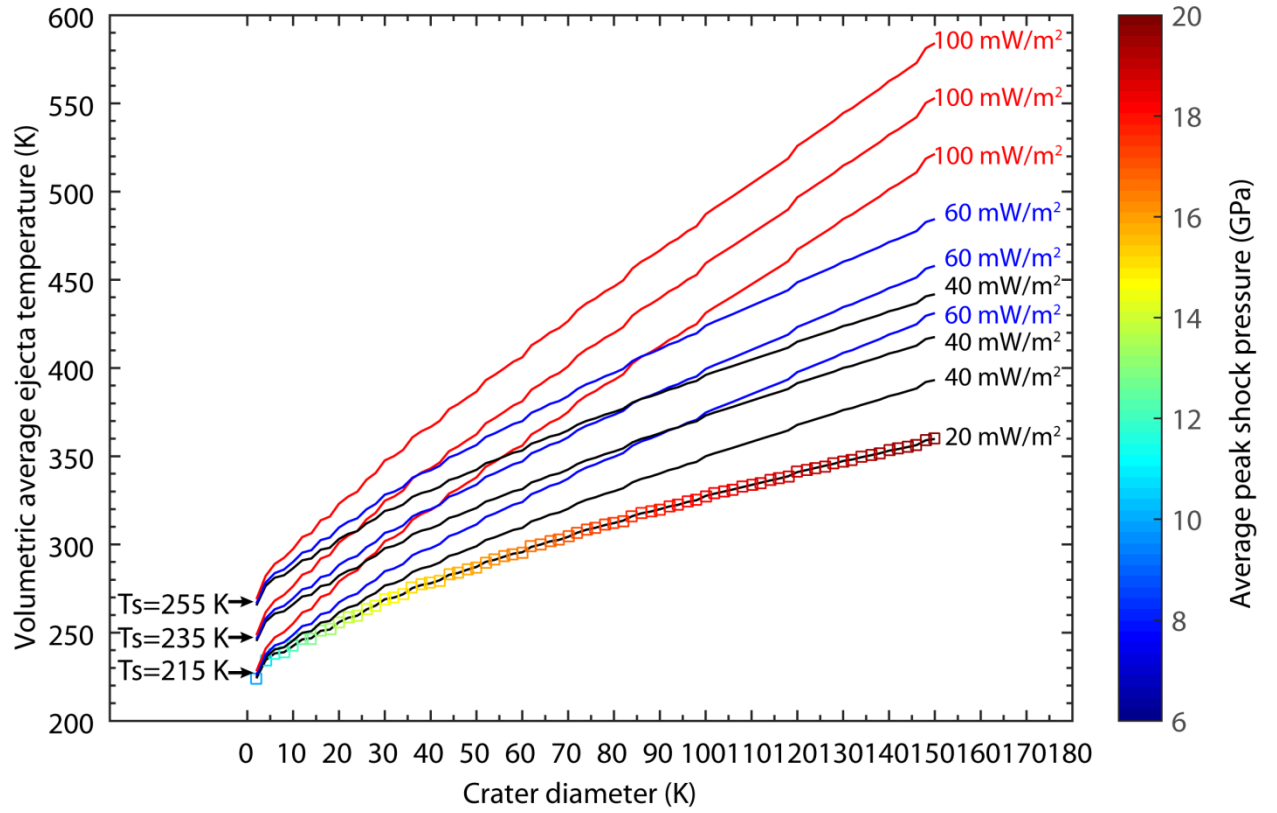


Figure 4. Volumetric average ejecta temperature (T_E) as a function of crater diameter for T_s of 215, 235, and 255 K, and surface heat fluxes of 20 mW/m², 40 (black lines), 60 (blue lines), and 100 mW/m² (red lines). Color bar indicates volumetric average peak shock pressures corresponding to each crater diameter (shown on the 20 mW/m² line).

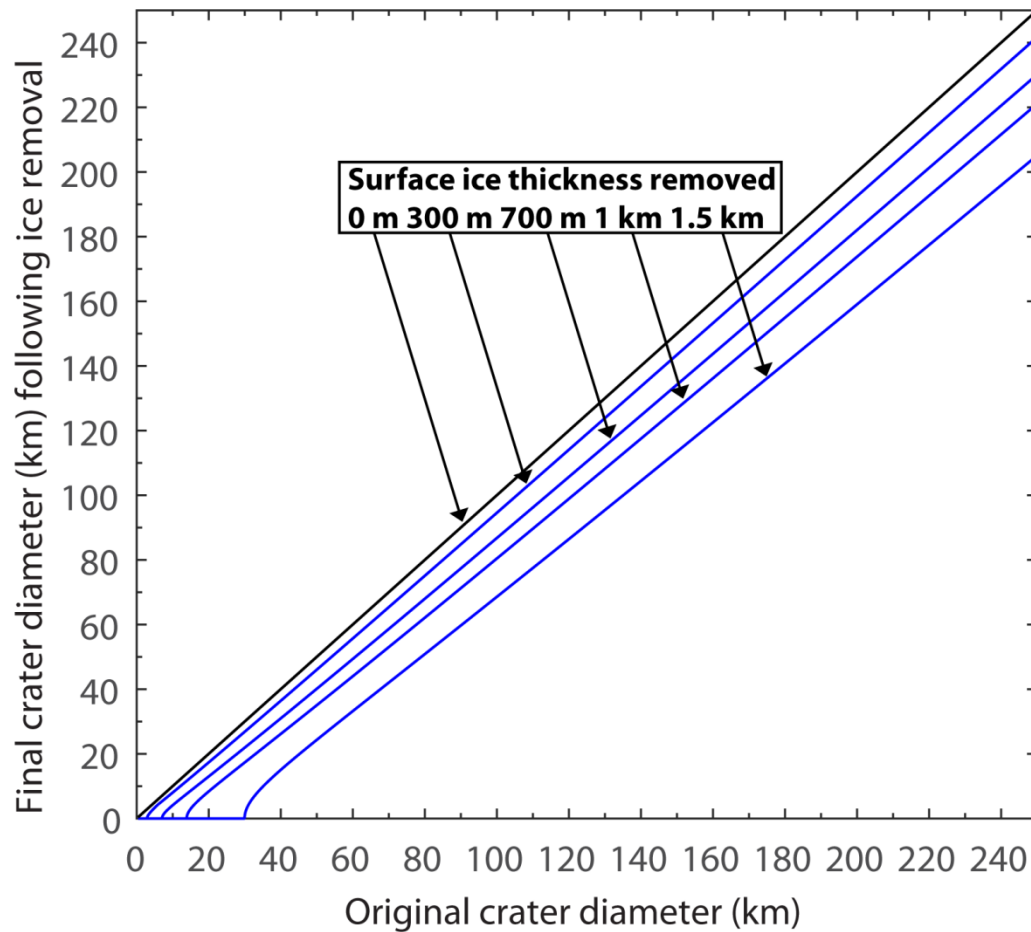


Figure 5. The final crater diameter (D) expected following removal of ice sheets (of thickness z) between 300 m and 1.5 km thick by equating an impact crater cavity as a parabola. Modified from Weiss and Head (2015): $D = \left(\frac{D_i^2(d-z)}{d} \right)^{0.5}$, where D_i is the original crater diameter and d is crater depth. If an impact occurs into a thick regional surface ice sheet, subsequent removal of the ice sheet in a different, later climate regime is predicted to decrease the observed crater diameter due to the removal of the upper hundreds of meters to kilometers of target material (the surface ice).

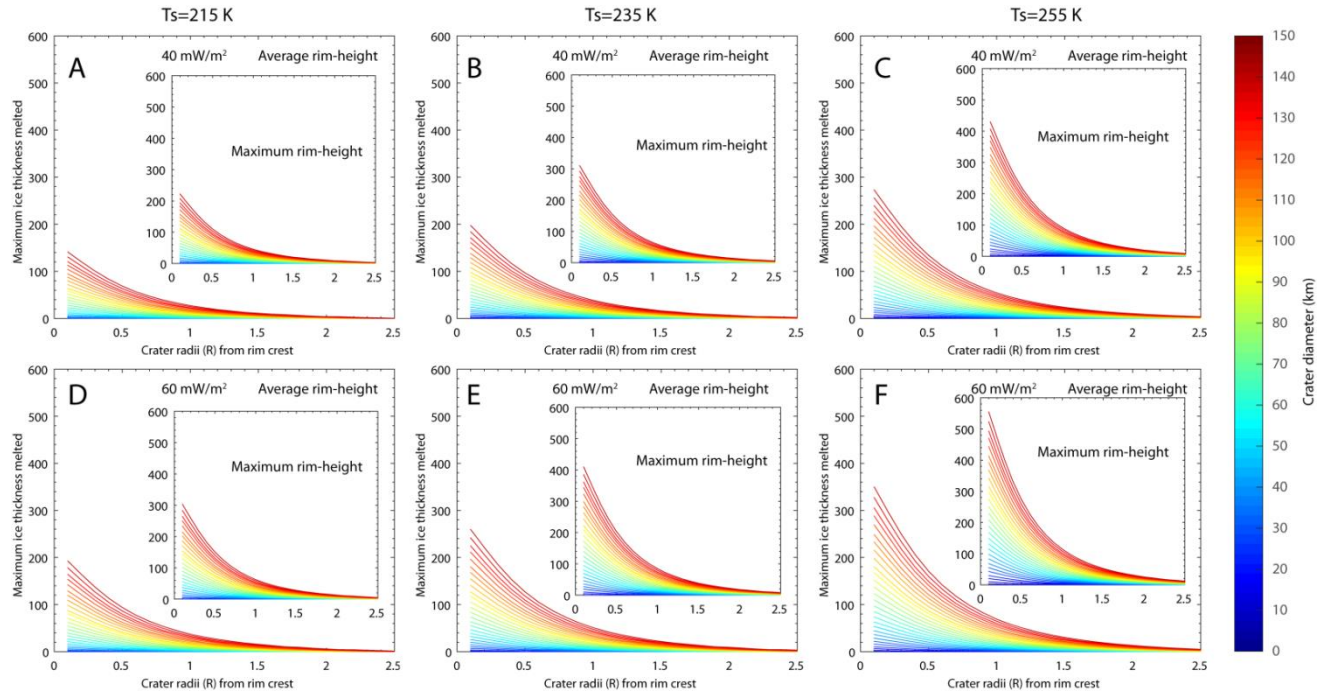


Figure 6. Contact melting thermal model results show the maximum thickness of ice melted (i.e., there is no ice thickness supply limit) for surface temperatures of 215 K (left panels), 235 K (middle panels), and 255 K (right panels). Surface heat fluxes of 40 mW/m^2 are shown in the top panels, and 60 mW/m^2 are shown in the bottom panels. The large panels show the average rim-height models, and the small inset panels show the maximum-rim height case.

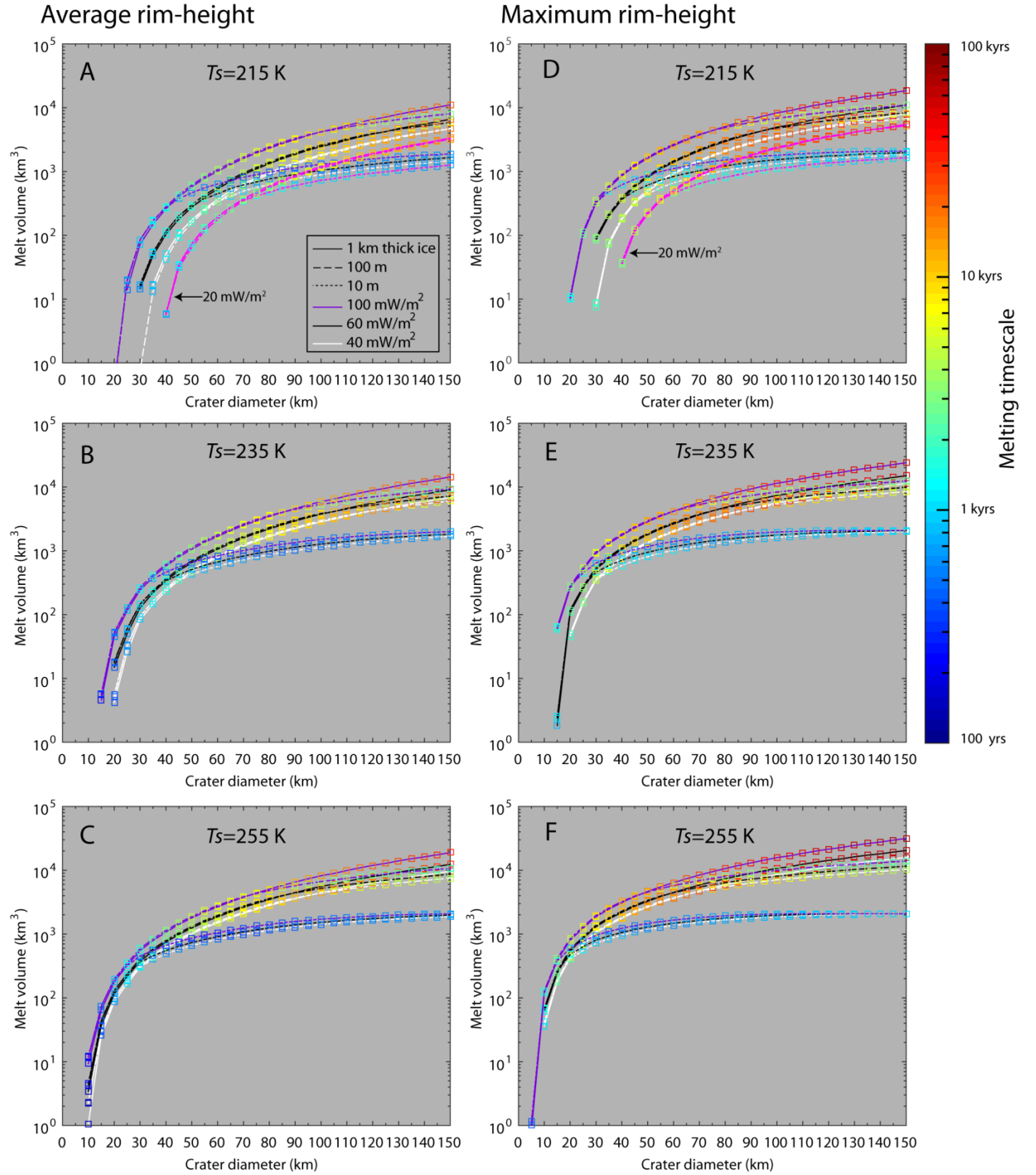


Figure 7. Contact melting thermal model results show the meltwater volume generated as a function of crater diameter, ice thickness, surface heat flux, and ejecta thickness. A) For $T_s=215\text{ K}$, B) $T_s=235\text{ K}$, C) $T_s=255\text{ K}$. Surface heat fluxes are 40 (white lines), 60 (black lines), and 100 (purple lines) mW/m^2 . We also show an Amazonian heat flux of 20 mW/m^2 (pink lines) for $T_s=215\text{ K}$ (top panels). Ice thicknesses of 1 km (solid lines), 100 m (dashed lines), and 10 m (dotted lines). D-F) Same as A-C but for the maximum rim-height case. Colored squares indicate the melting timescale.

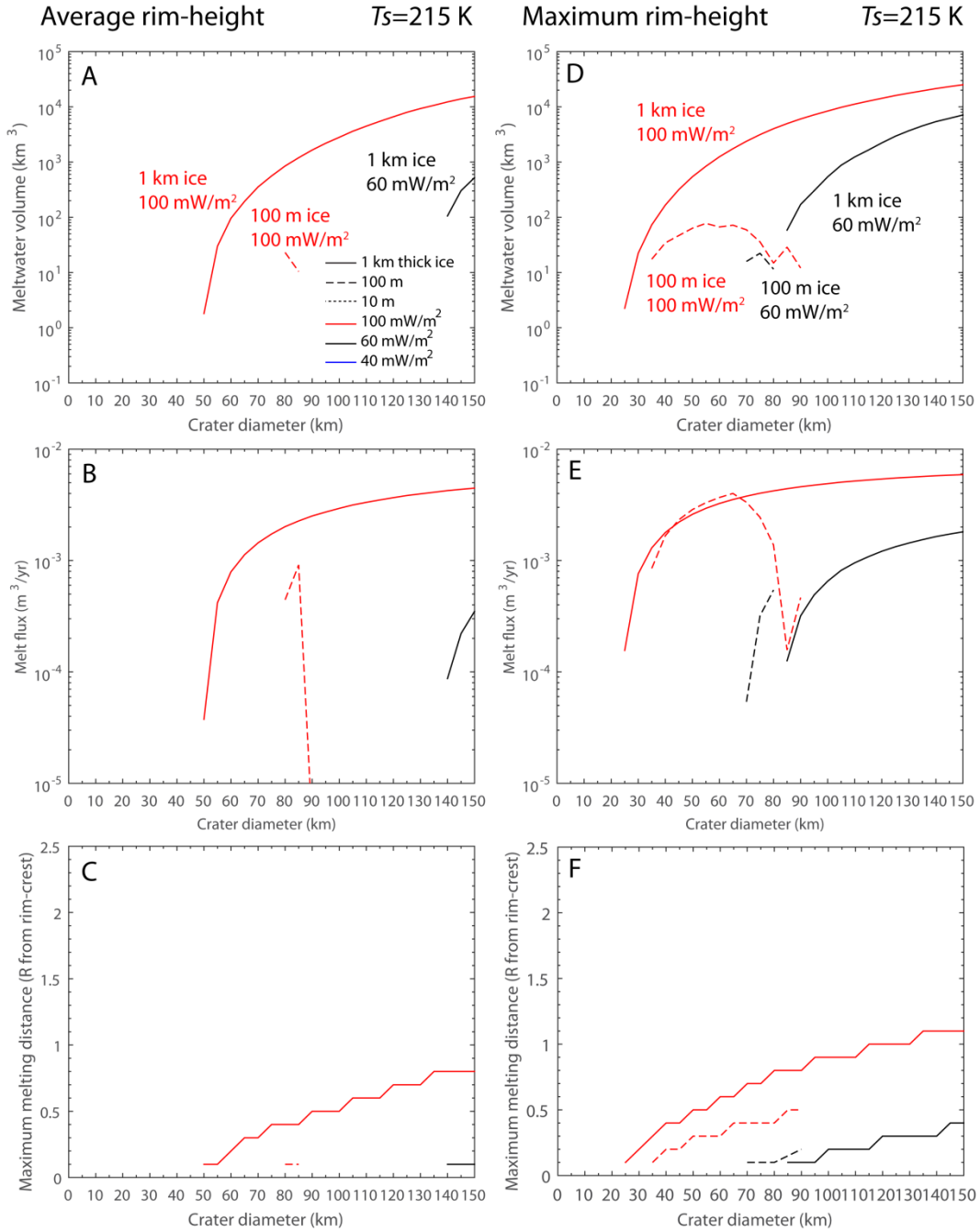


Figure 8. Model results for a surface temperature of 215 K for the average and maximum rim-height cases. (A) Meltwater volume generated for a 40 (blue) 60 (green) and 100 (red) mW/m^2 surface heat flux (average rim-height) and ice thicknesses of 1 km (solid lines), 100 m (dashed lines), and 10 m (dotted lines). Where model lines are not present, contact melting has already melted the surface ice, and so no basal meltwater is generated (except for the 40 mW/m^2 average rim-height case for $T_s = 215$ K, where basal melting is not predicted to occur). (B) Time-averaged basal melt flux as a function of crater diameter. (C) Maximum distance (in crater radii, R) of basal melting from the rim-crest versus crater diameter. (D-F) Same as (A-C) but for the maximum rim-height case.

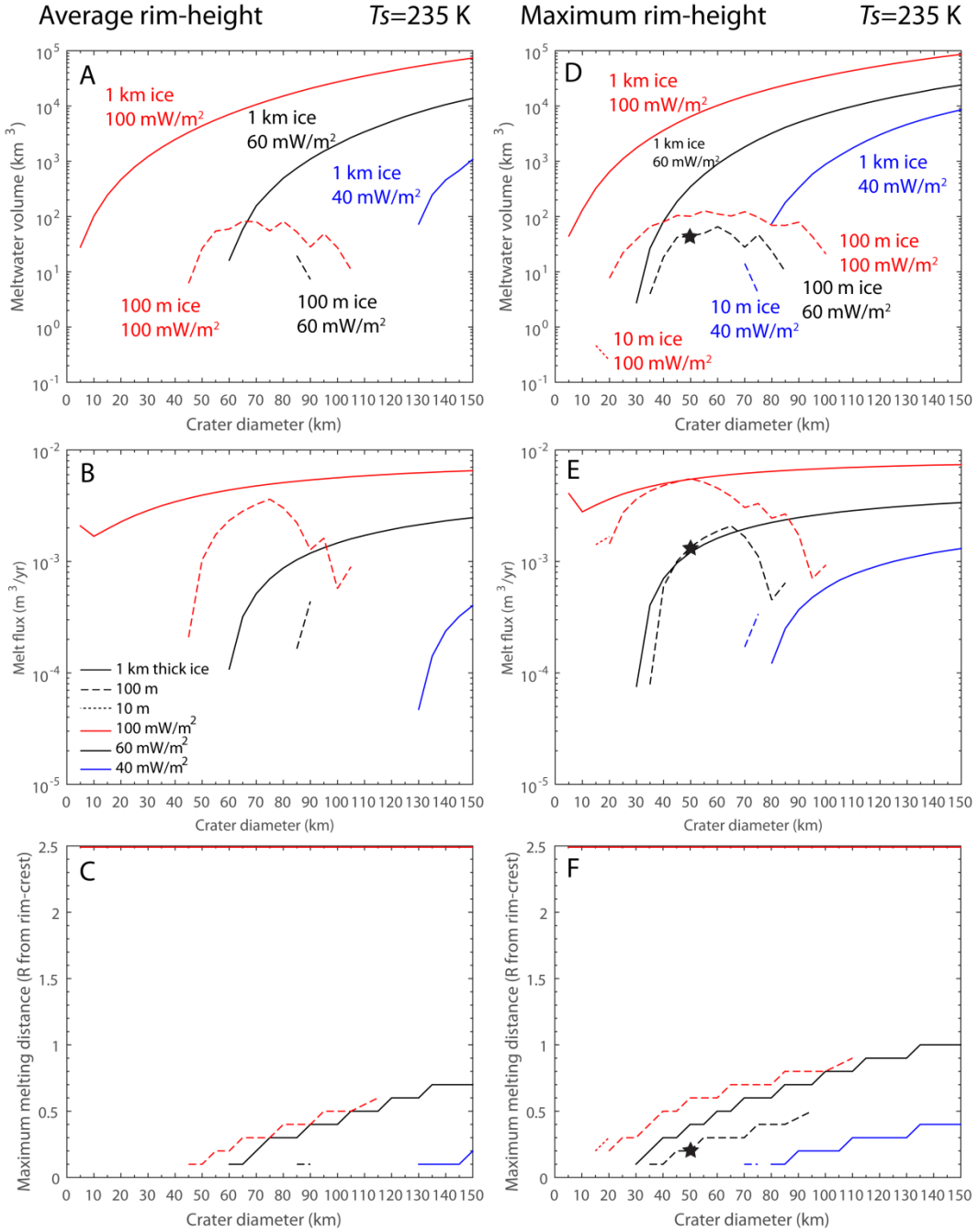


Figure 9. Model results for a surface temperature of 235 K for the average and maximum rim-height cases. (A) Meltwater volume generated for a 40 (blue) 60 (green) and 100 (red) mW/m^2 surface heat flux (average rim-height) and ice thicknesses of 1 km (solid lines), 100 m (dashed lines), and 10 m (dotted lines). Where model lines are not present, contact melting has already melted the surface ice, and so no basal meltwater is generated. (B) Time-averaged basal melt flux as a function of crater diameter. (C) Maximum distance (in crater radii, R) of basal melting from the rim-crest versus crater diameter. (D-F) Same as (A-C) but for the maximum rim-height case. See text in Section 3.2. for explanation of the black stars.

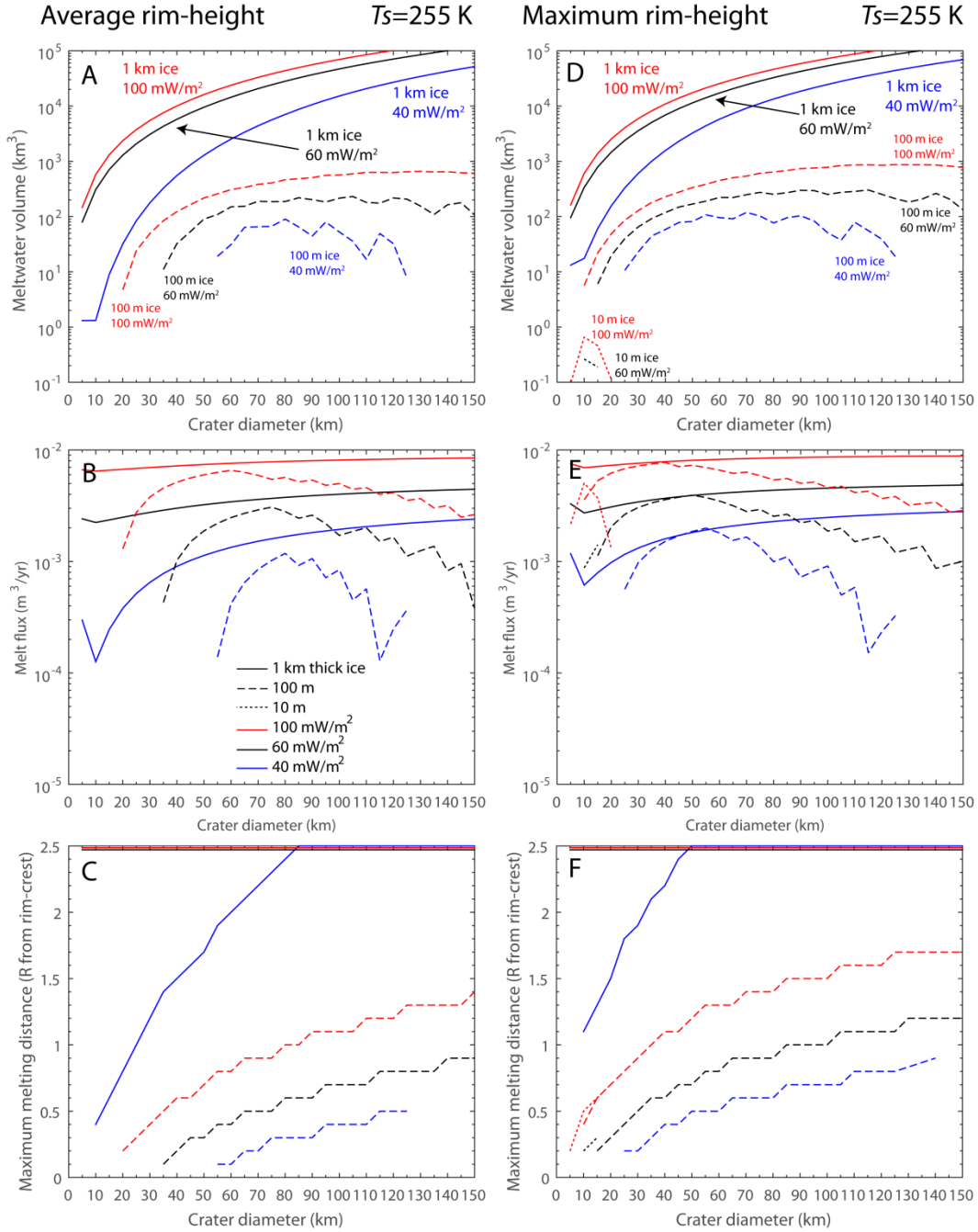
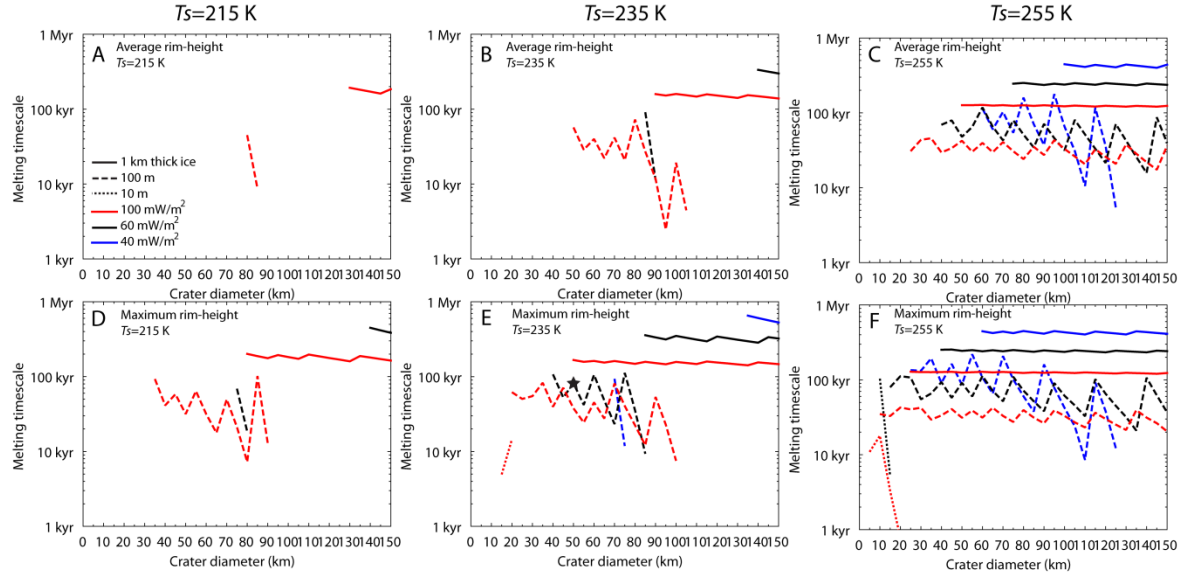


Figure 10. Model results for a surface temperature of 255 K for the average and maximum rim-height cases. (A) Meltwater volume generated for a 40 (blue) 60 (green) and 100 (red) mW/m^2 surface heat flux (average rim-height) and ice thicknesses of 1 km (solid lines), 100 m (dashed lines), and 10 m (dotted lines). Where model lines are not present, contact melting has already melted the surface ice, and so no basal meltwater is generated. (B) Time-averaged basal melt flux as a function of crater diameter. (C) Maximum distance (in crater radii, R) of basal melting from the rim-crest versus crater diameter. (D-F) Same as (A-C) but for the maximum rim-height case.

Surface ice basal melting timescale



Cryosphere melt-through (basal melting initiation) timescale

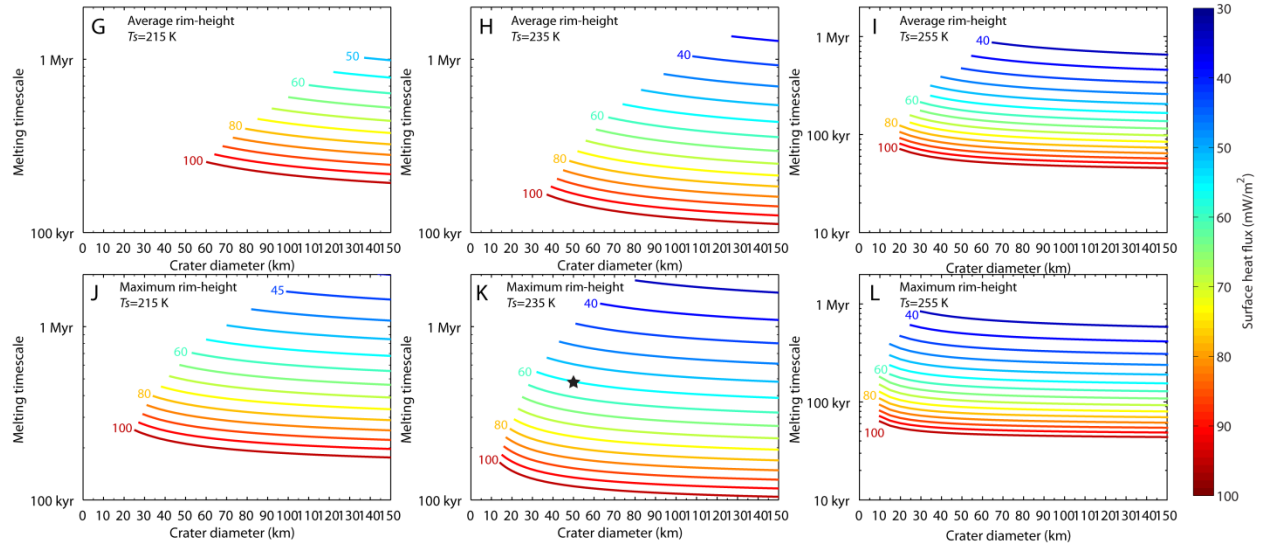


Figure. 11. Timescales for basal melting. (A-F) show the melting timescale for the surface ice, for surface temperatures of 215 K, 235 K, and 255 K for both the average and maximum rim-height scenario. Where model lines are not present, contact melting has already melted the surface ice, and so no basal meltwater is generated. (G-L) shows the timescale for the 273 K isotherm to reach the base of the ice sheet (i.e., the timescale to melt through the ice-cemented cryosphere). Melting timescales are shown only for the case where the entire thickness of the ice sheet (A-F) or ice-cemented cryosphere (G-L) is melted through. See text in Section 3.2. for explanation of the black stars.

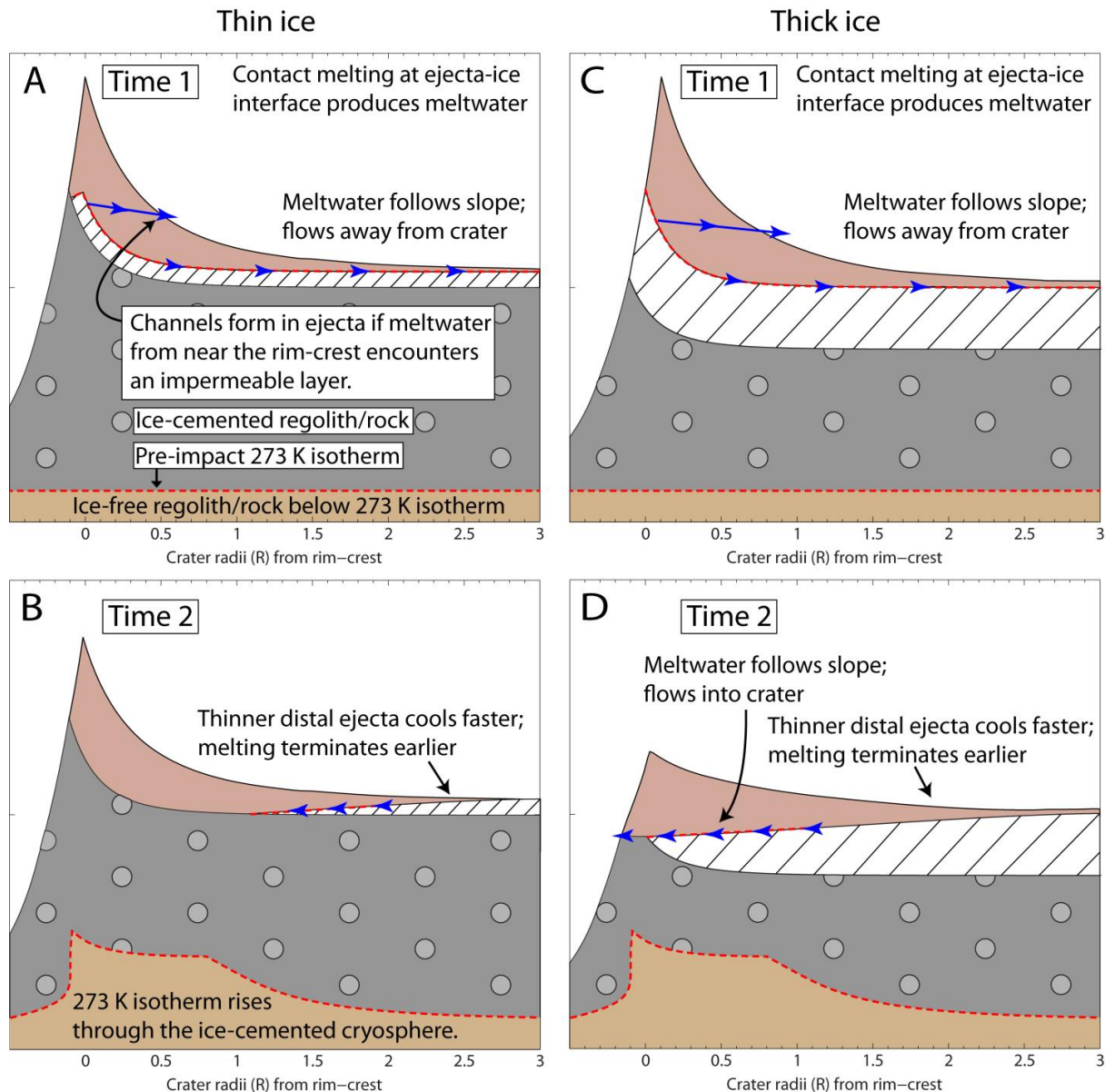


Figure 12. Contact melting scenario for thin ice (left panels; a, b) and thick ice (right panels; C, D) for early times (top panels) and late times (bottom panels). (A) For the thin ice case, melt will flow away from the crater during the early times when contact melting initiates. If the near-rim melt encounters an impermeable layer within the ejecta, spring-like breaching of the ejecta may generate fluvial channels on the ejecta facies. (B) In the later times, the near-rim ice has melted, reversing the slope of the topography. Melt may runoff toward the crater, but because the ice is lower than the structurally uplifted rim-crest, it is predicted to pool within the ejecta. (C) For the thick ice case, melt will flow away from the crater during early times when contact melting initiates. As in the thin ice case, breaching of the ejecta by a spring would form fluvial channels on the ejecta. (D) In the later times, a substantial thickness of near-rim ice has melted, which reverses the topographic slope. In this case, meltwater is predicted to flow into the crater interior, and could produce fluvial channels on the crater walls.

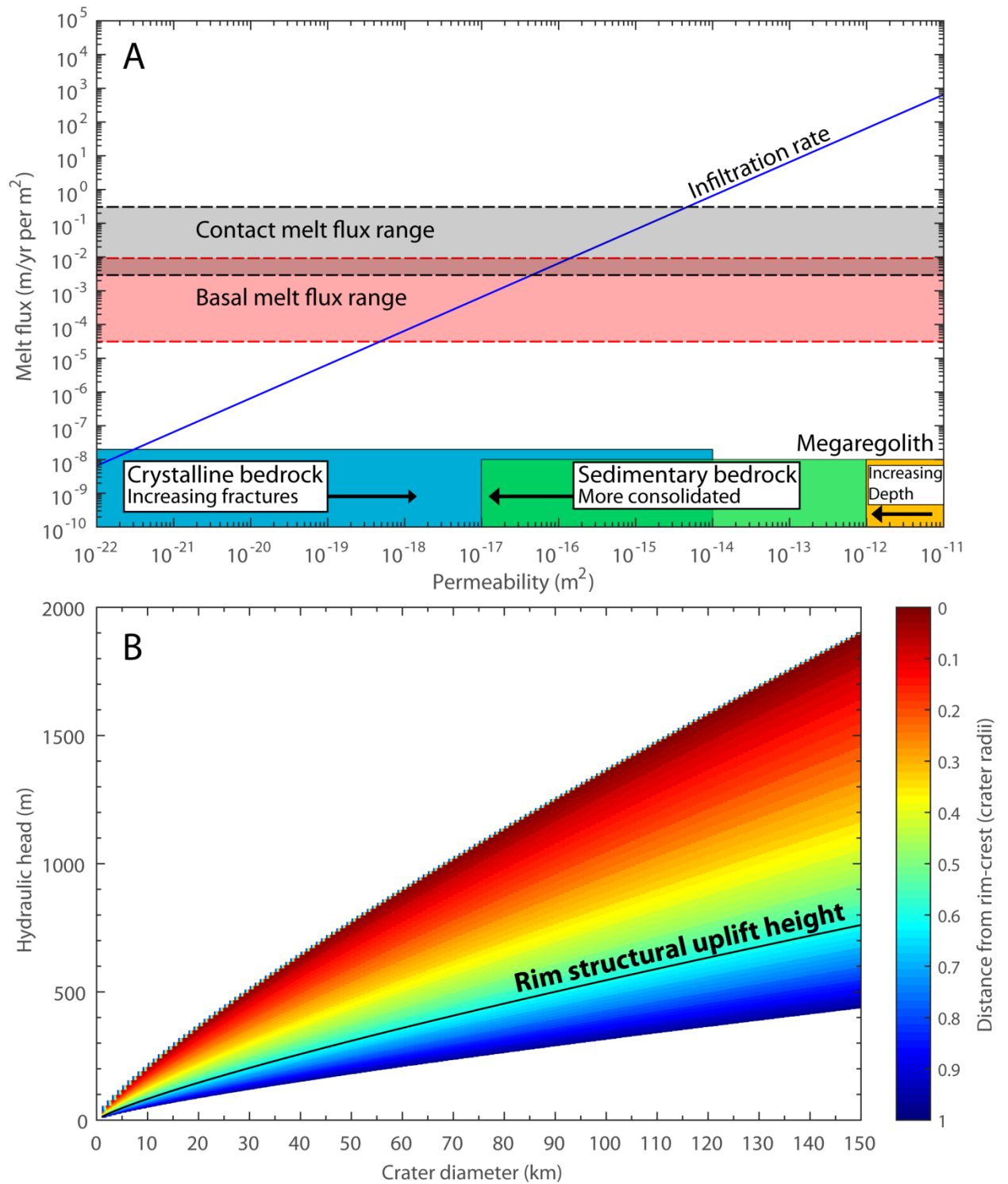


Figure 13. (A) Contact melt production rate (shaded grey area) and basal melt production rate (shaded red area) compared with maximum infiltration rate over a wide range of rock permeabilities (Bear, 1972; Thomson, 1978; Clauser, 1992; Lucia, 1999; Hanna and Phillips, 2005). Crystalline bedrock (blue box; arrow indicates how the permeability increases with increasing fractures), sedimentary bedrock (green box; arrow indicates

how permeability decreases with increasing levels of consolidation), and megaregolith (orange box, arrow indicates how permeability decreases with depth). When the infiltration rate exceeds the basal melt production, all basal melt drains through the underlying substrate. When the basal melt production rate exceeds the infiltration rate, the melt can be transported over the underlying substrate. This occurs for permeabilities $\leq 10^{-18}$ to 10^{-16} m^2 , representative of consolidated sedimentary bedrock or crystalline bedrock. (B) Hydraulic head produced by the weight of the overlying ejecta as a function of crater diameter and distance from the crater-rim crest. Basal melt will be transported up the rim as long as the hydraulic head exceeds the rim structural uplift height (black line shows the average case from the global average rim height function of Robbins and Hynek (2012). All basal meltwater within $0.6R$ from the rim-crest is predicted to be transported up the rim, given sufficiently low substrate permeability (A).

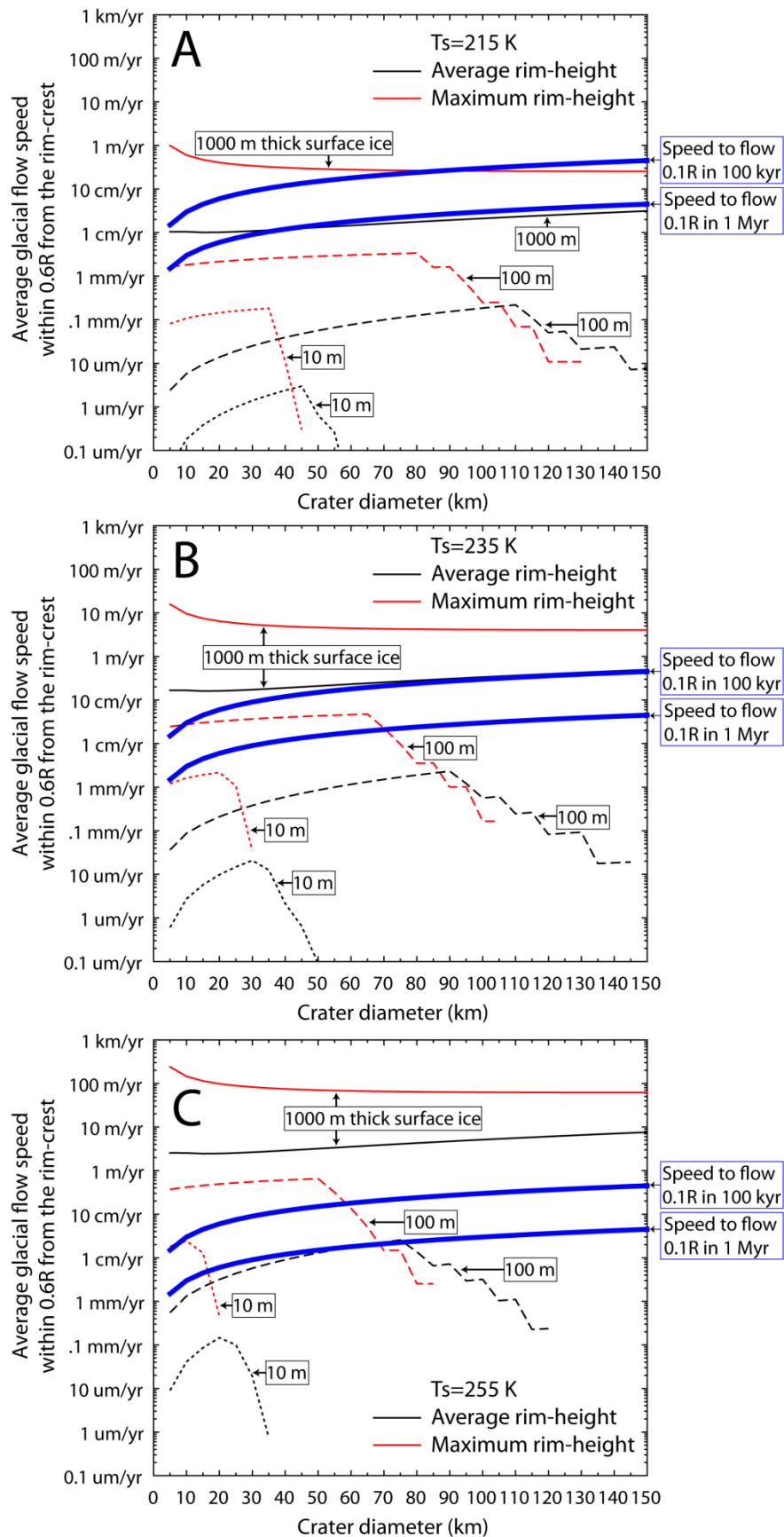


Figure 14. Average glacial flow speeds of the icy layer within 0.6R from the crater rim-crest for (A) surface temperatures of 215 K, (B) 235 K, and (C) 255 K. Model results are shown for the average rim-heights (black lines) and maximum rim-heights (red lines) for surface ice thicknesses of 1000 m (solid lines), 100 m (dashed lines), and 10 m (dotted lines). Surface heat flux in these models is 60 mW/m^2 ; these results do not vary significantly from using heat fluxes of 40 or 100 mW/m^2 . The thick blue lines show the flow speeds required for the ice to flow 0.1R within 100 kyr to 1 Myr. These low flow speeds represent the cold-based ice flow prior to basal melting. Ice flow speeds are generally predicted to be extremely low, with minimal movement of the icy layer. Where model lines are not present (e.g., at larger diameters for the 10 and 100 m ice sheets), all ice has been melted within 0.6R from the rim-crest, and so no ice flow is predicted to occur.

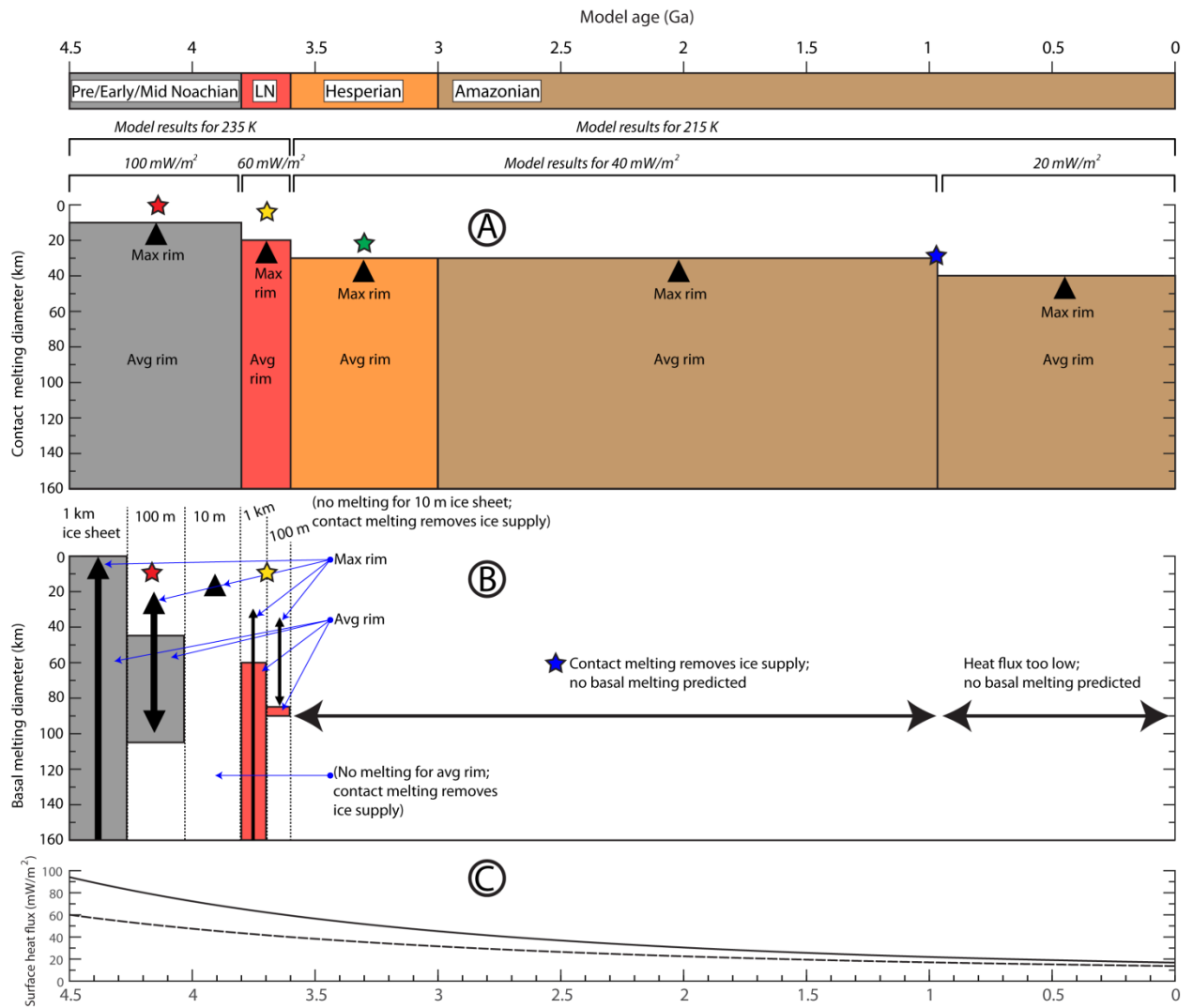


Figure 15. Summary of thermal model results as a function of age. Model ages are from Werner and Tanaka (2011) fit with the Hartmann (2005) production function. An average surface temperature of 235 K is used for the Noachian-age models, and 215 K for the Hesperian and Amazonian models. Pre/Early/Mid Noachian-age models use a heat flux of 100 mW/m², whereas the Late Noachian (LN) age models use a heat flux of 60 mW/m², the Hesperian-age models use a heat flux of 40 mW/m², and the Amazonian-age models use a heat flux of 20-40 mW/m² (Montési and Zuber, 2003; McGovern et al., 2004; Ruiz et al., 2011; Plesa et al., 2015). (A) shows the contact melting diameter range for the average rim-height models (filled boxes) and the maximum rim-height models (black arrows). (B) shows the basal melting diameter range for the average rim-height models (filled boxes) and the maximum rim-height models (black arrows) for ice thicknesses of 10, 100, and 1000 m (C) shows two heat flux models as a function of time from Montési and Zuber (2003). Blue stars refer to contact/basal melting in the Amazonian; green star refers to contact melting in the Hesperian; yellow star refers to contact/basal melting in the late Noachian; red stars refer to contact/basal melting in the Pre/Early/Mid Noachian.

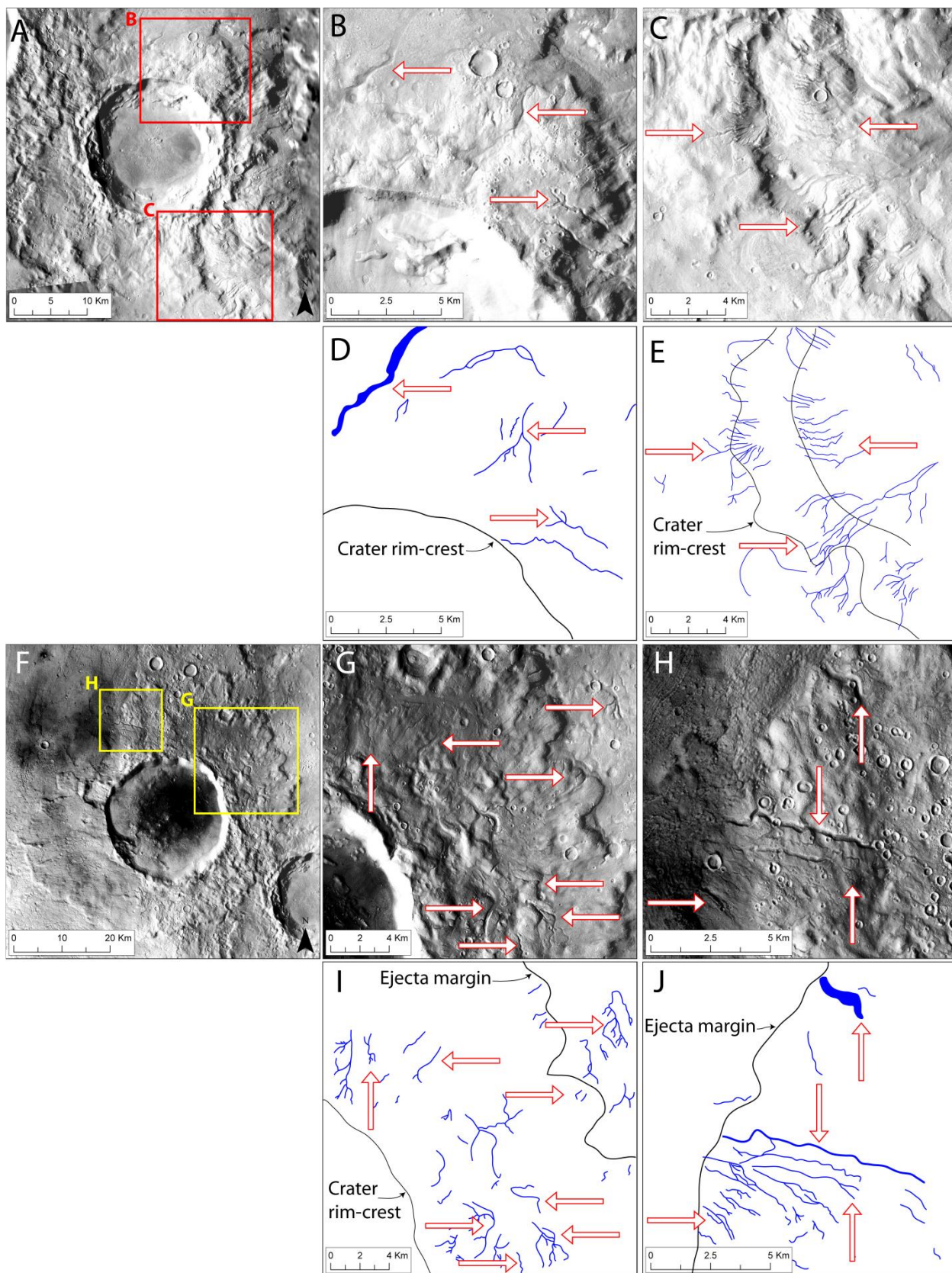


Figure 16. Impact craters which exhibit fluvial channels on the ejecta (red and white arrows), identified by Mangold (2012). These craters are candidate craters for contact melting. (A) 19 km diameter crater (39.8°N, 5.8°E). (B and C) Fluvial channels are located on the ejecta of this crater. (D and E) Sketch map of (B) and (C). CTX image B19_017184_2187. F) 26 km diameter crater (40.1°N, 5.1°E). G and H) Fluvial channels are located on the ejecta and within the surrounding terrain. (I and J) Sketch map of (G) and (H). CTX images P03_002139_2209.

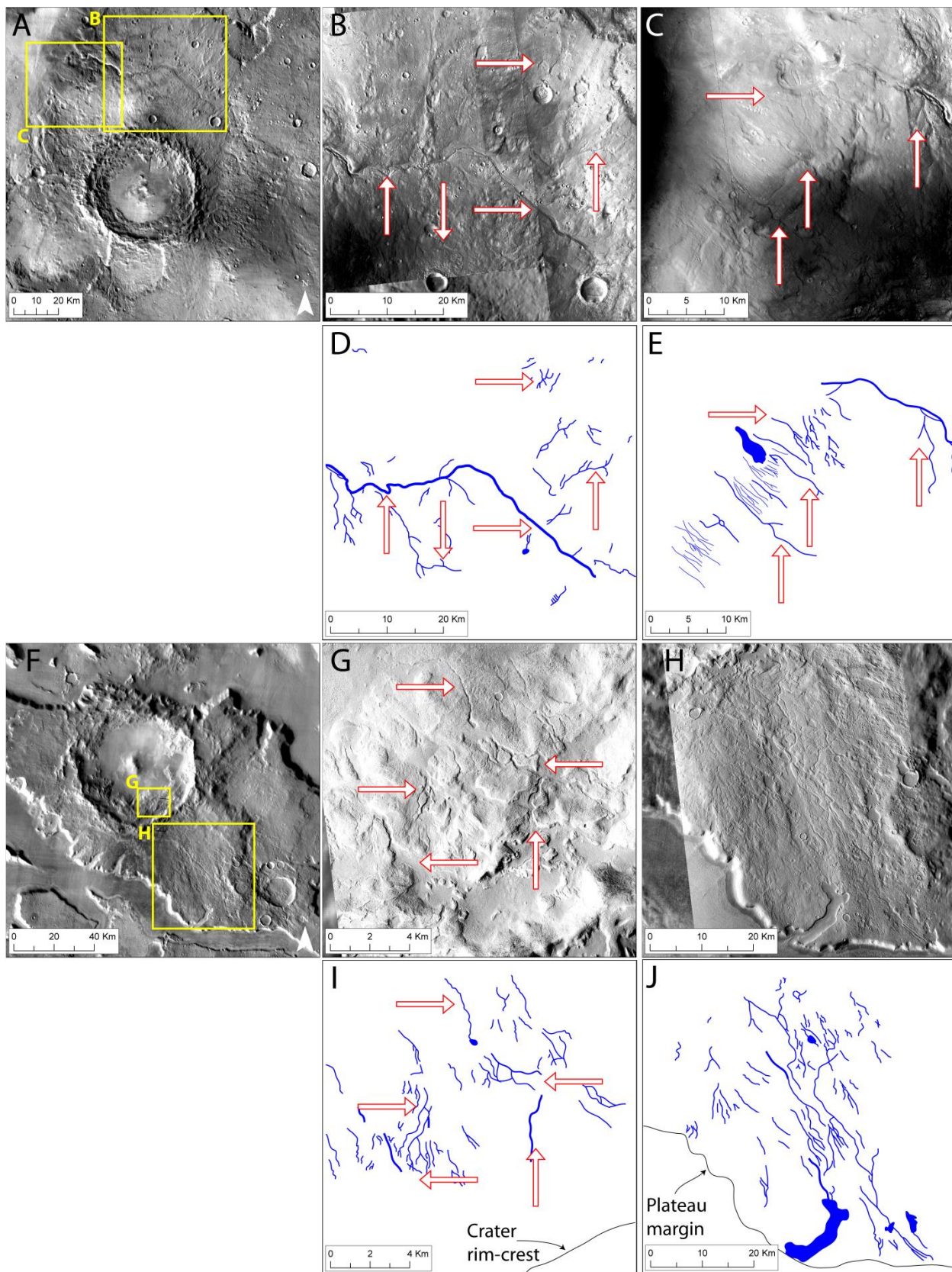


Figure 17. Larger impact craters which exhibit fluvial channels on the ejecta (red and white arrows). These craters are candidate craters for contact melting. (A) 47 km diameter crater (35.6°N, 0.5°E) identified by Mangold (2012). (B and C) Fluvial channels are located on the ejecta and surrounding terrain. (D and E) Sketch map of (B) and (C). THEMIS IR global daytime and CTX images B18_016815_2151, B03_010657_2168, G23_027153_2179, P17_007519_2147, P16_007374_2183. F) 63 km diameter Sinton crater (40.7°N, 31.7°E) characterized by Morgan and Head (2009). (G and H) Fluvial channels are located on the ejecta of this crater. (I and J) Sketch map of (G) and (H). THEMIS IR global daytime and CTX images P20_008968_2191, P03_002138_2194.

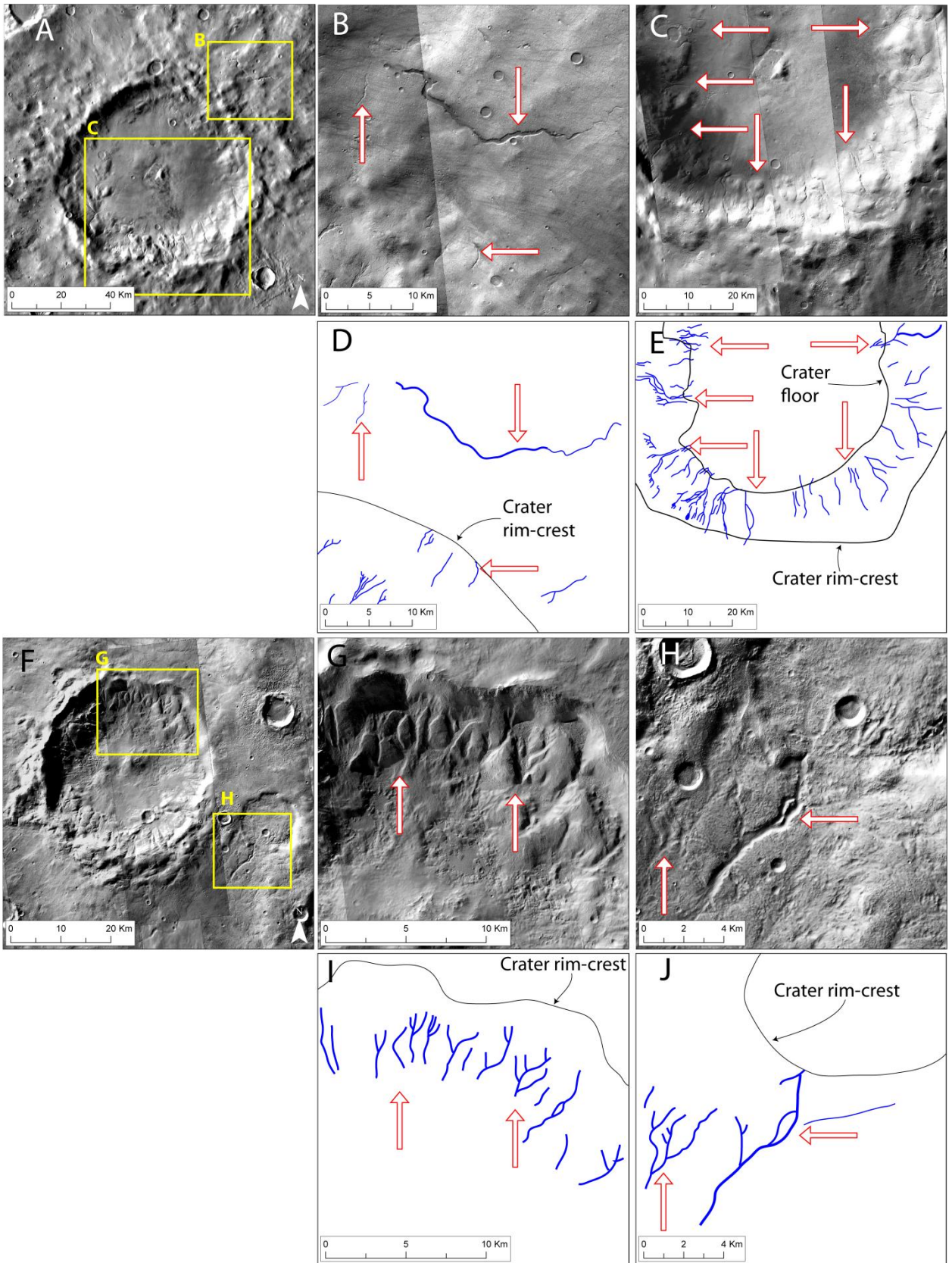


Figure 18. Two examples of Hesperian-aged impact craters characterized by Hobbs et al. (2016) which exhibit fluvial channels on the ejecta and interior walls (red and white arrows). These craters are candidate craters for contact melting. (A) 35 km diameter Choyr crater (32.4°S, 18.6°E). (B and C) Fluvial channels are located on the ejecta and on the crater interior walls. (D and E) Sketch map of (B) and (C). THEMIS IR global daytime and CTX images P16_007400_1414, P15_006965_1419, P16_007255_1411, B17_016340_1402.F) 74 km diameter crater (39.6°S, 19.1°E). (G and H) Fluvial channels are located on the ejecta and interior walls of this crater. (I and J) Sketch map of (G) and (H). In (H), ejecta appears to be infilling a 10 km diameter pre-existing crater, and a fluvial channel is draining out of the smaller crater, consistent with contact melting of ice deposits within the topographic low of the pre-existing crater. CTX images B22_018186_1473, B17_016485_1458.

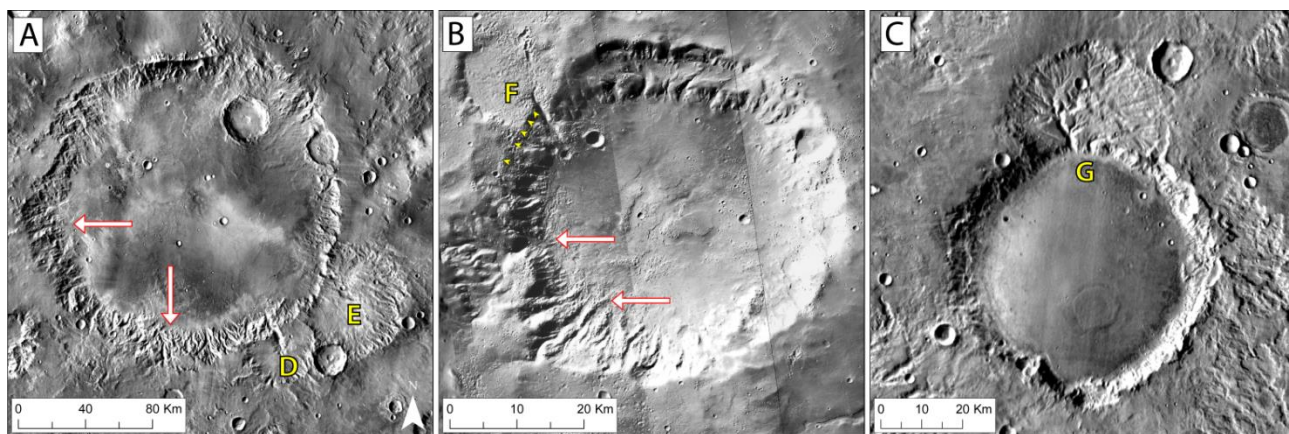


Figure 19. Noachian-aged craters with fluvial features. (A) ~180 km diameter crater adjacent to two pre-existing craters in the lower right hand corner (~40 and ~60 km in diameter) exhibits almost circumferential channels (red and white arrows) superposed on the rim and crater walls (38.1°E, 9.2°S). This crater, if formed on an ice sheet, is predicted to have produced 10^3 - 10^4 km³ of meltwater through contact melting, and a further 10^3 - 10^5 km³ of meltwater through basal melting. The basal meltwater would have migrated to the crater walls, causing runoff and channel formation. (B) ~50 km diameter crater which exhibits almost circumferential channels (red and white arrows) superposed on the rim crest and crater walls (-32.2°E, 29.4°S). This crater, if formed on an ice sheet, is predicted to have produced 10^2 - 10^3 km³ of meltwater through contact melting, and a further 10^1 - 10^4 km³ of meltwater through basal melting. The basal meltwater would have migrated to the crater walls, causing runoff and channel formation. Fluvial incision and meltwater-assisted mass-wasting on steep slopes could have contributed to the formation of the circumferential fluvial channels for the impact craters in both (A) and (B). (C) A 62 km diameter crater (64.7°E, 16.6°S) characterized by Howard (2007). Features D, E, F, and G denote anomalously large channels which superpose older adjacent craters (yellow arrowheads in (B) show that the rim-crest of the large crater cuts into the smaller crater (E), indicating that the smaller crater is stratigraphically older). In (A), (B), and (C), if these craters (features D, E, F, and G) were filled with surface ice at the time of impact, the excess volume of ice available for contact and basal melting could contribute to a more incised channel morphology. (A, C) THEMIS IR global day. (B) CTX images B19_016922_1495, G03_019546_1507, D22_035897_1508 superposed on THEMIS IR global day.

Chapter 4:
**Evidence for stabilization of the ice-cemented cryosphere in earlier
martian history: Implications for the current abundance of
groundwater at depth on Mars**

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Abstract

The present-day martian mean annual surface temperature is well below freezing at all latitudes; this produces a near-surface portion of the crust that is below the freezing point of water for >2 consecutive years (defined as permafrost). This permafrost layer (i.e., the cryosphere) is a few to tens of km thick depending on latitude. Below the base of the permafrost (i.e., the cryosphere), groundwater is stable if it exists, and can increase and decrease in abundance as the freezing isotherm rises and falls. Where water is available, ice fills the pore space within the cryosphere; this region is known as the *ice-cemented cryosphere* (ICC). The potential for a large reservoir of pore ice beneath the surface has been the subject of much discussion: previous studies have demonstrated that the theoretical thickness of the martian cryosphere in the Amazonian period ranges from up to ~9 km at the equator to ~10-22 km at the poles. The total thickness of ice that might fill the pore space within the cryosphere (the ICC), however, remains unknown. A class of martian crater, the Hesperian-Amazonian-aged single-layered ejecta crater, is widely accepted as having formed by impact into an ice-cemented target. Although the target structure related to the larger multiple-layered ejecta craters remains uncertain, they have recently been interpreted to be formed by impact crater excavation below the ice-cemented target, and here we tentatively adopt this interpretation in order to infer the thickness of the ice-cemented cryosphere. Our global examination of the excavation depths of these crater populations points to a Hesperian-Amazonian-aged ice-cemented cryosphere that is ~1.3 km thick at the equator, and ~2.3 km thick at the poles (corresponding to a global equivalent water layer of ~200 m assuming ~20% pore ice at the surface). To explore the implications of this result on the martian climatic and hydrologic evolution, we then assess the surface temperature, atmospheric pressure,

obliquity, and surface heat flux conditions under which the downward-propagating cryosphere freezing front matches the inferred ice-cemented cryosphere. The thermal models which can best reproduce the inferred ice-cemented cryosphere occur for obliquities between 25° and 45° and CO_2 atmospheric pressures ≤ 600 mbar, but require increased heat fluxes and surface temperatures/pressures relative to the Amazonian period. Because the inferred ice-cemented cryosphere is much thinner compared with Amazonian-aged cryosphere thermal models, we suggest that the ice-cemented cryosphere ceased growing when it exhausted the underlying groundwater supply (i.e., ICC stabilization) in a more ancient period in Mars geologic history. Our thermal analysis suggests that this ICC stabilization likely occurred sometime before or at $\sim 3.0\text{-}3.3$ Ga (during or before the Late Hesperian or Early Amazonian period). If groundwater remained below the ICC during the earlier Late Noachian period, our models predict that mean annual surface temperatures during this time were $\geq 212\text{-}227$ K. If the Late Noachian had a pure CO_2 atmosphere, this places a minimum bound on the Late Noachian atmospheric pressure of $\geq 390\text{-}850$ mbar. These models suggest that deep groundwater is not abundant or does not persist in the subsurface of Mars today, and that diffusive loss of ice from the subsurface has been minimal.

1. Introduction

Present-day global martian mean annual surface temperatures (MAST) are well below 273 K at all latitudes (Clancy et al., 2000; Christensen et al., 2001; Smith et al., 2001). In concert with the relatively low martian geothermal heat flux ($\sim 20\text{-}40$ mW/m²) in the Amazonian (the last ~ 3 Ga) (McGovern et al., 2004; Solomon et al., 2005; Plesa et al.,

2016), this yields temperatures below the freezing point of water throughout the shallow martian subsurface. Consequently, water ice is predicted to be thermally stable within the upper kilometers of the subsurface (Fanale, 1976; Clifford, 1993; Mellon et al., 1997; Kuzmin, 2005; Grimm and Painter, 2009; Clifford et al., 2010; Lasue et al., 2013). In the terrestrial literature, the subsurface zone which exhibits temperatures below the freezing point of water for two consecutive years is defined as the permafrost zone (Harrison et al., 1988). In the martian literature, this subsurface zone is referred to as the *cryosphere* (Clifford, 1991; Clifford et al., 2010) (dashed red line in Fig. 1), and we retain this designation here for continuity and clarity. Within the cryosphere (or permafrost), the zone in which ice fills the pore-space is referred to as the *ice-cemented cryosphere* (ICC) (shaded grey region in Fig. 1). Depending on the assumed crustal thermal and diffusive properties, porous ice may persist to considerable depth beneath the local ice table (e.g., Mellon et al., 1997; Grimm et al., 2016), and so we use the term “ice-cemented” but do not imply that the entire pore space within the ICC is necessarily fully saturated with ice. The ICC grows from the bottom-downwards, primarily through either upward thermal vapor diffusion of deeper groundwater, which freezes onto the downward-propagating cryosphere freezing front (Clifford, 1991; 1993); and/or groundwater freezing onto the cryosphere freezing front in places where groundwater is in direct contact with the freezing front (Clifford et al., 2010) (Fig. 1A).

The ICC is distinct from the shallow zone in which pore ice is in diffusive equilibrium with the atmosphere. This shallow zone is characterized by dry regolith which overlies a substrate that may be filled with pore ice that diffuses into the regolith as vapor from the atmosphere (Fanale, 1976; Farmer and Doms, 1979; Fanale et al., 1986;

Clifford and Hillel, 1983; Mellon and Jakosky, 1993; Mellon and Jakosky, 1995; Mellon et al., 1997; Schorghofer and Aharonson, 2005; Head and Marchant, 2014). The thickness of the dry regolith superposing the pore ice is predicted to encompass anywhere from the upper several tens to hundred meters of regolith at the equator, and the upper few centimeters to tens of meters at mid to high latitudes, with actual values determined by the local mean annual surface temperature (which varies as a function of latitude and obliquity), relative humidity of the atmosphere, geothermal gradient, and assumed thermal diffusive properties of the regolith (Fanale, 1976; Farmer and Doms, 1979; Fanale et al., 1986; Clifford and Hillel, 1983; Mellon and Jakosky, 1993, 1995; Mellon et al., 1997; Schorghofer and Aharonson, 2005; Grimm and Painter, 2009; Grimm et al., 2016).

The global ice-cemented cryosphere is the dominant thermodynamic sink for outgassed water and could thus represent a large portion of the water inventory of Mars (Clifford, 1993; Clifford et al., 2010; Lasue et al., 2013; Carr and Head, 2015). Because the pore ice within the cryosphere is sourced by underlying groundwater (Clifford, 1993; Grimm and Painter, 2009; Grimm et al., 2016), defining the thickness of the ICC is critical to the understanding of the aqueous history of the martian subsurface. Two fundamental end-member scenarios exist for the state of the martian cryosphere and groundwater:

Thermally-limited (Fig. 1A and B): The volume of water in the subsurface is approximately equal to the volume of pore space within the crust. In this case, as the planetary heat flux declines and the cryosphere freezing front advances deeper in the martian crust, the ICC grows downwards as it assimilates the underlying groundwater.

The thickness of the ICC in this case depends on the depth of the advancing freezing front.

Supply-limited (Fig. 1C and D): The volume of the water in the subsurface is less than the volume of pore-space within the crust. In this case, as the cryosphere freezing front advances deeper in the crust through time, the ICC will continue to grow until the supply of underlying groundwater is exhausted. The thickness of the ICC depends on the volume of water in the subsurface. At some time, the ICC will reach its maximum thickness and will not grow further as the freezing front advances (hereafter referred to as *ICC stabilization*).

To this end, previous investigators have performed calculations in an effort to constrain the maximum thickness of the cryosphere (Mellon et al., 1997; Clifford et al., 2010). Most recently, Clifford et al. (2010) modeled the Amazonian cryosphere thickness assuming a variety of ice melting isotherms, geothermal heat fluxes, and regolith thermal conductivity configurations, and found cryosphere thicknesses that range from ~10-22 km at the poles, and up to ~9 km at the equator, depending on a wide range of parameters. Clifford et al. (2010) found that the equatorial cryosphere can disappear entirely under special circumstances, for example: if the subsurface is saturated in groundwater that is a eutectic solution of magnesium perchlorate ($\text{Mg}(\text{ClO}_4)_2$), which depresses the ice-melting isotherm to 206 K (Chevrier et al., 2009), or in the case of a eutectic solution of sodium chloride (NaCl) (252 K ice-melting isotherm) and a thick thermally insulating regolith layer is present at the equator. While these models are necessary to estimate the thickness of the cryosphere based on thermal constraints, it remains unclear to what depth the cryosphere is actually filled with pore ice.

How deep is the ice-cemented cryosphere on Mars today, and how much of the water inventory of Mars (Lasue et al., 2013; Carr and Head, 2015) does it represent? What insight can the dimensions of the ICC provide on the abundance of martian groundwater? In this study, we provide an estimate of the thickness of the ice-cemented portion of the cryosphere using the excavation depths of impact craters interpreted to penetrate into a target rich in pore ice (Section 2). We then compare the inferred ICC thickness to thermal model predictions, and evaluate how varying the obliquity, atmospheric pressure, and surface heat flux affect the fit between the inferred ICC and the thermal models (Section 3 and 4). In Section 5, we explore the relevant parameter space to evaluate the thermal model parameters (i.e., atmospheric pressure, surface temperature, obliquity, surface heat flux) which provide the best fit to the inferred ICC thickness through time, and discuss implications for the age and climatic conditions under which the ICC could have reached the ice supply limit (Fig. 1C). Next, we evaluate the deviations between the inferred ICC thickness and the thermal models and discuss possible explanations which link surface geologic processes to the inferred configuration of the ICC (Section 6). Finally, we examine the implications of this study on the current and past presence of groundwater on Mars (Section 7).

2. Crater morphology and target structure

Previous investigators (e.g., Kuzmin, 1980; Kuzmin et al., 1988a; 1988b, 2004; Costard, 1989; Barlow and Bradley, 1990; Boyce and Roddy, 1997, 2000; Baratoux, 2002; Barlow, 2005; Barlow and Perez, 2003; Oberbeck, 2009; Weiss and Head, 2014; Jones and Osinski, 2015; Jones, 2015) have proposed that variations in martian impact

crater morphology can be used to constrain the structure of the target in which craters form. In this section, we review these crater morphologies and outline how they may be used to estimate the thickness of the ice-cemented cryosphere, and then present estimates on the volume of the pore ice within the ICC.

2.1. Single-layered ejecta craters

A class of Hesperian-Amazonian-aged martian layered ejecta craters, single-layered ejecta (SLE) craters (Barlow, 2005) (Fig. 2), are interpreted to form exclusively from impacts in the ice-cemented cryosphere (Carr et al., 1977, Mouginis-Mark, 1981; Costard, 1989; Barlow and Bradley, 1990; Barlow, 1994, 2005, 2006; Stewart et al., 2001; Baratoux, 2002; Barlow and Perez, 2003; Reiss et al., 2005, 2006; Oberbeck, 2009, Weiss and Head, 2014; Jones and Osinski, 2015). SLE craters range from ~1.5 to 40 km in diameter (~10 km on average), and are generally present throughout all latitudes, although they increase in frequency towards the equator (Barlow and Perez, 2003; Robbins and Hynek, 2012; Weiss and Head, 2014; Jones and Osinski, 2015). SLE craters typically display one ejecta lobe which extends ~1-1.5 crater radii from the rim crest (Barlow, 2005; Li et al., 2015) and terminates in a distal rampart (Mouginis-Mark and Baloga, 2006). The fluidized nature of SLE crater ejecta (Carr, 1977) and their blocky ramparts (Baratoux et al., 2005) are interpreted to indicate that these craters formed by an impact into an ice-rich target (Carr et al., 1977, Mouginis-Mark, 1981; Costard, 1989; Barlow and Bradley, 1990; Barlow, 1994, 2005, 2006; Stewart et al., 2001; Baratoux, 2002; Barlow and Perez, 2003; Oberbeck, 2009, Weiss and Head, 2014; Jones and Osinski, 2015). Indeed, Kuzmin (1980), Kuzmin et al. (1988a; 1988b, 2004), and Boyce

and Roddy (2000) found that the onset diameter of the martian layered ejecta craters decreases with increasing latitude, and that the ejecta runout distance (relative to the crater diameter) increases with increasing latitude. This is interpreted to indicate that the depth to the ice-table shallows and the ice content in the subsurface increases with increasing latitude, in agreement with predictions from thermal vapor diffusion models (Mellon et al., 1997).

Based on the interpretation that SLE craters are formed in an ice-rich target, previous studies (Baratoux, 2002; Barlow and Perez, 2003; Barlow, 2006; Weiss and Head 2014) have raised the possibility that the diameters of SLE craters may also be controlled by the thickness of the ICC. This hypothesis is supported by the observation that the maximum diameter of SLE craters increases at higher latitudes (Fig. 2C) (Barlow and Perez, 2003; Barlow, 2006; Weiss and Head 2014), and offers a minimum-bound estimate on the thickness of the ICC.

Although it remains unclear how much pore ice in the target is required to form a fluidized ejecta crater, it is important to note that terrestrial debris flows require high levels of pore-saturation (up to tens of wt% water) in order to produce ramparts (e.g., Major and Iverson, 1999; Savage and Iverson, 2003; Ilstad et al., 2004). Ramparts are interpreted to form through kinetic sieving (Middleton, 1970; Savage and Lun, 1988; Pouliquen and Vallance, 1999; Baratoux et al., 2005; Boyce et al., 2010), wherein larger grains are transported to the flow front, resulting in rapid dissipation of pore pressure (Gray and Ancey, 2009). The decrease in pore-pressure at the flow-front increases friction relative to the rest of the flow, causing the flow-front to decelerate (relative to the rest of the flow) and form a rampart (Iverson, 1997). The martian ramparts have also

been proposed to form by interactions with the atmosphere (Schultz, 1992), but this model predicts the ramparts to be dominated by fine-grained ejecta, in conflict with the observation that ramparts are generally composed of larger particles (Baratoux et al., 2005; Mouginis-Mark and Baloga, 2006; Wulf et al., 2013).

2.2. Multiple-layered ejecta craters

Single-layered ejecta craters are interpreted to impact within the ICC, and thus offer minimum-bounds on the thickness of the ICC. Can upper bounds be placed on the thickness of the ICC? Multiple-layered ejecta (MLE) craters (Fig. 2B) range from ~6 to ~80 km in diameter (~22 km on average) and exhibit ejecta which extends ~2 crater radii from the rim-crest (Barlow, 2005; Weiss and Head, 2014; Li et al., 2015). MLE craters are most common $\pm 40^\circ$ of the equator (Fig. 2; Barlow and Perez, 2003; Barlow, 2006; Weiss and Head, 2014), exhibit a highly sinuous ejecta facies consisting of multiple lobes, and display prominent distal ramparts (Barlow, 1994; Mouginis-Mark and Baloga, 2006). MLE craters have been hypothesized to form from (1) impact into a volatile-rich substrate (Carr et al., 1977; Wohletz and Sheridan, 1983; Costard, 1989, Barnouin-Jha et al., 2005; Komatsu et al., 2007; Oberbeck, 2009) and continuum flow of ejecta (Barnouin-Jha et al., 2005; Mouginis-Mark and Baloga, 2006); (2) interactions with the atmosphere (Schultz and Gault, 1979; Schultz, 1992; Barnouin-Jha and Schultz, 1998; Barnouin-Jha et al., 1999a, 1999b); (3) fuel-coolant interactions (Wohletz and Sheridan, 1983); (4) impact into a liquid water/brine-rich target (Barlow and Bradley, 1990; Boyce and Roddy, 1997, 2000; Oberbeck, 2009); (5) increased impact ejection angle resulting from a volatile-rich substrate causing oversteepening of impacting proximal rim ejecta to

form the lobes (Barnouin-Jha et al., 2005); and (6) impact and penetration below the ice-cemented cryosphere resulting in ejection angle variations (Weiss and Head, 2014).

Most of the hypothesized factors in the formation of MLE craters reviewed above are not necessarily mutually exclusive, with the exception of (4) and (6). Both of these models suggest that the class of multiple-layered ejecta (MLE) craters (Fig. 2B) may have formed by impact into an ice-rich target and ejecta excavation within and below the ICC (Fig. 2C) (Barlow and Bradley, 1990; Oberbeck, 2009; Boyce and Roddy, 1997, 2000; Weiss and Head, 2014) on the basis of their near-equatorial concentration, and relatively larger diameters and multiple ejecta facies compared with SLE craters. Barlow and Bradley (1990) and Oberbeck (2009) suggested that the multiple ejecta lobes characteristic of MLE craters are due to excavation beneath the ICC into groundwater. Barlow (2006) later noted, however, that the excavation depths of MLE craters are likely too shallow for them to excavate groundwater. As we will discuss later (Section 4.1), a theory of origin in which MLE craters excavate groundwater would require an Amazonian surface heat flux that is a factor of ~ 2 -7 times higher than currently inferred (e.g., Montési and Zuber, 2003; Ruiz et al., 2011), and we therefore consider this formation mechanism unlikely. Weiss and Head (2014) alternatively suggested that the difference in strength between the ice-cemented regolith/rock and underlying ice-free regolith/rock would produce variations in the ejecta excavation angles (e.g., see Fig. 9 and 10 in Senft and Stewart, 2008) which could contribute to the formation of the multiple layers/lobes. In this model (Weiss and Head, 2014), the geometry of the excavation streamtubes (e.g., Fig. 1 in Croft, 1980) is predicted to cause ejecta from different depths (e.g., derived from both above and below a strength discontinuity

generated by the ICC) to be ballistically emplaced along the entire extent of the ejecta facies (before flow initiates). Because this ejecta was excavated at contrasting ejection angles (and horizontal velocities), multiple lobes may then form during ejecta flow/sliding. The large sizes of MLE craters (relative to SLE craters) also enhances the shock pressures within the ejecta (Weiss and Head, 2016). This produces more meltwater within the ejecta that contains pore ice from the ICC (Stewart et al., 2004). In this scenario the more distal ejecta, which is derived from the upper part of the target which hosts pore ice (i.e., the ICC), exhibits enhanced fluidization and runout distances relative to SLE craters. Critically, the larger sizes and near-equatorial concentration of MLE craters (relative to SLE craters) is consistent with MLE crater excavation beneath the ICC because the thicker ICC predicted at the high latitudes would prevent frequent MLE crater formation (Weiss and Head, 2014). We emphasize that further work is required to better understand the enigmatic formation of MLE craters, but here we adopt the assumption that the formation of MLE craters is related to excavation beneath the ice-cemented portion of the martian crust in order to proceed with our analysis.

In the context of this interpretation, the thickness of the martian ice-cemented cryosphere may be estimated by finding the “*transition diameter*” between SLE and MLE craters. By determining the threshold diameter at which SLE craters cease forming and MLE craters begin forming (i.e., the *transition diameter*), and then using standard crater scaling laws to determine the corresponding excavation depth (i.e., the *transition depth*), we can provide an estimate of the thickness of the ICC. The transition from an SLE to an MLE crater should not begin exactly when the excavation cavity of the crater penetrates through the ICC because the volume of ejecta excavated below the cryosphere would

initially be negligible. Consequently, we predict the *transition depth* to lie between the maximum SLE and minimum MLE crater excavation depths in any given region.

2.3. Crater relationships and the ICC thickness

In Figure 3A, we examine the latitudinal trends in diameter of the SLE and MLE crater population samples from Weiss and Head (2014). This database has since been updated following the classification criteria from Barlow (2015). The database is complete at latitudes above 40° , but includes only the most confident identifications of an SLE or MLE crater at lower latitudes due to their high frequency near the equator (total $N=882$ MLE craters, 2087 SLE craters). We find SLE crater diameters to typically be ~ 10 km at the equator, and increase to ~ 35 km towards the south pole and up to ~ 40 km towards the north pole (Fig. 3A), confirming the observations of previous investigators (Barlow and Bradley, 1990; Barlow and Perez, 2003; Barlow, 2006). Our detailed review of crater morphologies show that there exist numerous examples of confidently classified MLE craters at all latitudes, and that MLE craters are generally larger than SLE craters in each latitudinal band (Fig. 3A). We interpret this to indicate that the larger MLE crater excavation depths provide an upper limit to the ICC thickness. Thus, the ICC thickness estimates derived from this method are not considered lower bounds.

Because there is a lower frequency of MLE craters at high-latitudes, we also examine the radial (lunar-like) ejecta craters poleward of 40° . The craters we examine are from the Barlow (1988) crater database, but newer images (THEMIS and CTX data) were used to refine several classifications and we thus omitted a small number of the craters ($N=14$). We co-plot the remaining radial ejecta craters poleward of 40° ($N=12$) in Fig. 3A (only

nine radial craters are shown in the figure because three of the radial craters are larger than 100 km in diameter). On the basis of their large sizes and lunar-like (non-fluidized) ejecta morphology, this crater class is interpreted to have excavated in a target that is largely free of water/ice (Barlow and Bradley, 1990). Considering that these craters are generally between ~60-100 km in diameter at the high latitudes (black triangles in Fig. 3A), they are predicted to excavate ejecta from depths between ~4.2 km and 6.5 km. The ejecta is likely to be volatile-poor, either because groundwater is not present at these depths, or alternatively because the porosity at such great depths is too low for sufficient pore ice to fluidize the ejecta. We find the porosity argument difficult to explain this observation because the porosity at 4.2 km should be between ~7 to 13% (for an initial porosity of 0.20 to 0.35), and the porosity at 6.5 km would be between ~4-8% (using Eq. 1). Furthermore, the large diameters (and shock pressures; e.g., Fig. 4 in Weiss and Head, 2016) of these craters imply that they are melting a larger proportion of their pore ice relative to the smaller craters, and so it remains uncertain whether the lower porosity actually corresponds to lower volumes of meltwater. While it remains unclear how much water is actually needed to fluidize ejecta, it is important to note that most of the excavated volume of ejecta in a near-paraboloidal excavation cavity (Croft, 1980) is derived from shallower depths where the porosity (and thus the ice content) is higher than the lower limits discussed above, and where the distal ejecta (i.e., the ejecta diagnostic of fluidization) is derived from. In concert, these points suggest that the radial ejecta craters are not excavating groundwater, and so we proceed with the interpretation that groundwater was unlikely to have been in contact with the ice-cemented cryosphere when these craters formed. Consequently, we consider these craters to be absolute upper

bounds on the depth of the ICC.

In order to find the zonally averaged *transition depth* on Mars, we sort the SLE/MLE crater populations into an equal-area grid on the martian surface. We use latitude bins of 15° , and longitude bins of 15° at the equator. In order to maintain bins of equivalent surface area, the longitudinal bin size progressively increases with latitude to account for decreasing area with latitude. For example, the longitudinal bin sizes increase from 15° between 0° - 15° latitude, up to 60° longitude in the 75° - 90° latitude bin. Next, we find the *maximum* SLE crater diameter and *minimum* MLE crater diameter in each latitude/longitude bin, and then find the zonal average of these two crater diameters at each latitude interval. We find the transition diameter by averaging these maximum and minimum values within each latitude bin (green squares in Fig. 3A). The large bin sizes presented here minimize error from regions with a low frequency of SLE or MLE craters, although we note that varying the bin dimensions does not drastically alter our results. For example, Fig. 3C shows that the transition diameters derived using a variety of different bin dimensions are not significantly different in magnitude and form to those using the equal-area bins described above (green squares; Fig. 3B).

We find the excavation depth (D_E) of these impact craters as $D_E = 0.1D_T$ (Croft, 1980; Melosh, 1989), where $D_T = D_{SC}^{0.15 \pm 0.4} D_R^{0.85 \pm 0.04}$ (Croft, 1985). D_T is the transient crater diameter, D_{SC} is the simple-complex crater transition diameter (global average is ~ 6 km on Mars; Robbins and Hynek, 2012), and D_R is the rim-to-rim crater diameter. Based on these scaling relations, the martian crater latitude-depth relationships (Fig. 3B) are interpreted to represent the presence of a Hesperian-Amazonian (the age of the SLE/MLE craters; e.g., Reiss et al., 2006) equatorial ICC thickness of ~ 1.3 km that thickens to a

maximum of ~ 2.3 km towards the poles (Fig. 3B). The ICC thickness estimates presented here are based on 15° latitude bins and 15 - 60° equal-area longitude bins (Fig. 3B), and thus represent a zonally averaged estimate. While regional variations in geothermal heat flux and crustal thermal properties (e.g., thermal conductivity) would affect the cryosphere thickness locally (e.g., Reiss et al., 2005, 2006; Cassanelli and Head, 2015, 2016; Cassanelli et al., 2015; Weiss and Head, 2016), these effects are damped out in our estimate due to the zonal-averaging method used. Interestingly, Baratoux et al. (2002) applied dimensional analysis to the sinuosity of impact ejecta to 250 SLE craters within $\sim 15^\circ$ of the equator and found that the trends between sinuosity and crater diameter could be explained by impact into a target of low viscosity in the upper ~ 1 km, which overlies material of higher viscosity. Baratoux et al. (2002) pointed out that this could be related to a rheologic transition between an upper zone saturated in pore-ice above a zone free of pore-ice, or due to declining porosity with depth. This result is in good agreement with the finding of a ~ 1.3 km thick ice-cemented cryosphere at the equator inferred in our study on the basis of SLE/MLE crater excavation depths.

Because the surface temperature in radiative equilibrium (and the thickness of the cryosphere) varies with the cosine of latitude (e.g., Pierrehumbert, 2010), the latitude-dependent distribution of the *transition diameter* between SLE and MLE craters (green squares in Fig. 3A) is highly suggestive of a cryosphere control: the formation of larger SLE/MLE craters at high latitudes is consistent with impact into a thicker ICC, and the relatively smaller SLE/MLE craters near the equator are consistent with impact into a relatively thinner ICC. The frequency of SLE and MLE craters is lower at higher latitudes, which may limit confidence in the observed latitudinal trend. We note,

however, that the error bars shown in Fig. 3 account for the sample size in each latitudinal bin. If the lower-end ICC thickness estimate is adopted from the error bars, a latitude-dependence is still observed, and so we consider the latitude-dependence shown in Fig. 3 to be a reasonable basis for further analysis. If the interpretation that MLE craters excavate through the ICC is incorrect (e.g. if MLE craters instead formed due declining porosity with depth), the derived ICC thicknesses would not be applicable, but in that case MLE crater diameters and excavation depths would not be expected to show any latitude-dependence, which is not the case (Fig. 3B). Furthermore, if the ICC extended to deeper depths than MLE crater excavation depths (and MLE craters were not formed by impacts which excavate through the ICC), it would remain unclear how radial ejecta craters, interpreted to form in a largely water/ice-free target, excavated only ~1-2 km deeper than MLE craters (black triangles in Fig. 3A) in the same latitudinal bands. Consequently, we consider our estimate of the thickness of the martian ICC to provide a reasonable basis for further analysis.

2.4. Pore volume in the ice-cemented cryosphere

How much ice is contained within the ICC? We calculate the total pore volume of the ICC (Table 1) inferred from SLE/MLE crater excavation depths by integrating the volume of the pore-space down to the depth of the ICC in each latitudinal band (Fig. 3B) on a spherical Mars. We exclude the upper ~300 m of crust equatorward of $\pm 40^\circ$ interpreted to be depleted of volatiles (Kuzmin, 1980; Kuzmin et al., 1988a, 2004; Clifford, 1993; Mellon et al., 1997; Boyce and Roddy, 2000; Kirchoff and Grimm, 2016). We use the porosity (Φ) profile from Athy's law (Athy, 1930):

$$\Phi_{(Z)} = \Phi_0 \exp\left(\frac{-Z}{K}\right) \quad (1)$$

where Φ_0 is the porosity at the surface, and Z is depth in km. Clifford (1993) adjusted the lunar porosity decay constant ($K_{Lunar}=6.5$ km) to martian gravity (g), which yielded a K value of 2.82 km. New results from the GRAIL mission suggest a lunar $K_{Lunar}=9.8$ km (Besserer et al., 2014), which, when adjusted for martian gravity ($K_{Mars} = K_{Lunar} \frac{g_{Lunar}}{g_{Mars}}$), yields a value of 4.28 km (Weiss and Head, 2017). This results in an ICC volume of 3.21×10^7 km³, equivalent to a martian global equivalent water layer (GEL) of 203 m ($\Phi_0=0.2$; Table 1).

Despite the higher crustal porosity predicted by the updated decay constant, our estimates of the volume of ice within the cryosphere (~ 200 m GEL) are lower than previous estimates of the volume of ice that may be available within the deep cryosphere (435-1025 m for a melting isotherm of 273 K; Clifford et al., 2010). Similarly, Carr and Head (2015) recently provided an estimate of the surface/near-surface reservoir of water on Mars to be 24 m GEL in the Hesperian period, in contrast to earlier, higher values.

2.5. Age of the ice-cemented cryosphere

The layered ejecta craters are believed to be Hesperian through Amazonian in age on the basis of (1) their superposition over Hesperian-and Amazonian-aged terrains (Barlow and Bradley, 1990; Barlow and Perez, 2003; Jones and Osinski, 2015) (Fig. 4A); (2) inferred moderate erosional state (Reiss et al., 2005); and (3) the dating of individual layered ejecta craters (e.g., Reiss et al., 2006; Mouginis-Mark and Boyce, 2012; Sun and

Milliken, 2014; Werner et al., 2014; Viola et al., 2015; Wulf and Kenkmann, 2015; Kirchoff and Grimm, 2016). As pointed out by Reiss et al. (2006), because SLE and MLE craters are Hesperian through Amazonian in age, it is possible that the ICC thickness inferred in this study is simply a snapshot from an earlier period in martian history (e.g., the Hesperian). If the bulk of SLE and MLE craters used in this study formed in the Hesperian (during a period of higher geothermal heat flux than the present) for example, their excavation depths would record a relatively thinner ICC (Fig. 1A). After this period, however, groundwater present below the ICC would have continued to assimilate onto the deepening cryosphere and thicken the ICC (Fig. 1B). If this is the case, the ICC inferred in this study would not reflect the present-day ICC thickness on Mars. Could the inferred ICC thickness reflect a snapshot from a changing cryosphere thickness through time?

In order to address this question, we examine the distribution of SLE and MLE craters on different aged surfaces from the updated geologic map of Mars (Tanaka et al., 2014a). SLE and MLE craters are found to superpose terrains which span from the Amazonian through the Noachian in age (Fig. 4A), which places minimum bounds on crater ages: Craters forming on Hesperian terrains could be younger (Amazonian) in age, but they cannot be older (i.e., Noachian). Note that none of these craters are likely to be Noachian in age based on their degradation state (Mangold et al., 2012), and so the SLE and MLE craters present on Noachian-aged terrains are likely Hesperian or Amazonian in age. The latitudes and excavation depths of SLE and MLE craters present in Amazonian-aged terrains are shown in Fig. 4B; terrains which may be either Amazonian or Hesperian (Fig. 4C); Hesperian-aged terrains (Fig. 4D); and Noachian or Hesperian-aged terrains

(Fig. 4E). If the ICC recorded by SLE and MLE craters (Fig. 3B) has thickened through time, the excavation depth transition between SLE (red triangles) and MLE craters (blue squares) is also expected to increase through time in Fig. 2E.

The SLE and MLE craters present on Amazonian-aged terrains (Fig. 4B) are fewest in number, likely because Amazonian units comprise only 10% of the surface area of Mars as mapped by Tanaka et al. (2014a, 2014b). Based on the overlap between SLE and MLE craters, this population appears to record an ICC that is between ~0.8-1.5 km thick between 20°N and 40°N, which encompasses the ICC thickness predicted by the entire SLE/MLE populations at the same latitude (~1.3 km thick; Fig. 3B). More SLE and MLE craters are present on terrains denoted as Amazonian/Hesperian and Hesperian by Tanaka et al. (2014a), which may be due to an older age for the craters (these units comprise 9% of the surface area of Mars; Tanaka et al., 2014b). These craters appear to record an ICC that is also between ~0.8-~1.5 km thick $\pm 40^\circ$ of the equator, and ~2.5 km thick at the high latitudes (Fig. 4C), consistent with the global trends shown in Fig. 3B. Craters located on exclusively Hesperian-aged terrain are also abundant, and suggest an ICC thickness of ~1 km $\pm 40^\circ$ of the equator; this unit comprises 27% of the surface area of Mars (Tanaka et al., 2014b). We have grouped Noachian-aged terrain (44% of the surface area of Mars; Tanaka et al., 2014b) and Hesperian/Noachian-aged terrain (10% of the surface area of Mars; Tanaka et al., 2014b) in Fig. 3E. The craters within these units appear to record an ICC that is ~1 km thick at the equator and up to ~2.5 km thick in the high southern latitudes, consistent with the global trends shown in Fig. 3B.

If the ICC thickness recorded by SLE and MLE craters (Fig. 3B and C) has increased through time, the excavation depth transition between SLE and MLE craters present on

Noachian- and Hesperian-aged terrains (Fig. 4D and E) is expected to be shallower than those present on Amazonian-aged terrains (Fig. 4B and C). This does not appear to be the case: SLE/MLE crater excavation depths present on younger terrains are not deeper than those on older terrains. The SLE/MLE transition excavation depth in the mid- and low-latitudes remains a constant ~ 1.3 km regardless of terrain-age. It appears from this data that the SLE/MLE craters in this study are sampling an ICC which has not observably thickened during the Amazonian and Hesperian periods. These observations may indicate that the SLE/MLE craters used in this study are either primarily Amazonian in age, or if many are Hesperian in age, then the ICC stopped thickening at some time during or before the Hesperian period. In either case, the craters used to determine the ICC thickness appear to have impacted into the ICC after it reached the supply limit of ice and stopped thickening through time (Fig. 1D). This is consistent with the observation (Barlow, 2004) that craters of varying degradation (a proxy for time) do not exhibit any changes in ejecta runout distance (a proxy for fluidization by shock-induced melting of pore ice): Barlow (2004) interpreted these data to indicate that the volatile-content of the subsurface has remained relatively constant since the end of the Noachian period.

In summary, we used the transition between the excavation depths of SLE and MLE craters to estimate the ICC to be ~ 1.3 km thick at the equator, and up to ~ 2.3 km thick toward the poles (corresponding to a ~ 200 m GEL layer). These ICC thickness estimates are consistent with the prediction of a latitude-dependent cryosphere thickness (e.g., Clifford et al., 2010). Based on terrain-age and excavation depth relationships (Fig. 4), we suggest that these craters largely formed after the ICC stopped growing.

If indeed the SLE/MLE craters formed in the ICC after it stopped growing, it raises

the possibility that the ICC was supply-limited (i.e., the supply of deep groundwater was exhausted as the ICC grew). For example, the thickness of the cryosphere (i.e., the depth of the ice melting isotherm) increases with time as the planetary heat flux declines (Fig. 1). In the supply-limited scenario (Fig. 1C and D), the downward-propagating freezing front of the cryosphere may have reached the base of the ICC (i.e., the ICC assimilates all underlying groundwater and stops growing; Soderblom and Wenner, 1978; *ICC stabilization*, Fig. 1D) prior to the Amazonian period.

We acknowledge that a hydrologic model of Mars with a supply-limited cryosphere is seemingly incompatible with an origin for the outflow channels involving groundwater discharge from a globally integrated, pressurized groundwater system (e.g., Clifford, 1993; Fig. 6 in Carr, 2002; Fig. 1 in Harrison and Grimm, 2009), but we proceed in our analysis with the assumption that outflow channels may not be fundamentally linked to globally integrated subsurface groundwater aquifers. We discuss this potential inconsistency in Section 7, and proceed in our analysis. Is the hypothesis of a supply-limited ICC consistent with thermal constraints? Next, we model the thickness of the martian cryosphere (following Clifford et al., 2010) for comparison with the inferred ICC configuration (Fig. 3B) in order to evaluate the possibility of a supply-limited ICC.

3. Cryosphere thermal models

Could the ICC have stabilized during an earlier period in the history of Mars? Under what obliquity, geothermal heat flux, atmospheric pressure, and global mean annual surface temperature (MAST) conditions can the ICC stabilize? In order to address these questions, we produce thermal models (following the approach of Clifford et al., 2010) of

Amazonian-age through Late Noachian-age cryosphere thicknesses for comparison with the inferred ICC thickness derived from the excavation depths of SLE/MLE craters (Fig. 3B). Because the thickness of the ICC is dependent upon MAST and geothermal heat flux, a comparison between the inferred ICC thickness and thermal model predictions offers a way to investigate ancient martian conditions. In order to assess the relationship between the thermal model parameters and the thickness of the inferred ICC, we illustrate how surface heat flux, obliquity, and atmospheric pressure can affect the thickness of the cryosphere, and how large changes to these parameters affect the fit between the thermal models and the inferred thickness of the ice-cemented cryosphere.

3.1. Thermal profile

We find the depth of the cryosphere using the one-dimensional steady state heat equation:

$$T_{(Z)} = T_{(Z-1)} + \frac{Q\Delta Z}{\kappa_{(Z)}} \quad (2)$$

where $T_{(Z)}$ is temperature as a function of depth (Z), where the surface temperature $T_s = T_{(Z=0)}$ and Q is the geothermal heat flux (in W/m^2); we use a ΔZ of 1 meter. The depth of the cryosphere is defined where $T_{(Z)}$ reaches the melting point of ice. We adopt the thermal conductivity structure of the upper martian crust from Clifford (1993) and Clifford et al. (2010), given by (Hobbs, 1974):

$$\kappa_Z = \frac{488.19}{T_{(Z)}} + 0.4685 \quad (3)$$

Clifford (1993) noted that the κ of basalt spans the range of κ for terrestrial permafrost, and that the κ for ice (Eq. 3) (Hobbs, 1974) is generally equal to that of basalt. Thus, a basaltic bedrock or megaregolith substrate saturated with pore ice is also predicted to share this thermal conductivity. Following Clifford et al. (2010), we adopt Eq. (3) for the thermal conductivity of the substrate rock within the cryosphere.

Due to desiccation of the shallow regolith at the low latitudes, the shallow equatorial zone is predicted to be devoid of pore ice (Clifford and Hillel, 1983; Clifford et al., 1993; Mellon et al., 1997; Grimm and Painter, 2009; Grimm et al., 2016). On the basis of Fanale et al. (1986), Kuzmin (1980), Kuzmin et al. (1988a, 2004), Boyce and Roddy (2000), Clifford et al. (2010), and Kirchoff and Grimm (2016), we set the depth of the ice-free regolith to 0.1 m at $>40^\circ$ latitude, 1 m at 40° , 200 m at 20° , and 300 m at the equator. This differs slightly from Clifford et al. (2010), who used a 180 m thick equatorial desiccated zone. We explore the case of a desiccated equatorial zone of thermal conductivity $\kappa_{eq}=1$ W/mK (i.e., consolidated ice-free sedimentary/volcanic rock), 0.1 W/mK (unconsolidated rock), and for the simple case of no equatorial desiccated zone.

3.2. Mean annual surface temperatures (MAST)

We use martian mean annual surface temperatures for $T_s = T_{(z=0)}$ in Eq. (2). In order to explore cryosphere thickness through time, we implement Amazonian and Late Noachian surface temperature conditions. Our thermal models adopt the present-day Amazonian MAST climate model results from Haberle et al. (2003) for obliquities of 0° ,

15°, 30°, 45°, 60° (Fig. 5A). For the Late Noachian MAST, we use results from recent 3D Late Noachian (solar luminosity at 3.8 Ga) general circulation models (GCMs) (Horan and Head, 2016), which include a pure CO₂ atmosphere, eccentricity of 0, and a water cycle (the Laboratoire de Météorologie Dynamique (LMD) GCM from Forget et al., 2013 and Wordsworth et al., 2013, 2015). We explore obliquities of 25°, 35°, 45°, and 55° and surface pressures of 125 mbar (Fig. 5B), 400 mbar (Fig. 5C), 600 mbar (Fig. 5D), 800 mbar (Fig. 5E), and 1000 mbar (Fig. 5F). The obliquity range used in this study falls within that suggested by the statistical solutions of Laskar et al. (2004), which predicted that the average obliquity of Mars over its entire history is 37.62° with a standard deviation of 13.82°. Note that as atmospheric pressure increases in the Late Noachian models, the lapse-rate strengthens and the effects of topography on temperature become more pronounced, leading to lower temperatures in the southern highlands for increasing atmospheric pressures (Fig. 5B-F). A zonally averaged pole-to-pole MOLA topographic profile (5° latitude bins) is shown in Fig. 5G for comparison.

3.3. *Ice melting isotherm*

In order to define the base of the ICC in the thermal models, we must determine the ice-melting isotherm (for pure ice this is 273.15 K). For example, Fig. 6 reproduces the Amazonian cryosphere thickness estimates of Clifford et al. (2010) for a variety of ice-melting isotherms and surface heat fluxes. The lower ice melting isotherms (206 and 252 K) explored by Clifford et al. (2010) illustrate the case where a salty eutectic groundwater solution is in direct contact with the cryosphere freezing front, and freezes directly onto the base. The 206 K isotherm (Mg(ClO₄)₂ brine) is a poor choice because it

cannot produce an equatorial ICC (blue lines in Fig. 6).

As noted in Clifford (1993), a eutectic solution is a natural consequence of the cryosphere freezing front advancing through time. As groundwater is progressively cold-trapped to the cryosphere, the salts are concentrated in the underlying groundwater, depressing the freezing point. This concept has led to the adoption of eutectic freezing points throughout the literature. We note, however, that the salt concentration through time from this process is highly dependent on the depth of the freezing front. We consider it unlikely to have caused groundwater in the upper kilometers of the martian subsurface (where the base of the inferred ice-cemented cryosphere is in this study) to be a eutectic solution based on the following lines of reasoning.

Based on the inferred ICC thickness in our study, freezing the upper ~1.3-2.3 km of groundwater in a ~10 km thick water column using the porosity profile from Eq. (1) is equivalent to freezing ~28% of the groundwater in the subsurface (assuming a thermally-limited groundwater system from Fig. 1A and B, a 10 km pore closure depth from Hanna and Phillips, 2005, accounting for the density difference between water and ice, and using volumes of the ICC and ice-free pore space below the ICC from Table 1). Therefore, if the entire column of water started with 5 wt% salt before it was concentrated by freezing, freezing the upper regions within the ice-cemented cryosphere would lead the groundwater below the ice-cemented cryosphere to have a salt content of 7%, a scenario in which the groundwater isotherm would be only slightly lower (~1-6 K) than 273 K (Table 2). In order to achieve the eutectic solution (Chevrier et al., 2009), the initial salt content of the global groundwater inventory before concentration by freezing must be unreasonably large (Table 2): for example, 17 wt% for NaCl, or 32 wt% for magnesium

perchlorate. For comparison, terrestrial seawater hosts ~3.5 wt% salts, and terrestrial briny groundwater is typically composed of ≤ 10 wt% salts (Van Weert et al., 2009).

The eutectic solution is attainable if the cryosphere freezing front advanced to a much greater (deeper) depth than the thickness of the ice-cemented cryosphere inferred in our study. For example, if 80% of the volume of groundwater has been frozen in a fully saturated subsurface (with pore closure at 10 km), only ~3-10 wt% initial (pre-freezing) salt is required to reach a eutectic solution. This scenario is not realized in our models because the inferred thickness of the ice-cemented cryosphere only reaches depths of ~1.3-2.3 km, which is only ~30% of the available pore space above 10 km. The supply-limited scenario thus predicts that groundwater was not in contact with the ICC.

In summary, even if the groundwater had up to 5-10 wt% salt, the freezing point would only be depressed between ~1-6 K (Table 2), which would lead the ice-cemented cryosphere to be only ~30-200 m deeper than the 273 K isotherm (Eq. 1). We therefore consider the 273 K isotherm to be the most reasonable because the depth of the melting isotherm for 5-10 wt% salts is not quantitatively or qualitatively different than for the 273 K isotherm. Furthermore, the radial ejecta craters, which are unlikely to form in a groundwater-rich target, are excavating even deeper than MLE craters (Fig. 3A), which, in tandem with our volume calculations above, suggests that direct contact between groundwater and the ICC is unlikely (in which case the cryosphere grows through vapor diffusion, and the 273 K isotherm is valid). For these reasons, we proceed in our thermal model analysis favoring the 273 K (pure ice) melting isotherm. To be thorough, we also explore models using the 252 K isotherm as reference point in order to explore the case of a highly depressed freezing point, which may be valid locally or regionally (but not

globally) in areas of perched aquifers. The 252 K isotherm represents the eutectic for an NaCl solution (23.3 wt% salt), or a solution of Mg perchlorate with ~32 wt% salt, or Na perchlorate with ~37 wt% salt (Table 2). Notably, the 252 K isotherm is also representative of a model where the melting isotherm remains 273 K, but the thermal conductivity of the upper martian crust is approximately half of that given in Eq. (3), corresponding to the case where a large portion of the pore space within a porous megaregolith comprising the ICC is devoid of pore ice.

4. Cryosphere model results

We now evaluate the thermal model fits to the inferred ICC by varying surface heat flux, obliquity and atmospheric pressure. We attempt to isolate the parameters which are able to reproduce the form and magnitude of the inferred ICC in order to understand better the climatic conditions at the time when the ICC stopped growing.

4.1. Amazonian cryosphere models

The Amazonian cryosphere thickness estimates of Clifford et al. (2010) are reproduced in Fig. 6 under a variety of different Amazonian geothermal heat flows (15 and 30 mW/m²; McGovern et al., 2004; Solomon et al., 2005) and ice melting isotherms (206 K; eutectic Mg(ClO₄)₂ brine, 252 K; eutectic NaCl brine, and 273 K; pure ice; Clifford et al., 2010). We find that the ICC is anomalously thin (~1.3-2.3 km) compared with the cryosphere thicknesses predicted by Amazonian thermal models (Fig. 6) (typically ~3-22 km; Clifford, 1993; Mellon et al., 1997; Clifford et al., 2010). The models predict either an excess cryosphere thickness (~5-14 km) at high latitudes (252

and 273 K isotherms) or an absence of an equatorial cryosphere (206 K isotherm), irrespective of heat flow conditions. One difference between the model shown in Fig. 6 and that of Clifford et al. (2010) is that we do not include a hydrate-rich cryosphere. For simplicity, we do not consider the case of a global subsurface methane hydrate layer due to the lack of globally distributed methane detections: previous investigators (Formisano et al., 2004; Mumma et al., 2009; Webster et al., 2015) attribute the origin of the methane to localized sources, and it remains unclear whether methane hydrate is generating the methane.

Because the obliquity of Mars varies on a 10^5 - 10^6 yr timescale (Laskar et al., 2004), we first explore the effects of varying obliquity on the thickness of the Amazonian cryosphere (which can respond to the 10^6 yr variations; Grimm and Painter, 2009; Clifford et al., 2010; Grimm et al., 2016). Using these models we find the R^2 values (a measure of the correlation between the datasets) (Fig. 7A), root mean squared error (RMSE; Fig. 7B), and sum of squares error (SSE) of the thermal models (Fig. 7C) over a wide range of surface heat fluxes. We present the corresponding least squares fit between the thermal models and the ICC thickness in Fig. 7D (Table 3). The model results shown in Fig. 7 illustrate the case where $\kappa_{eq}=1$ W/mK using the 273 K isotherm model.

Our model results show that the R^2 values exhibit near-normal distributions around a range of surface heat fluxes for each obliquity model (Fig. 7A). It appears that the 30° obliquity (near the present day value of 25.2°) and 45° obliquity models offer the best fit to the inferred ICC thickness ($R^2=0.80, 0.87$), but the surface heat flux is required to be ~ 100 mW/m², which is a factor of ~ 2.5 -7 too large for the Amazonian period (e.g., Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005; Ruiz et al.,

2011). These relationships (Fig. 13A) also apply to the 252 K isotherm model (Fig. 13C), but for lower surface heat fluxes of $\sim 80 \text{ mW/m}^2$ (a factor of $\sim 2\text{-}5$ too large). Thus, if MLE craters excavated groundwater-rich crust, the Amazonian heat flux is required to be elevated to unrealistic levels. A surprising finding is that the inferred ICC thickness is far thinner than predicted by the Amazonian thermal models, regardless of the obliquity: surface heat fluxes are required to be vastly in excess of typical Amazonian heat flux estimates in order for the thermal models to reproduce the ICC thickness.

The disparity between the thin inferred ICC and the thick ICC predicted by Amazonian thermal models (Fig. 6) could have important implications for the water inventory and geologic history of Mars. The difference between the inferred and modeled ICC thickness suggests that the maximum modeled cryosphere thickness (Fig. 6) (Clifford, 1993; Mellon et al., 1997; Clifford et al., 2010) was not reached in the Amazonian due to a supply limit of ice (i.e., the volume of the pore space in the cryosphere exceeded the volume of ice available to fill the pores; Fig. 1D). Because the ICC thickness appears to be anomalously thin compared with the modeled Amazonian cryosphere thickness, we raise the possibility that the cryosphere freezing front reached the maximum thickness of the ICC (and the supply-limit of ice) during an earlier period in martian history (Fig. 1C).

Mars is predicted to have had a thicker atmosphere during the more ancient Noachian period (e.g., Kasting, 1991; Haberle, 1998; Forget et al., 2013; Wordsworth et al., 2013, 2015; Kite et al., 2014; Hu et al., 2015). Could a thicker atmosphere on ancient Mars allow the thermal models to better reproduce the ICC thickness? Next, we examine the effects of increasing the atmospheric pressure on the thermal models.

4.2. Late Noachian cryosphere models

Does changing the atmospheric pressure allow the thermal models to better reproduce the inferred ICC thickness? In order to assess this, we evaluate surface temperatures/pressures predicted for the more ancient Late Noachian martian climate (Fig. 5B-F). The model results shown in Fig. 8-12 illustrate the case where $\kappa_{eq}=1$ W/mK using the 273 K isotherm model. Much like for the Amazonian models, the R^2 values appear to exhibit near-normal distributions around a range of surface heat fluxes for each atmospheric pressure and obliquity model (Fig. 8A-12A). For the 125 mbar atmosphere, the 25° and 35° obliquity models (black and blue lines in Fig. 8) offer the best fit to the ICC, and provide R^2 values >0.8 for heat fluxes of 105 and 107 mW/m². Similarly, for the 400 mbar atmosphere, the 25° and 35° obliquity models (black and blue lines in Fig. 9) offer the best fit to the ICC, and provide R^2 values >0.8 for heat fluxes of 81 and 82 mW/m². For the 600 mbar atmosphere, the 25° and 35° obliquity models (black and blue lines in Fig. 10) also offer the best fit to the ICC, and provide R^2 values >0.69 for heat fluxes of 70 and 73 mW/m². The 800 mbar atmosphere provides poorer fits: the 35° and 45° obliquity models (green and blue lines in Fig. 11) offer the best fit to the ICC but provide R^2 values >0.4 for heat fluxes of 63 and 66 mW/m². The 1000 mbar atmosphere provides the worst fits (Fig. 12), with all R^2 values approaching zero. These relationships (Fig. 13A) also apply to the 252 K isotherm model (Fig. 13C), but for comparatively lower surface heat fluxes (~60-80% the heat flux values of the 273 K isotherm model). Table 3 summarizes the parameters and statistics of the best-fitting cryosphere thermal models for $\kappa_{eq}=1$ W/mK.

In a manner similar to the Amazonian models, the Late Noachian models between 125 and 600 mbar provide good fits to the inferred ICC data. Figure 13 shows each of the best-fitting thermal models displayed as an individual marker for a given atmospheric pressure and obliquity. The higher surface temperatures provided by the increased atmospheric pressure reduces the heat flux requirements of the Late Noachian models to reproduce the magnitude of the inferred ICC compared with the Amazonian models (Fig. 13A and B). The decreased freezing point of the 252 K isotherm models compared with the 273 K isotherm models also serves to reduce the heat flux requirements of these models to reproduce the ICC (Fig. 13C and D). The model results for $\kappa_{eq}=0.1$ W/mK and the case of no desiccated equatorial zone are co-plotted with the nominal model ($\kappa_{eq}=1$ W/mK) results in Fig. 13A-D. The models where $\kappa_{eq}=0.1$ W/mK eliminate the equatorial cryosphere entirely, providing a poor fit, and so all R^2 values are zero in this case. Figure 13E and F and Table 3 show that the best correlating models are for atmospheric pressures ≤ 600 mbar and obliquities between 25° and 45° , and that the 273 K isotherm models typically have higher R^2 values and lower SSE and RMSE than the 252 K isotherm models. Interestingly, the highest frequency of the peak R^2 values for the 273 K isotherm model at a given atmospheric pressure is at 35° obliquity, a result comparable to the time-averaged martian obliquity of 37.62° predicted by Laskar et al. (2004).

None of the surface heat fluxes which produce the least squares fits in Fig. 13 are representative of the Amazonian period, which further suggests that the cryosphere stabilized in a more ancient period of martian history. Based on the R^2 values, RMSE, and SSE of the different models (Fig. 13; Table 3) we suggest that when the ICC stabilized, atmospheric pressures were likely to have been $\leq \sim 600$ mbar and obliquity

was likely between 25° and 45°. These models, however, represent only a snapshot in time, atmospheric pressure, and obliquity conditions. The cryosphere freezing front may reach the base of the ICC over any range of atmospheric pressures and obliquities. For example, in order for two different thermal models to achieve identical cryosphere thicknesses (i.e., the same depth of the ice melting isotherm), a model with lower surface pressure (or higher κ) must have a higher surface heat flux. In the following section, we use the results of these thermal models to assess the ICC stabilization parameter range as a function of time.

5. Some speculations on the ice-cemented cryosphere through time

The best-fit model analysis (Section 4) offers the opportunity to explore MAST and heat flux as a function of time. In this section, we first use the least square fit thermal models (Fig. 13) to constrain the surface temperature and heat flow conditions at the time when the cryosphere freezing front reached the base of the ICC (Section 5.1 and 5.2). Further, because vapor diffusion timescales (Clifford and Hillel, 1983) are much shorter than geothermal heat flux decay timescales (Montési and Zuber, 2003) (i.e., as the planetary heat flux declines, the ICC can concomitantly grow through vapor diffusion), we can then speculate on the age during which the subsurface ice-supply was reached by the ICC (i.e., when all groundwater is assimilated into the overlying ICC) and the ICC stops growing (*ICC stabilization*) (Section 5.3).

The global MAST, atmospheric pressure, and heat flux of the best-fit cryosphere thermal models (Fig. 13) can be fit by linear functions, as shown in Fig. 13A-D. For the nominal case of $\kappa_{eq}=1$ W/mK, the best-fit global MAST (T_F) and atmospheric pressure

(P_F) can be related to the best fit heat flux (Q_F) by:

$$T_{F(273)} = -612.545Q_F + 263.914 \quad (4)$$

$$P_{F(273)} = -18.427Q_F + 1.985 \quad (5)$$

$$T_{F(252)} = -603.0437Q_F + 247.742 \quad (6)$$

$$P_{F(252)} = -18.273Q_F + 1.506 \quad (7)$$

These functions represent the best-fit global MAST, atmospheric pressure and heat flux required for the ICC to stabilize for both the 273 K isotherm model (Eqs. 4 and 5) and the 252 K isotherm model (Eqs. 6 and 7). The atmospheric pressures are for a CO₂ atmosphere with a water cycle in the LMD GCMs of Forget et al. (2013), Wordsworth et al. (2013, 2015), and Horan and Head (2016). The ancient martian atmospheric composition is not yet known, and individual climate models generate somewhat different surface temperatures under the same atmospheric pressure (e.g., Mischna et al., 2013; Wordsworth et al., 2013; Urata and Toon, 2013) due to differing physics parameterizations. The thickness of the cryosphere, however, is fundamentally a function of geothermal heat flux and surface temperature. Thus, the MAST- Q_F relationship (Eqs. 4 and 6) in Fig. 13A is largely independent of the different assumptions and parameters within individual climate models.

Using these function (Eqs. 4 and 6), we can estimate the MAST required for the ICC to stabilize over a range of heat fluxes. In order to link MAST from Eqs. (4) and (6) to the heat flux as a function of time from the martian interior, we set Q_F in Eqs. (4) and (6) equal to the surface heat flux from the heat balance models of Montési and Zuber (2003)

(red and blue lines in Fig. 14) and Ruiz et al. (2011) (black line in Fig. 14). These heat balance models (Fig. 14) have been shown to be consistent with surface heat fluxes derived from lithospheric elastic thickness measurements (McGovern et al., 2004; Solomon et al., 2005; Ruiz et al., 2011) and wrinkle ridge mechanical models (Montési and Zuber, 2003). We refer to the upper end heat flux estimate from Montési and Zuber (2003) as MZ1, the lower end heat flux estimate as MZ2, and the heat flux estimate from Ruiz et al. (2011) (which uses a Urey ratio of 1) as RUr1.

Solving Eqs. (4) and (6) with Q_F equal to the MZ1, MZ2, and RUr1 heat flux functions predicts the MAST and heat flux requirements through time which allow ICC stabilization (Fig. 15). Figure 15 thus shows the minimum MAST required for the ICC to stabilize at any given time (higher MAST would allow groundwater below the ICC). As time progresses and the internal heat of the planet declines, MAST is required to increase to compensate for the decreasing heat flux in order to preserve the depth of the cryosphere freezing front. In other words, the slope of the lines in Fig. 15 do not indicate that surface temperatures increase through time, but rather that if the cryosphere freezing front reached the base of the ICC at 3 Ga rather than 3.5 Ga, for example, higher surface temperatures at 3 Ga are needed to compensate for the lower heat flux.

5.1. Minimum Late Noachian temperatures

In this section, we use the MAST- Q_F relationship from Eqs. (4) and (6) to provide estimates on the mean annual surface temperatures on ancient Mars. We first review the physical and geologic constraints that are relevant to the analysis, and then determine the lower limits of the MAST in the Late Noachian period.

The outflow channels (Tanaka, 1986) are predominantly Hesperian in age and are believed to form through groundwater discharge from beneath the ICC (e.g., Baker and Milton, 1974; Carr, 1979, 1996, 2002; Clifford, 1993; Clifford and Parker, 2001; Head et al., 2003; Manga, 2004; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007; Cassanelli et al., 2015). If this interpretation is correct, the ICC seems unlikely to have stabilized prior to the beginning of the Hesperian period (Late Noachian-Early Hesperian boundary is ~ 3.6 Ga; Hartmann, 2005; Werner and Tanaka, 2011; Michael, 2013). We thus rule out the MAST and heat flow configurations for ICC stabilization prior to 3.6 Ga in Fig. 15 (grey shading), but we note that this assumption would require the outflow channels to be sourced by perched and highly compartmentalized aquifers (e.g., Harrison and Grimm, 2009) in order to maintain pressurization in a supply-limited ICC. In order to exclude unrealistically low or high surface heat fluxes through time, we exclude all heat flux values greater than MZ1 and lower than MZ2 (grey shading in Fig. 15) from Montési and Zuber (2003) (red and blue lines; Fig. 15).

Taking into account the two conditions outlined above, we are left with a more confined range of MAST and heat flow configurations in which the cryosphere freezing front could have reached the ICC between 3.6 and 0 Ga (white and yellow-shaded areas in Fig. 15). The predicted minimum MAST at the end of the Late Noachian (3.6 Ga) for the 273 K isotherm model is 227 K (Fig. 15A), corresponding to a surface heat flux of $\leq 60 \text{ mW/m}^2$ (MZ1 high heat flow) (Table 4). For the 252 K isotherm model, the minimum MAST at 3.6 Ga is 212 K (Fig. 15B). Any MAST less than 212-227 K at 3.6 Ga would allow the ICC to stabilize prior to 3.6 Ga, and may thus be unlikely based on the presence of outflow channels, which are interpreted to result from groundwater discharge from

beneath the ICC. The lower heat flux estimates predict relatively higher minimum MAST: the RUr1 heat flux estimate (black line in Fig. 15) predicts a minimum MAST of 233 K at 3.6 Ga for the 273 K isotherm model (Fig. 15A), and 224 K for the 252 K isotherm model (Fig. 15B). The MZ1 low heat flow model predicts the minimum MAST at 3.6 Ga to be 238 K for the 273 K isotherm model (Fig. 15A), and 231 K for the 252 K isotherm model (Fig. 15B). If the atmosphere was pure CO₂, the equivalent minimum atmospheric pressures in the LMD GCMs (Forget et al., 2013; Wordsworth et al., 2013, 2015; Scanlon et al., 2013; 2016; Horan and Head, 2016) are 850 mbar for the 273 K isotherm model and 390 mbar for the 252 K isotherm model (for MZ1 heat flux) (Table 4), after accounting for increasing solar luminosity through time (~30% in 4.5 Gyr; Gough, 1981). The 252 K isotherm model is also representative of a model with the 273 K isotherm but a crustal thermal conductivity of approximately half of the value used in Eq. (3), corresponding to the case where a large portion of the pore space within the ICC is devoid of pore ice.

In summary, if we assume that the ICC did not stabilize before the Late Noachian (so that the outflow channels can form through groundwater discharge in the Hesperian), the minimum mean annual surface temperature in the Late Noachian predicted by our models is 212-227 K. In a pure CO₂ atmosphere with a water cycle (i.e., the LMD GCM; Forget et al., 2013; Wordsworth et al., 2013, 2015), this corresponds to a minimum Late Noachian atmospheric pressure of 390-850 mbar.

5.2. Comparison with previous paleopressure estimates

Because our lower limit atmospheric pressure estimates at 3.6 Ga (minimum of 390-

850 mbar CO₂ atmosphere) are based on the LMD general circulation model of Forget et al. (2013) and Wordsworth et al. (2013, 2015), they are inherently climate model-dependent. Despite the uncertainty of the presence of additional greenhouse gases (e.g., Ramirez et al., 2013; Halevy and Head, 2014; Horan and Head, 2016), our results appear to be consistent with previous bounds on the martian paleoatmospheric pressure in the Noachian: (1) the ≥ 120 mbar surface atmospheric pressure inferred from the terminal velocity of a volcanic bomb sag at Gusev crater (Manga et al., 2012); (2) the 0.5-2.0 bar Noachian atmospheric pressure range inferred from chemical equilibrium thermodynamics for rocks exposed in Gusev Crater (van Berk et al., 2012) ; (3) the 0.5-5.0 bar Noachian atmospheric pressure range inferred from the carbonate content of martian dusts and soils (Lammer et al., 2013); (4) the ~ 0.2 -2.7 bar range of early Mars atmospheric pressures predicted by 3D general circulation models to be stable against atmospheric collapse (Forget et al., 2013); (5) the upper bound Late Noachian atmospheric pressure of < 2 bars which can match orographic precipitation patterns (Scanlon et al., 2013); (6) the upper limit atmospheric pressure estimate of 0.9 ± 0.1 bar at 3.6 Ga by Kite et al. (2014) on the basis of atmospheric filtering of impactors; (7) the suggestion that the martian atmosphere may have had $\lesssim 500$ mbar of CO₂ during the Late Noachian on the basis of the spectrally-derived carbonate contents within a Noachian-aged rock unit (Edwards and Ehlmann, 2015); (8) the upper limit atmospheric pressure estimate of ~ 1 bar at 3.8 Ga indicated by the modern day carbon isotope ratios in the martian atmosphere and rocks/soil (Hu et al., 2015); and (9) the estimated range of 0.25-2 bar Noachian atmosphere based on models for impact-induced atmospheric escape and volatile delivery (Pham and Karatekin, 2016).

5.3. Cryosphere stabilization age

When during martian geologic history did the ICC exhaust the underlying groundwater supply and stop growing (i.e., ICC stabilization)? Because the decay of planetary heat flux (Montési and Zuber, 2003) occurs over longer timescales than vapor diffusion (Clifford and Hillel, 1983), the rate at which the ICC can grow is limited by the rate in which the geothermal heat flux declines. Thus, by placing an upper bound on either MAST or atmospheric pressure at the time during or before ICC stabilization, we may estimate the latest time period in which the ICC can stabilize. We first review a recently published upper bound placed on atmospheric pressure, and then discuss implications for the age of ICC stabilization.

Kite et al. (2014) compared the size-frequency distribution of small craters in Aeolis Dorsa to predictions of atmospheric impactor-filtering and found that the maximum atmospheric pressure at 3.6 Ga was 0.9 ± 0.1 bar. Hu et al. (2015) modeled the evolution through time of carbon reservoirs and atmospheric escape on Mars and found that the modern day carbon isotope ratios suggest that the atmospheric pressure at 3.8 Ga was likely less than ~ 1 bar. Although the ancient atmospheric composition remains unknown, the results of Kite et al. (2014) and Hu et al. (2015) allow us to make predictions about the age of ICC stabilization. Because atmospheric pressure is predicted to have declined through time (e.g., Lammer et al., 2013; Hu et al., 2015), atmospheric pressures > 1 bar after 3.6 Ga are unlikely. If we assume that the ancient martian atmospheric composition after 3.6 Ga was CO_2 (e.g., Forget et al., 2013; Wordsworth et al., 2013, 2015) and no more than 1 bar (Kite et al., 2014; Hu et al., 2015), the area of “unrealistic solutions”

(defined by the shaded grey regions) grows to encompass the shaded yellow area in Fig. 15. This shaded yellow region corresponds to MAST greater than or equal to a 1 bar CO₂ atmosphere; the temperature of the 1 bar CO₂ atmosphere increases with time due to the increasing solar luminosity (Gough, 1981). The latest age at which ICC stabilization is predicted to occur is thus 3.3 Ga for the MZ1 heat flux (intersection of red line and shaded yellow region in Fig. 15A) in the 273 K isotherm model. Because the 252 K isotherm model (which is also representative of a model with the 273 K isotherm but a crustal thermal conductivity of approximately half of the value used in Eq. 3) reduces the heat flux required for the thermal models to match the inferred ICC, the area of realistic solutions in this case occurs at temperatures lower than produced for the 1 bar CO₂ atmosphere, and so the atmospheric pressure does not offer any constraint on the stabilization age. We note, however, that for the ICC to avoid stabilization by 3 Ga, MAST is required to be > 220 K (corresponding to CO₂ atmospheric pressures > 600 mbar at 3 Ga in the LMD GCM). For the ICC to avoid stabilization by 2 Ga, MAST is required to be $\geq$ 230 K, and $\geq$ 240 K to avoid ICC stabilization by 1 Ga. Given that Mars is believed to experience modern-day, cold conditions (modern day MAST=210 K) for the duration of the Amazonian period (e.g., Carr and Head, 2010), it seems unlikely that the 252 K isotherm model would allow ICC stabilization beyond the beginning of the Amazonian period, at 3.24 Ga (age from Michael, 2013).

We note that these estimates assume that the martian atmospheric composition at the time of cryosphere stabilization was pure CO₂. The addition of a greenhouse gas (or a grey gas) would change the relationship between atmospheric pressure and MAST, which would change the linear function in Fig. 13B and D (Eqs. 5 and 7) and thus the estimated

ICC stabilization age. Given that the Hesperian period is believed to have been characterized by an Amazonian-like climate without a substantial greenhouse effect (e.g., Bibring et al., 2006; Carr and Head, 2010), however, we suggest that the nominal estimate for the latest ICC stabilization age of ~ 3.0 to ~ 3.3 Ga remains reasonable.

In summary, previous estimates on the Late Noachian atmospheric pressure (Kite et al., 2014; Hu et al., 2015) in concert with the results of thermal models (Fig. 13B) allow us to provide an estimate on the latest age of ICC stabilization of ~ 3.0 to ~ 3.3 Ga.

5.4. Summary of thermal model results

Our analysis (Fig. 13 and 15) shows that the depth of the cryosphere freezing front could have plausibly reached the base of the ICC (and the ice volume supply limit) in a more ancient period in the history of Mars (Fig. 1C), when heat fluxes, and possibly atmospheric pressure, MAST, and obliquity, were higher. On the basis of the varying degrees of correlation among model runs with different atmospheric pressure and obliquity, (Fig. 13) our models indicate that when the ICC stabilized, atmospheric pressures were likely to be $\leq \sim 600$ mbar and obliquity was likely to be between 25° and 45° (Section 4.2).

Our MAST- Q_F ICC stabilization model (Fig. 15) may further constrain Late Noachian (>3.6 Ga) atmospheric temperatures. If we assume that the ICC did not stabilize before 3.6 Ga (so that groundwater may persist into the Hesperian to form outflow channels), Late Noachian temperatures at 3.6 Ga are constrained to ≥ 212 - 227 K assuming surface heat flows ≤ 60 mW/m² (Fig. 17). If the Late Noachian atmosphere was pure CO₂, the corresponding atmospheric pressure at 3.6 Ga is required to be ≥ 390 -

850 mbar. This value appears to be consistent with estimates from previous researchers (Section 5.2).

Assuming a pure CO₂ atmosphere (from Forget et al., 2013 and Wordsworth et al., 2013, 2015) at the time of ICC stabilization, our models (Fig. 15) predict that the stabilization of the ice-cemented cryosphere will occur within the Amazonian or Hesperian period (~3.0-3.3 Ga at the latest; Fig. 17). It is difficult to envision ICC stabilization later than ~3.0 to 3.3 Ga (the beginning of the Amazonian period; Michael, 2013), given that this would require MAST in excess of 231 K (273 K isotherm model) or 218 K (252 K isotherm model) (Table 4) in the cold and dry Amazonian period (Section 5.3). For frame of reference, the modern-day global mean annual surface temperature is ~210 K. Because the modern-day sun is ~29% brighter than at 3.3 Ga (Gough, 1981), the MAST at 3.3 Ga with the modern-day 6 mbar CO₂ atmosphere would yield a MAST of only ~199 K, and so mean annual surface temperatures would be required to be elevated by ~20-30 K in the Amazonian period for the ~10⁶ year timescales required for the thermal wave to penetrate to the base of the ice-cemented cryosphere. In summary, the Late Noachian atmospheric pressure is required to be ≥ 390-800 mbar to avoid ICC stabilization before 3.6 Ga, but the martian atmospheric pressure was likely < 600 mbar when ICC stabilization did occur (sometime at or before ~3.0 to 3.3 Ga).

6. Deviation between thermal models and the ICC

In this section, we evaluate the major disparity between the inferred ICC and the results of the thermal models and discuss a possible explanation which links surface

geologic processes to the inferred configuration of the ICC. It appears that the Amazonian-aged crater excavation depths decrease sharply at 75°S (Fig. 3A), suggesting a shallower ICC at the southernmost high latitudes. Critically, this feature (dashed red circle in Fig. 7D) is unable to be reproduced by any of the thermal models.

We note that a shallow ICC at the southern high-latitudes could result from the thermally insulating effect of a polar ice cap. As pointed out by Clifford (1993) and Cassanelli and Head (2016), the insulating effects of a kilometers-thick ice sheet would elevate the ice-melting isotherm and thin the underlying cryosphere (Fig. 16). Although the current south polar cap extends contiguously to only 85°S, the more ancient expanded southern-polar cap, the Dorsa Argentea Formation (DAF), is mapped extending down to ~65°S (Tanaka and Scott, 1987, Head and Pratt, 2001; Tanaka and Kolb, 2001; Tanaka et al., 2014a), but may have been much larger (Scanlon et al., 2016). For comparison, the northern polar cap currently extends down to 80°N (Fig. 16) (Zuber et al., 1998), and does not appear to be reflected in the inferred ICC thickness (Fig. 3B) because it is present at latitudes higher than the SLE and MLE craters used in our study (Fig. 3A).

The DAF is characterized by eskers interpreted to result from basal melting of the DAF ice sheet at the Late Noachian-Early Hesperian boundary (Head and Pratt, 2001; Fastook et al., 2012; Scanlon and Head, 2014; Kress and Head, 2015; Butcher et al., 2016). The suggestion that basal melting formed the eskers under the Dorsa Argentea Formation (Head and Pratt, 2001; Fastook et al., 2012; Scanlon and Head, 2014; Kress and Head, 2015) requires that the underlying ice-cemented cryosphere was melted first.

The best-fit thermal models (Fig. 7-12) predict the southern hemisphere cryosphere at 75°S to be 2.3-2.7 km thick, in contrast to the $\sim 1.5 \pm 0.3$ km thickness inferred. The

deviation between the cryosphere model thickness and the inferred ICC data (dashed red circle in Fig. 7) could be explained by 0.5 to 1.5 km thick snow and ice deposits (i.e., the DAF) present on the surface within this latitudinal band at a time period during or before ICC stabilization. We note that after the surface temperature and/or heat flux reduced sufficiently to terminate melting of the ICC below the DAF, any leftover deep groundwater could have diffused upwards and thickened the ICC below the DAF, and so this thickness estimate of the DAF (1 ± 0.5 km) is a minimum estimate. Interestingly, our DAF thickness estimate is in agreement with the average $\sim 1.4 \pm 0.7$ km height of tuyas present within the DAF (Ghatan and Head, 2002). Tuyas are volcanic edifices that erupt subglacially, and their height is interpreted to record the thickness of the ice at the time of eruption (e.g., Jakobsson and Gudmundsson, 2008). We suggest that the close correspondence of the measured tuya heights within the DAF ($\sim 1.4 \pm 0.7$ km) to our thermal model deviation at 75°S (1 ± 0.5 km) is highly suggestive of the signal from DAF melting and thinning the ICC during the Noachian-Hesperian.

In summary, it appears that the inferred ICC is anomalously shallow at the high southern latitudes, which may be a remnant from an expanded south-polar ice cap, the DAF, during a more ancient climate regime on Mars. This hypothesis is consistent with the results of our thermal modeling (Section 5), which independently suggests that the ICC stabilized during or shortly after the presence of the DAF (Fig. 17).

7. Implications for groundwater

In this section, we review the implications of our cryosphere thermal models for the martian groundwater inventory through time. We first review the expected behavior of

groundwater with respect to a growing ice-cemented cryosphere (Section 7.1). Then, using observations from geomorphology, numerical modeling, and radar sounding, we evaluate whether groundwater was in direct contact with the cryosphere (Section 7.2). We next assess whether our observations are consistent with outflow channel formation through groundwater discharge (Section 7.3), and finally we discuss the implications of our cryosphere thermal models for the martian groundwater inventory (Section 7.4).

7.1. Interaction between the ICC and groundwater

A globally integrated groundwater system, wherein groundwater can migrate down subsurface topographic gradients across the planet, has been proposed by Clifford (1993) and Clifford and Parker (2001) on the basis of several working assumptions: (1) an upper few kilometers of crust that is both permeable and porous; (2) a cryosphere saturated with pore ice; and (3) high heat flow and low crustal thermal conductivity (to permit the stability of liquid water above the pore closure depth). In this model, as the cryosphere freezing front advances downwards through time, groundwater can freeze onto the cryosphere where in direct contact with the cryosphere, or may instead diffuse upwards as vapor through the vadose zone (Figure 1A). In either case, ice would saturate the pores of the cryosphere until either the pore space were filled (Fig. 1B), or the groundwater supply was exhausted (Fig. 1D).

7.2. Was groundwater in direct contact with the cryosphere?

If salty groundwater was in contact with the advancing cryosphere freezing front, groundwater is required to be present down to the pore-closure depth (Fig. 1A) (estimated

at ~10 km depth; Hanna and Phillips, 2005), a scenario in which the Amazonian ICC could be ~4-9 km thick assuming the groundwater was a eutectic solution of NaCl (black line in Fig. 6; Table 2), which is not observed (Fig. 3B and 6). The amount of ice required in the pore space would be in excess of the volume inferred by a factor of ~2 (Table 1). We find that for a depressed ice freezing point of 252 K (salt wt% shown in Table 2), the surface heat flux of Mars would be required to be ~80 mW/m² in order for the depth of the freezing front to match the inferred ICC thickness (and therefore for salty groundwater to be in contact with the cryosphere of the inferred thickness). This is a factor of ~2-5 too large for the Amazonian period (e.g., Montési and Zuber, 2003; Ruiz et al., 2011), and so we consider it more likely that groundwater was not in contact with the cryosphere freezing front as it advanced (e.g., Fig. 1C). Indeed, Russell and Head (2002) found no evidence for a post-impact lake from sub-cryospheric groundwater inflow (e.g., Newsom et al., 2006; Schwenzer et al., 2012) in the Early Amazonian-aged ~215 km diameter Lyot crater in the northern lowlands, leading these researchers to favor the interpretation that groundwater may not have been present below the ICC by the Early Amazonian. Lyot is the deepest location in the northern lowlands, where groundwater is most likely to be in contact with the cryosphere due to the low elevation. The lack of groundwater inflow in Lyot thus suggests that groundwater was not present in the upper martian crust at the time Lyot formed. As pointed out by Russell and Head (2002), however, unusual (and ad-hoc) permeability configurations that prevented the groundwater inflow cannot be ruled out. Harrison et al. (2010) proposed that the fluvial features emanating from the Lyot ejecta are caused by impact-induced groundwater release, but recent work by Head et al. (2016) suggested that impact-ejecta induced

melting (e.g., Weiss and Head, 2016) of surface/near-surface ice deposits might be a more likely explanation on the basis of Lyot's latitudinal association with other surface-ice related features, and distribution of fluvial channels and secondary craters. In this scenario, Lyot is unlikely to have formed in a target hosting underlying groundwater at the time of impact based on the results of Russell and Head (2002). Conversely, the formation of the outflow channels by groundwater discharge implies direct-contact between groundwater (i.e., a thermally-limited cryosphere; Fig. 1A and B) and the ICC to generate hydraulic head (e.g., Baker and Milton, 1974; Carr, 1979, 1996, 2002; Clifford, 1993; Clifford and Parker, 2001; Head et al., 2003; Manga, 2004; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007; Cassanelli et al., 2015).

Another form of data regarding the interaction between groundwater and the cryosphere are the results of numerical models. Grimm and Painter (2009) and Grimm et al. (2016) used a three-phase numerical model of water migration to model the behavior of a 2D pole-to-equator transect of the martian cryosphere and groundwater over time. They found that the ICC within $\sim 30^\circ$ of the equator is entirely sublimated unless a steady groundwater supply exists below the ICC to replenish the equatorial ICC. This is in contrast to the results of our study, which suggest the presence of an equatorial ICC in the absence of underlying groundwater. Grimm et al. (2016) found that the amount of ice lost from the equatorial ICC depended primarily on obliquity (higher obliquities inhibit loss), but was also affected by porosity, pore radius, tortuosity, and heat flux. Our models indicate that obliquity was likely to be between 25° and 45° when the cryosphere freezing front advanced beneath the ICC, which would favor lower loss rates. A better understanding of subsurface ice loss rates (e.g., Bramson et al., 2016) are required in

order to further evaluate our prediction of a thin ICC with no underlying groundwater in the context of multiphase water migration models (Grimm and Painter, 2009; Grimm et al., 2016). For example, Bramson et al. (2016) found that subsurface ice loss rates predicted by current vapor diffusion models (e.g., Schorghofer and Forget, 2012) require the rapid loss of thick excess ice deposits, in contrast to their documented existence in the mid-latitudes from the Middle to Late Amazonian until today (Kress and Head, 2008; Holt et al., 2008; Plaut et al., 2009; Head et al., 2010; Stuurman et al., 2012; Viola et al., 2015; Bramson et al., 2015) and the equator (Head and Weiss, 2014). As pointed out by Grimm et al. (2016), the presence of thin low-porosity layers within the upper crust of Mars not considered in their models (e.g., equatorial regolith hosting pore-ice deposited during periods of high obliquity; Steele et al., 2017) could increase tortuosity and impede sublimation. These factors should be further evaluated to assess whether underlying groundwater is in fact required to replenish the equatorial ICC to avoid complete sublimation as suggested by Grimm et al. (2016).

An additional dataset regarding the interaction between groundwater and the cryosphere are the results of ground penetrating radar. To date, no detections of groundwater reflectors have been made by the Mars Advanced Radar for Subsurface and Ionospheric Sounding (MARSIS) instrument onboard Mars Express, which has a theoretical sounding depth up to ~3-5 km (Picardi et al., 2004). As discussed by Clifford et al. (2010) and Lasue et al. (2013), the absence of groundwater detection can be explained by four possible factors: (1) groundwater may not exist below the ICC at the present time; (2) groundwater is present below the ICC but below the maximum sounding depth of MARSIS (deeper than ~3-5 km); (3) the attenuating properties of the martian

subsurface may prevent MARSIS from reaching its maximum sounding depth (Farrell et al., 2009); and (4) the possibility that thin films of water eliminate the dielectric contrast between the ICC and groundwater, preventing detection of a reflector. Thus, as noted by Farrell et al. (2009) and Clifford et al. (2010), the lack of detection of groundwater by orbiting radar instruments does not rule for or against the presence of sub-cryospheric groundwater on Mars.

7.3. Formation of outflow channels in a supply-limited cryosphere

The primary line of evidence for a global groundwater system on Mars (in contact with the ice-cemented cryosphere) are the outflow channels (Clifford, 1993; Clifford and Parker, 2001), which are hypothesized to result from groundwater discharge sourced by aquifers that fully saturate the pore space beneath a thermally-limited (Fig. 1A and B) ice-cemented cryosphere (Baker and Milton, 1974; Carr, 1979, 1996, 2002; Clifford, 1993; Clifford and Parker, 2001; Head et al., 2003; Manga, 2004; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007) in the Hesperian and Amazonian periods (e.g., Rodriguez et al., 2015). Critically, any model of outflow channel formation that requires a global subsurface fully saturated with groundwater is inconsistent with our results. One such model for aquifer pressurization relies on hydraulic head supplied by groundwater recharge from basal melting of a south polar cap (Clifford, 1993). As noted by Carr (2002), however, the elevation of some outflows channels are too high for this mechanism to operate for all of the outflow channels. Recharge by basal melting of ice caps on Tharsis have alternatively been proposed to supply the recharge because the elevation of Tharsis is sufficient to provide hydraulic head for all of the outflow channels

(Harrison and Grimm, 2004; Russell and Head, 2007; Cassanelli et al., 2015). This model is also uncertain, however, because (1) basal melting is generally not predicted to occur except in localized regions of highly elevated heat flux (“heat-pipe drain pipe” effect; Cassanelli et al., 2015); (2) basal melting of ice sheets on Tharsis is unlikely to have supplied sufficiently high volumes of water to form the outflow channels (Cassanelli et al., 2015); and (3) groundwater flow models do not predict Tharsis-sourced groundwater to discharge in the locations where outflow channels are observed, even in the case where groundwater may follow preexisting fractures so that superlithostatic groundwater pressures are not required (Harrison and Grimm, 2009).

An alternative model for aquifer overpressurization that does not rely on recharge from the surface was explored by Carr (1979, 1996, 2002). In this model, as the freezing front of the cryosphere advances deeper in the martian crust and groundwater freezes onto the growing cryosphere, the volume expansion from water to ice causes the pore pressure of the underlying groundwater to increase. When the pore pressure of the groundwater exceeds the lithostatic pressure, the groundwater may fracture the cryosphere and discharge on the surface to produce the outflow channels. Hanna and Phillips (2005) point out that any lateral confinement of the aquifer makes this hypothesis unlikely because the groundwater would diffuse away toward the edges of the confined portion of the aquifer, thereby reducing the pore pressure. Wang et al. (2006) further tested whether this model could provide sufficient pore pressures and water discharge volumes in the best-case scenario of a fully confined aquifer. Wang et al. (2006) found that, for the updated K value used in our study (4.28 km; Section 2.4) and a pore closure depth of 10 km (Hanna and Phillips, 2005), pore pressures are insufficient to breach the

cryosphere. Wang et al. (2006) found that the pore closure depth must be at least 5 km for the pore pressures to breach the cryosphere, but that the water volumes discharged in this case were negligible. Thus, pore-pressure increase by an advancing cryosphere freezing front may not be a viable candidate to form the outflow channels. In summary, none of the groundwater recharge and aquifer overpressurization mechanisms quantitatively explored in the literature to date (summarized above) adequately explain the formation of the outflow channels.

Even if sufficient recharge and pressurization can be supplied an additional complication arises: are groundwater discharge rates sufficiently high to carve the outflow channels? Outflow channel events are typically estimated to have required flow rates on the order of $\sim 10^6$ - 10^8 m³/s (e.g., Table 2 in Kleinhans, 2005; Leask et al., 2007; Wilson et al., 2009) in order to generate the necessary erosion. Previous investigators who modeled groundwater discharge adopted the upper limit of terrestrial crustal permeability and found that the discharge rates are indeed sufficient (Manga, 2004; Hanna and Phillips, 2005). Later work used a more realistic range of aquifer permeability in their 3D groundwater models to calculate the discharge, frequency, and duration of groundwater-sourced outflow channel events (Harrison and Grimm, 2008). Their models predicted extremely low discharge rates (generally below $\sim 10^6$ m³/s after only the first few minutes to hours after flooding initiates) and an unreasonably high frequency of discharge events (hundreds to thousands), which led these authors to “doubt the ability of groundwater flows to produce the large erosive forms observed in the outflow channels,” and alternatively proposed that breaching of large standing bodies of water at the surface or near-surface may be more consistent with the formation of outflow channels (Harrison

and Grimm, 2008).

The discrepancy between a supply-limited ICC and evidence for pressurized groundwater in the Hesperian and Amazonian (e.g., Rodriguez et al., 2015) might be explained by the regional compartmentalization of groundwater aquifers (Harrison and Grimm, 2009). Harrison and Grimm (2009) conducted 3D numerical groundwater models with recharge above Tharsis and the south pole and found that a globally-integrated groundwater aquifer system could not produce groundwater breakout at the locations of the outflow channel sources, even in the modeled case where groundwater discharge did not require cryosphere disruption through overpressurization. These authors thus concluded that if the outflow channels did form through groundwater discharge, either (1) the martian aquifer system was compartmentalized on local to regional scales (e.g., geologic features such as Tharsis or regional dike systems could act as lateral or vertical aquicludes), or (2) the distribution of groundwater was spatially heterogeneous in the martian crust. Harrison and Grimm (2009) thus suggested that either the martian groundwater system was global but regionally compartmentalized, or the amount and spatial distribution of groundwater in the subsurface was limited. Alternatively, other proposed mechanisms for the formation of these outflow channels which do not require that aquifer pressurization is operating, include: (1) breaching of standing bodies of water at the surface/near-surface (Coleman and Baker, 2007; Harrison and Grimm, 2008) generated by, for example, top-down heating and melting of surface ice deposits (e.g., Cassanelli and Head, 2016); (2) melting of the cryosphere and discharge by dike intrusions (McKenzie and Nimmo, 1999; Head et al., 2003; Craft and Lowell., 2012); (3) bottom-up heating (Zegers et al., 2010); and/or (3) an exclusively volcanic origin for

these outflow channels (Leverington, 2004, 2007, 2009, 2011; Hurwitz and Head, 2012; Hopper and Leverington, 2014). A reassessment of individual outflow channel flow rates and erosive potential (Wilson et al., 2004, 2009, Kleinhans, 2005) may provide insight as to whether any of the alternative formation mechanisms discussed above warrant further investigation.

In summary, our model of a supply-limited ICC is generally incompatible with outflow channel formation sourced by groundwater discharge because this model requires that the pores of the subsurface are fully saturated with groundwater down to the pore-closure depth (i.e., a thermally-limited cryosphere). On the basis of the complicating factors for outflow channel formation discussed above, we suggest that other mechanisms for outflow channel formation should be further evaluated. It is not our goal in this paper to revise any outflow channel formation hypotheses—rather, we present our evidence and analysis independently and suggest that this work may motivate a second look at the formation of outflow channels. If the outflow channels did not form through discharge of a pressurized globally integrated groundwater system, note that our minimum estimates for the Late Noachian mean annual surface temperature ($\geq 212\text{--}227\text{ K}$) and atmospheric pressure ($\geq 390\text{--}850\text{ mbar CO}_2$ atmosphere) (Section 5.2) may be overestimated. For example, if the martian groundwater system was cold-trapped to the cryosphere during the Late Noachian period, atmospheric temperatures and pressures could have been lower during this period.

7.4. Consequences for groundwater abundance

Our model results suggest that the cryosphere freezing front could have propagated

beneath the base of the ice-cemented cryosphere, at which point there was no longer an abundant groundwater source to input ice in the thickening cryosphere layer (e.g., Fig. 1D). This led to the thickness stabilization of the ICC by ~ 3.0 to ~ 3.3 Ga at the latest (assuming a predominantly CO₂ atmosphere) (Fig. 17). Because our models with atmospheric pressures ≥ 800 mbar are unable to reproduce the form of the inferred ICC (Fig. 13B and C), we suggest that the groundwater supply was likely to have been exhausted during a period where the martian atmospheric pressure was $\leq \sim 600$ mbar (Fig. 17). If large volumes of groundwater were present and globally integrated below the ICC beyond the Hesperian period (i.e., available to thicken the global ICC through upward vapor diffusion), the ICC would better match the thermal models using Amazonian heat fluxes (e.g., Fig. 6 and 7). Additionally, the inferred ICC would not be expected to retain the thinned ICC at the southernmost high latitudes (dashed red circle in Fig. 7) because underlying groundwater would have diffused upwards and frozen onto the growing ICC. We suggest (Section 6) that this feature (dashed red circle in Fig. 7) could be caused by cryosphere melting from the overlying insulating Dorsa Argentea Formation during the Late Noachian-Hesperian period (Fig. 17) (Head and Pratt, 2001; Ghatan and Head, 2002, 2004; Fastook et al., 2012; Scanlon et al., 2013; Scanlon and Head, 2014).

Based on the anomalously thin ICC thicknesses (~ 1.3 - 2.3 km) derived in Section 2 (Fig. 3B), the results of our thermal models (Fig. 13 and 15), and the lack of an observed deep globally integrated groundwater system in the Amazonian (e.g., Russell and Head, 2002), we suggest that the total groundwater supply below the ICC was insufficient to fill the pore space of the cryosphere, and that a deep, globally or regionally integrated

groundwater system did not persist in the subsurface beyond the Late Hesperian or Early Amazonian period (Fig. 17).

8. Conclusions

The martian cryosphere is the zone in the subsurface characterized by temperatures below the freezing point of water, allowing water ice to be thermally stable (Fig. 1). The martian ice-cemented cryosphere (ICC) is the reservoir of pore ice within the cryosphere that extends into the subsurface (Fig. 1). Previous investigators have assessed the theoretical thickness of the martian cryosphere on the basis of thermal models (Fig. 6), but the depth to which ice fills the pore space has remained unknown. Estimating the thickness of the portion of the cryosphere that is ice-cemented is critical to our understanding of the martian global water inventory and the presence, extent, and/or absence of a groundwater system during the history of Mars. For example, was the martian cryosphere thermally-limited (Fig. 1A and B), or supply-limited (Fig. 1C and D)? We evaluated thermal models and crater excavation-depth relationships in tandem to examine the characteristics of the martian ICC. We surveyed the excavation depths of (1) an Amazonian- to Hesperian-aged crater population interpreted to form in an ice-cemented target, single-layered ejecta (SLE) craters; and (2) crater classes that we tentatively interpret to penetrate through an ice-cemented target: radial ejecta and multiple-layered ejecta (MLE) craters (Fig. 2). These excavation depths are interpreted to reflect the Amazonian- to Hesperian-aged ICC thickness. We compared this ICC thickness estimate with cryosphere thermal models using Amazonian through Late Noachian heat flux, surface temperature, atmospheric pressure, and obliquity

configurations. Our results suggest the following:

- (1) The ICC thickness inferred from SLE and MLE crater excavation depths is ~ 1.3 km thick at the equator, and ~ 2.3 km thick at the poles (Fig. 3B) during the Hesperian-Amazonian periods.
- (2) This corresponds to a pore ice volume of $\sim 3 \times 10^7$ km³, equivalent to a martian global equivalent layer (GEL) of water of ~ 200 m, much lower than previous estimates based on the available pore space within the cryosphere (~ 580 - 1160 m GEL; Table 1, and Clifford et al., 2010).
- (3) The inferred ICC thickness is not in agreement with Amazonian cryosphere models, which generally predict a much thicker cryosphere (Fig. 6). This suggests that the martian cryosphere is supply-limited. Thermal models which incorporate higher heat fluxes, atmospheric pressures, and obliquities, however, can reproduce the inferred ICC thicknesses (Fig. 13). This suggests that the ice-cemented cryosphere reached its current thickness in a more ancient period of martian history (Fig. 1C), under obliquities between 25° and 45° and atmospheric pressures likely to be $\leq \sim 600$ mbar, and that no abundant, globally-integrated groundwater system exists below the cryosphere in the present day (Fig. 1D).
- (4) If this interpretation is correct, our thermal models constrain Late Noachian (>3.6 Ga) mean annual surface temperatures to ≥ 212 - 227 K, assuming that groundwater persisted in the Late Noachian period and that the surface heat flux was ≤ 60 mW/m². If the Late Noachian exhibited a pure CO₂ atmosphere, atmospheric pressures at 3.6 Ga are then predicted to be ≥ 390 - 850 mbar.
- (5) Thermal models constrain the age during which the ice melting isotherm reached the

base of the ice-cemented cryosphere to a time period of ~3.0-3.3 Ga (the Late Hesperian to Early Amazonian) at the latest (assuming a pure CO₂ atmosphere with a water cycle). After ~3.0-3.3 Ga, our models predict that abundant groundwater did not persist in the deep martian subsurface (Fig. 17).

(6) The thinner ICC in the southernmost high-latitudes (75°S) is interpreted to be due to the presence of a $\sim 1 \pm 0.5$ thick thermally insulating ice cap on the surface out to 75°S during the Late Noachian-Early Hesperian periods (the Dorsa Argentea Formation; Fig. 16).

(7) Our model of a supply-limited cryosphere (Fig. 1A) is generally inconsistent with an origin for the outflow channels involving discharge from a globally-integrated subcryospheric groundwater system. Future work is required to reconcile these contrasting models for the martian hydrologic evolution.

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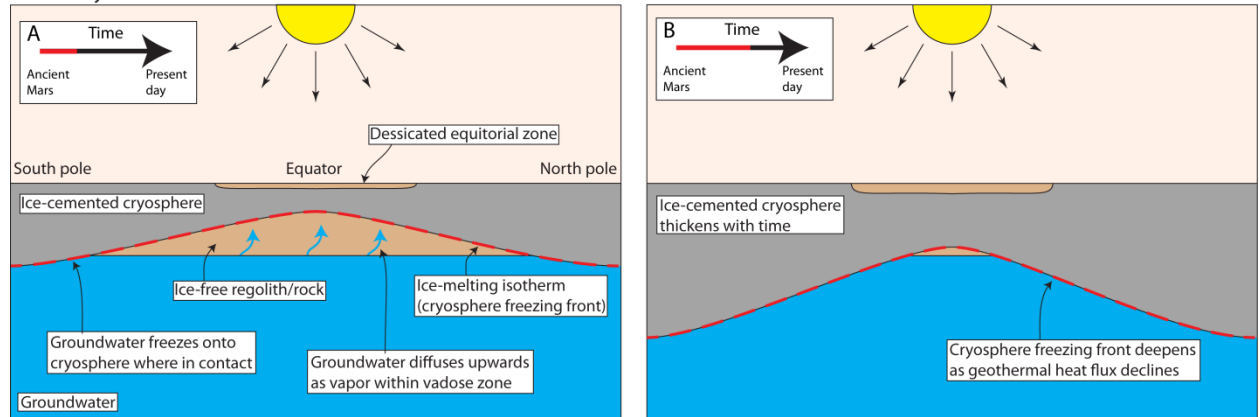
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Figures, tables, and captions:

Thermally-limited



Supply-limited

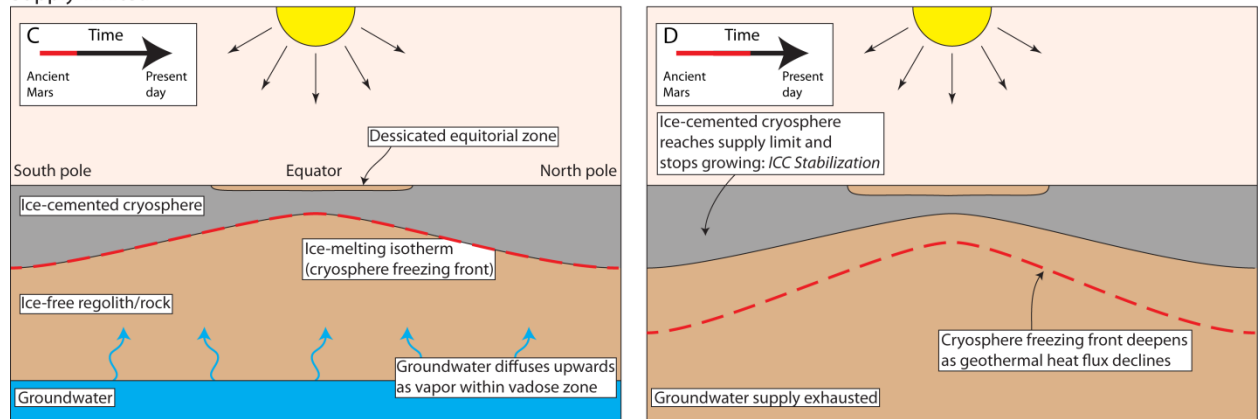


Figure 1. Schematic of the martian cryosphere (dashed red line), and the ice-cemented cryosphere (shaded in grey). (A) The top panels show the case of a cryosphere that is thermally-limited, with no groundwater supply limit. Groundwater freezes onto the freezing front where in contact, and diffuses upwards as vapor in places where groundwater is not in contact with the freezing front. (B) As the geothermal heat flux declines with time, water continues to freeze onto the freezing front and the ice-cemented cryosphere grows. (C) The bottom panels show the case of a cryosphere with a groundwater supply-limit. (D) Once the groundwater supply is exhausted, the ice-cemented cryosphere stops growing, even as the freezing front advances deeper in the subsurface.

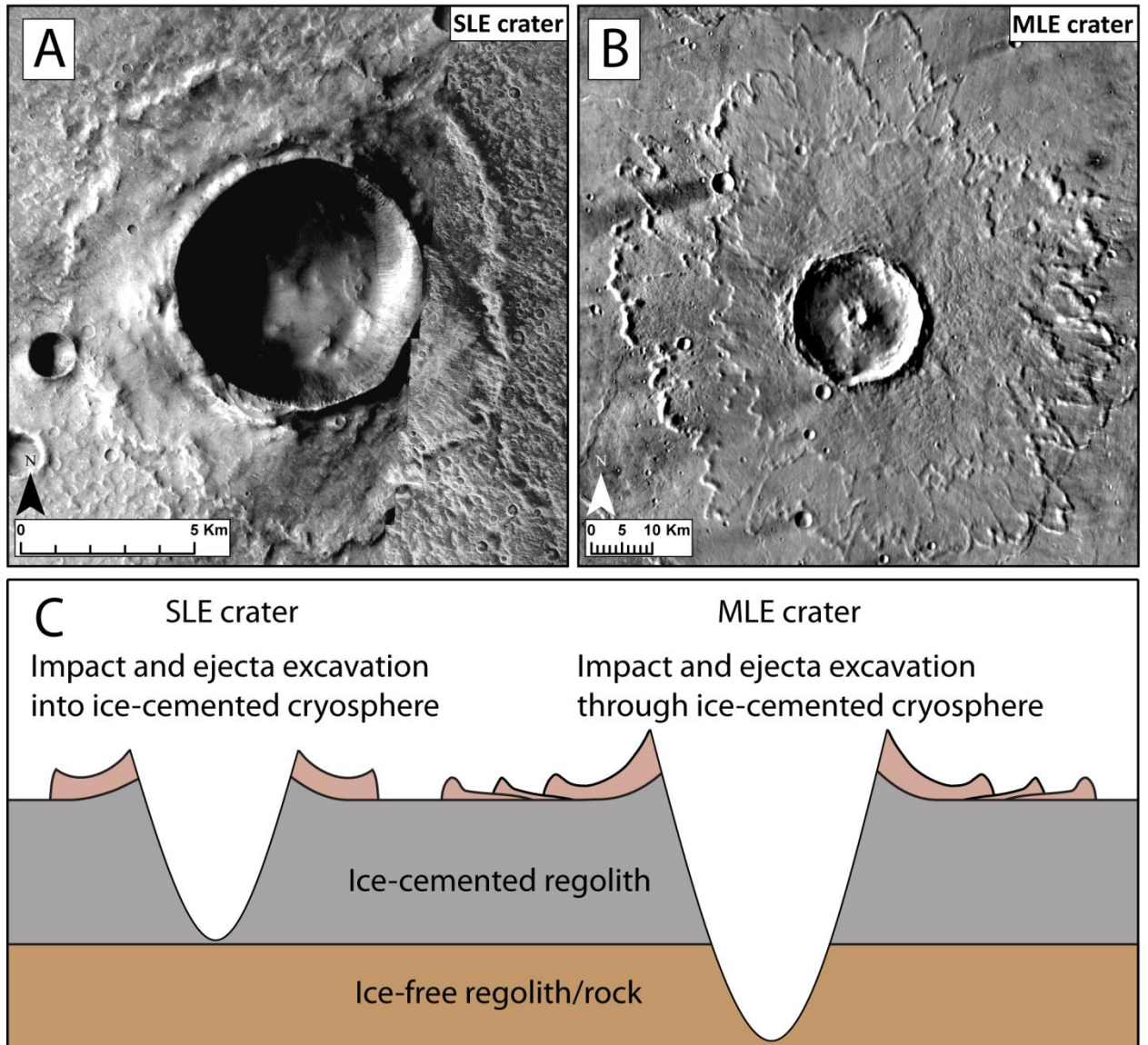


Figure 2. Martian impact craters interpreted to form in the ice-cemented cryosphere. (A) SLE crater, 7.2 km diameter; 2.76°N, 74.5°E; THEMIS VIS V26756014, (B) MLE crater, 21 km diameter; 5.9°N, 70.53°E; THEMIS IR day global mosaic, (C) Simplified target structure for SLE and MLE craters. SLE craters are interpreted to excavate within the ice-cemented cryosphere, while MLE craters are interpreted to excavate below the ice-cemented cryosphere.

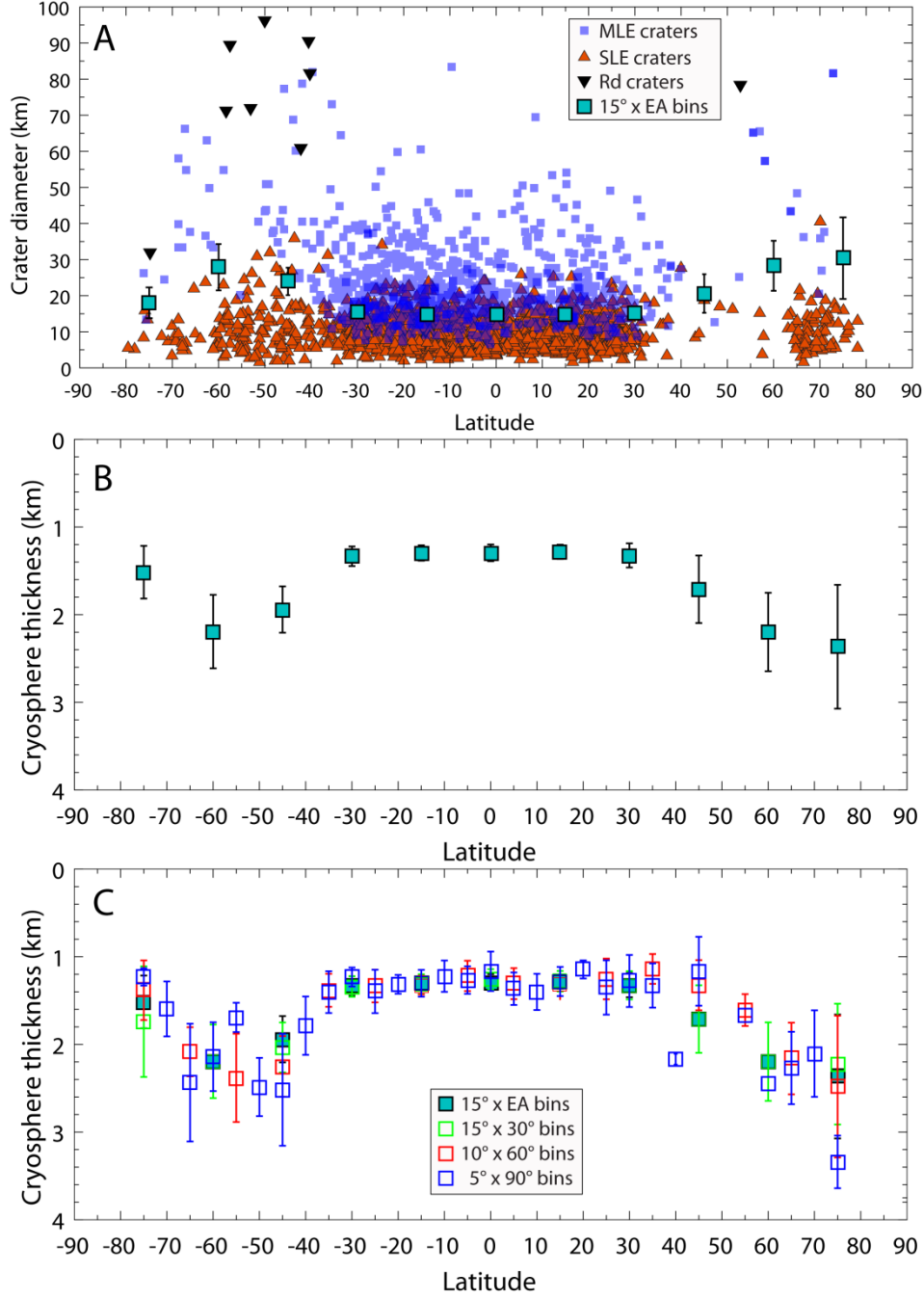


Figure 3. Cryosphere thickness estimate inferred from SLE and MLE craters. (A) Latitudinal relationships of the MLE (blue squares), SLE crater populations (red triangles) modified from Weiss and Head (2014), and radial (Rd) craters modified from Barlow (1988). SLE/MLE transition diameter is shown for 15° latitude bins averaged across equal-area (EA) longitude bins (green squares; 15° at the equator, increasing in size toward the poles to account for decreasing area). Error bars show the standard error (SE) of the difference between the mean of the SLE and MLE craters in each bin:

$$SE_{\sigma_{MLE}-\sigma_{SLE}} = \sqrt{\frac{\sigma_{MLE}^2}{N_{MLE}} + \frac{\sigma_{SLE}^2}{N_{SLE}}}, \text{ where } \sigma \text{ is standard deviation and } N \text{ is the sample}$$

number in each bin. (B) Ice-cemented cryosphere thickness inferred from SLE/MLE crater transition diameter. (C) Inferred ice-cemented cryosphere thickness derived using different bin dimensions: the 15° latitude by EA longitude bins (filled green squares), 15° latitude by 30° longitude bins (open green squares), 10° latitude by 60° longitude bins (red squares), and 5° latitude by 90° longitude bins (blue squares).

	Clifford (1993) porosity model			
Φ_0	0.15	0.20	0.25	0.3
$V_{\text{ICC}} (10^7 \text{ km}^3)$	2.41	3.21	4.01	4.81
$\text{GEL}_{\text{ICC}} (\text{m})$	152	203	254	305
$V_{\text{below}} (10^7 \text{ km}^3)$	5.57	7.43	9.29	11.15
$\text{GEL}_{\text{below}} (\text{m})$	385	513	642	770
$V_{\text{total}} (10^7 \text{ km}^3)$	8.36	11.48	13.94	16.72
$\text{GEL}_{\text{total}} (\text{m})$	577	770	962	1155

Table 1. Volume of the inferred ice-cemented cryosphere (V_{ICC}) and global-equivalent water layer of the ICC (GEL_{ICC}) derived from varying the initial porosity (Φ_0) from Eq. (1) using a porosity decay constant of 4.28 km (Weiss and Head, 2017). Also shown is the volume (V_{below}) and corresponding global equivalent layer ($\text{GEL}_{\text{below}}$) of the pore space between the ICC and a 10 km pore closure depth, and the total volume (V_{total}) and global equivalent layer ($\text{GEL}_{\text{total}}$) of pore space within the upper crust of Mars.

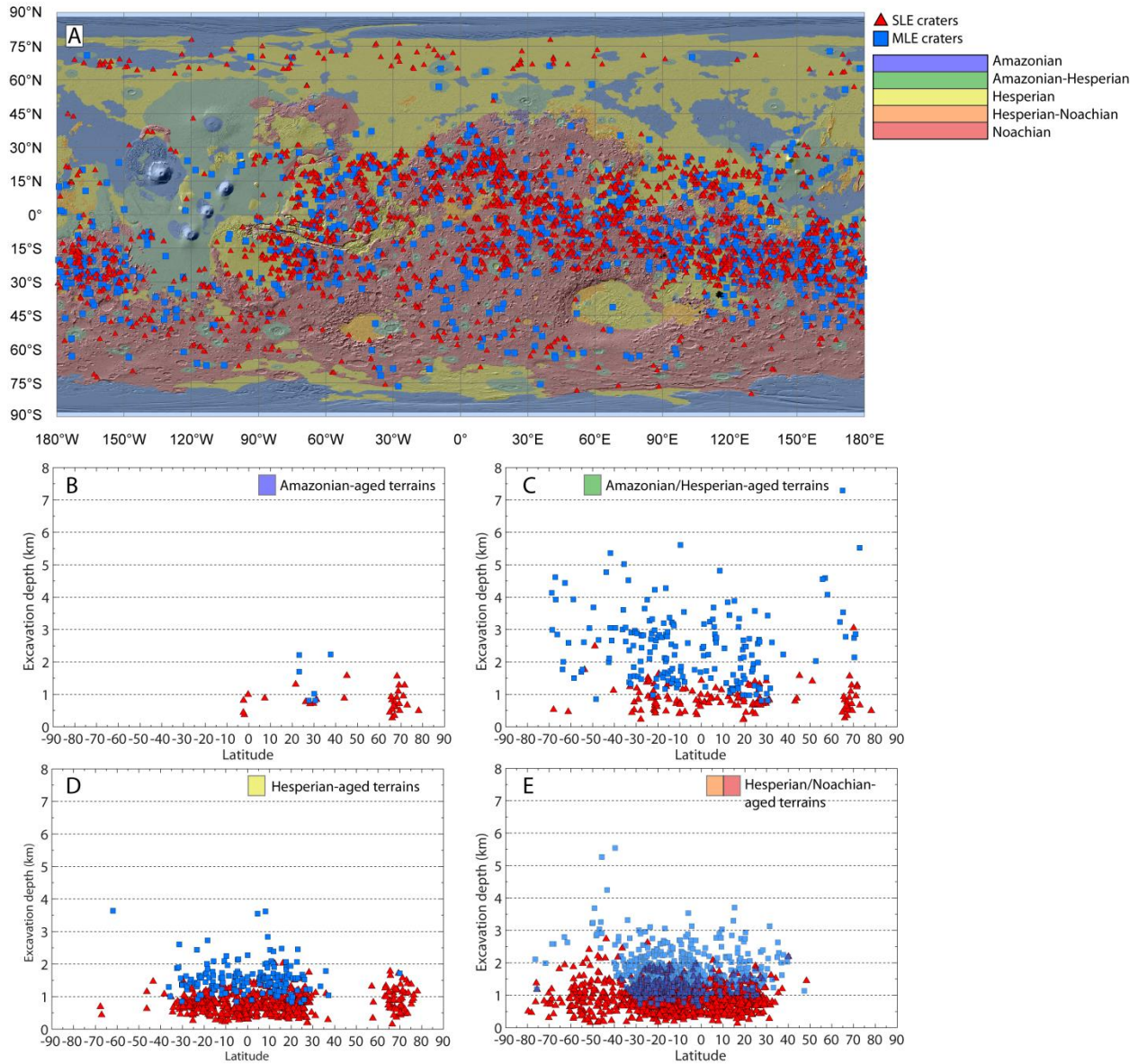


Figure 4. Terrain-age and excavation depth relationships for the SLE and MLE craters. (A) Terrain age units from the geologic map of Tanaka et al. (2014a) overlain on MOLA shaded relief map. Amazonian-aged terrain (blue), Amazonian- or Hesperian-aged terrain (green), Hesperian-aged terrain (yellow), Hesperian- or Noachian-aged terrain (orange), Noachian-aged terrain (red). Distribution of single-layered ejecta (SLE; red triangles) and multiple-layered ejecta (MLE; blue squares) used in this study. Latitude and excavation depths of SLE and MLE craters in (B) Amazonian-aged terrains, (C) Amazonian- or Hesperian-aged terrains, (D) Hesperian-aged terrains, and (E) Noachian- (or Hesperian-) aged terrains.

Salt species	Eutectic melting isotherm in K (wt% salt required)	Melting isotherm (K) with salt		Salt wt% required to reach 252 K melting isotherm	Initial salt content required (wt%) to reach eutectic through freezing of the inferred ice-cemented cryosphere
		5 wt%	10 wt%		
Halite NaCl	252 (23.3 wt%)	270.1	266.5	23.3	16.7
Magnesium perchlorate ^a Mg(ClO ₄) ₂	206 (44 wt%)	271.2	269.2	30	31.5
Sodium perchlorate ^a NaClO ₄	236 (52 wt%)	272.7	270.9	42	37.3
Magnesium sulfate ^b MgSO ₄	269 (17 wt%)	272.5	271.7	N/A	12.2

Table 2. Eutectic temperatures and wt% required for a variety of candidate martian salt species. Also shown is the melting isotherm for 5-10 wt% salt, the salt content required to reach the 252 K isotherm, and the initial salt content required to reach the eutectic through concentration of salts in the underlying groundwater by progressive freezing of the thickness of the inferred ice-cemented cryosphere.

^a Chevrier et al. (2009)

^b Hogenboom et al. (1991)

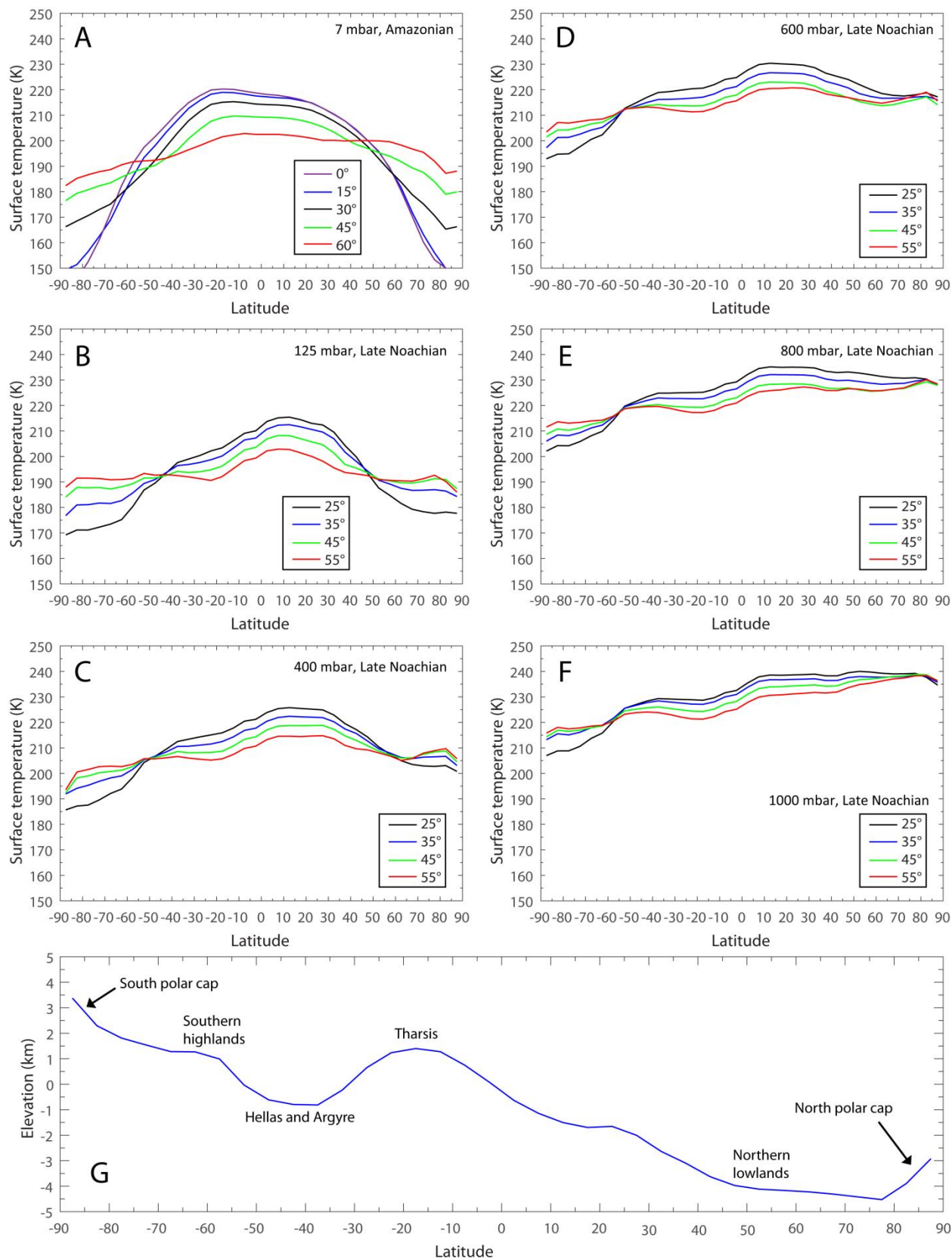


Figure 5. Mean annual surface temperatures used in the thermal models. (A) Zonally averaged martian temperatures for the Amazonian period from the climate models of Haberle et al. (2003) for different obliquities. (B) Zonally averaged martian temperatures for the Late Noachian period (3.8 Ga) from the climate models of Horan and Head (2016) (GCM from Forget et al., 2013 and Wordsworth et al., 2013, 2015) for an atmospheric pressure of 125 mbar (CO₂ atmosphere with a water cycle) and obliquities of 25° (black), 35° (blue), 45° (green), and 55° (red). (C) 400 mbar atmosphere. (D) 600 mbar atmosphere. (E) 800 mbar atmosphere. (F) 1000 mbar atmosphere. (G) Longitudinally-averaged pole-to-pole MOLA topographic profile (5° bins).

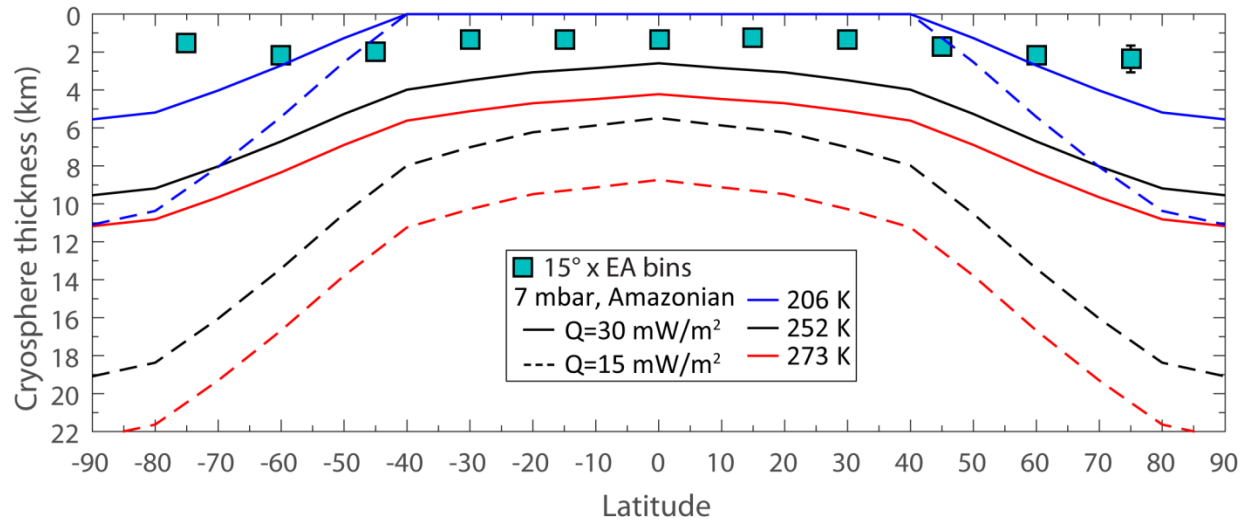


Figure 6. Modeled cryosphere thickness relationships for the Amazonian period of Mars following Clifford et al. (2010). Heat flow used is 15 mW/m^2 (dashed lines) and 30 mW/m^2 (solid lines), 206 K melting isotherm (blue lines), 252 K melting isotherm (black lines), and 273 K melting isotherm (red lines). Ice-cemented cryosphere derived from SLE and MLE crater excavation depths (green squares).

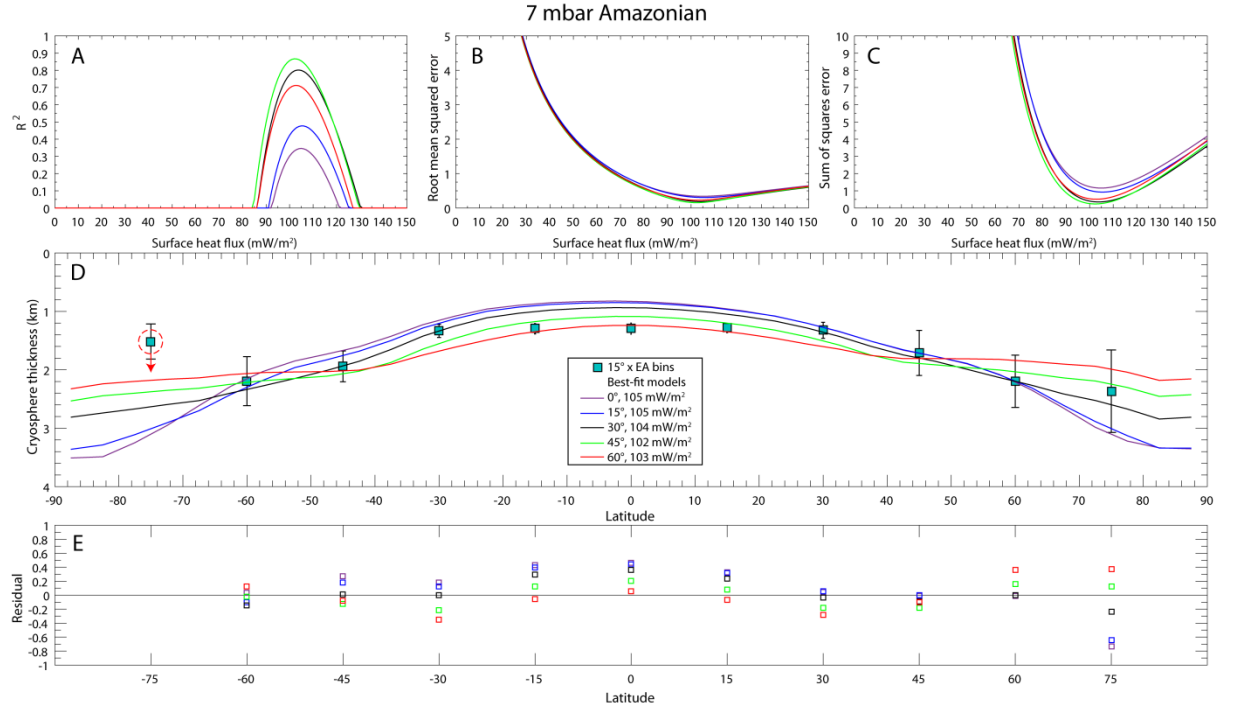


Figure 7. Comparison between the best-fit Amazonian-age thermal model (surface temperatures from Haberle et al., 2003) and ice-cemented-cryosphere (ICC) using a 273 K ice-melting isotherm, and a 300 m equatorial zone of low thermal conductivity ($\kappa_{eq}=1$ W/mK). (A) R^2 values as a function of heat flux between cryosphere thermal models and ice-cemented cryosphere thickness for different obliquities. (B) Root mean squared error. (C) Sum of squares error. (D) Least squares fit cryosphere thermal models compared with inferred ice-cemented cryosphere thickness. Dashed red circle points to anomalously thin ICC in the southern high latitudes (see Section 6). (E) Residuals for (D).

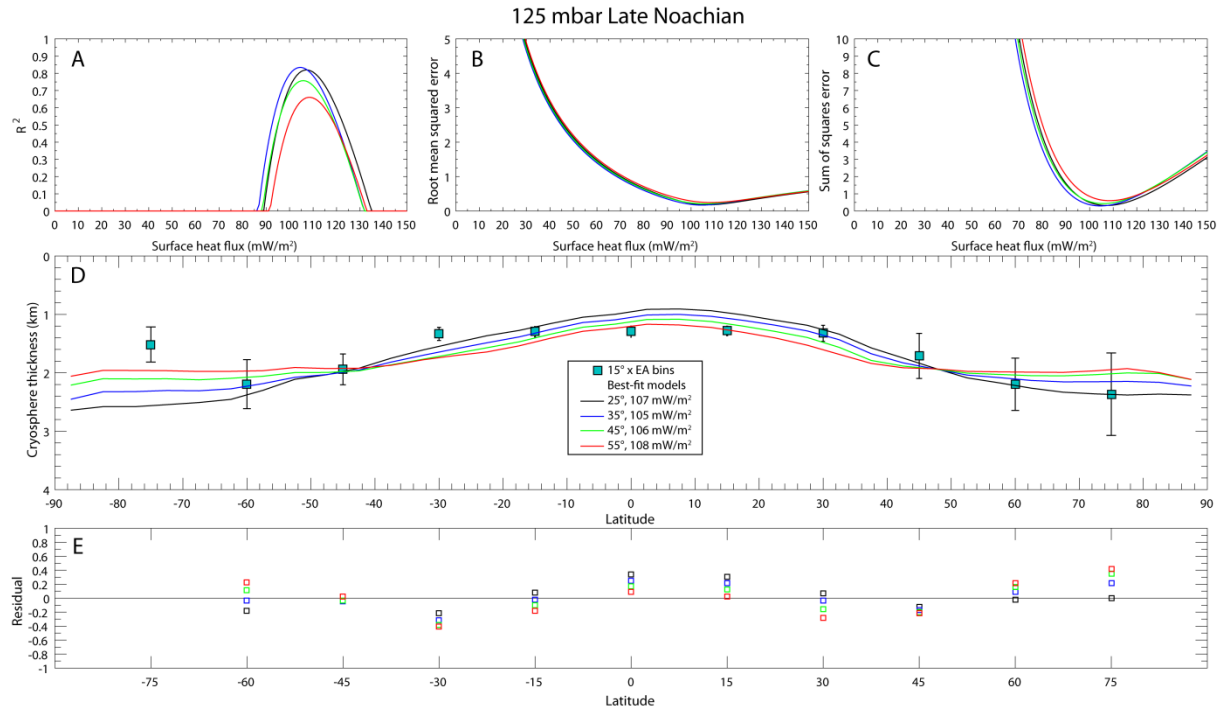


Figure 8. Comparison between the 273 K isotherm model and ICC thicknesses for a 125 mbar Late Noachian CO_2 atmosphere (with a water cycle), and a 300 m equatorial zone of low thermal conductivity ($\kappa_{\text{eq}}=1 \text{ W/mK}$). (A) R^2 values as a function of heat flux between cryosphere thermal models and ice-cemented cryosphere thickness for 25° obliquity (black line), 35° (blue line), 45° (green line), and 55° (red line). (B) Root mean squared error. (C) Sum of squares error. (D) Least squares fit cryosphere thermal models compared with inferred ice-cemented cryosphere thickness. (E) Residuals for (D).

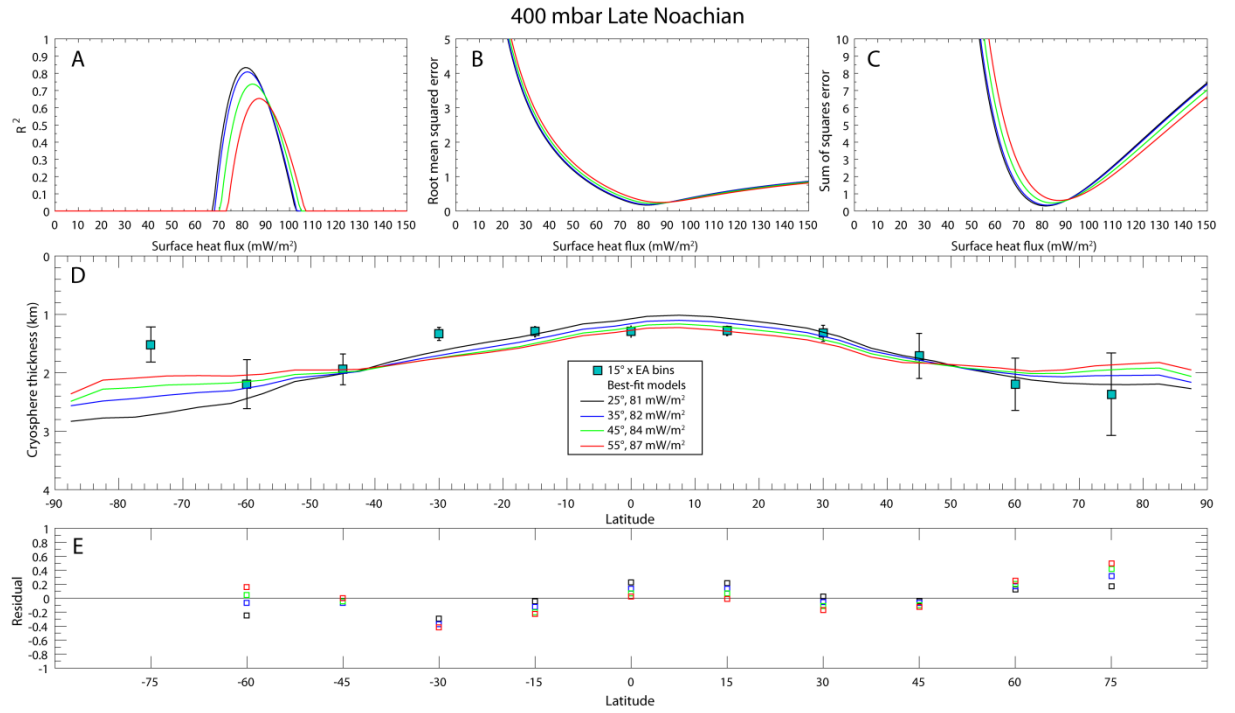


Figure 9. Same as Fig. 8 but for a 400 mbar atmosphere. The 400 mbar atmosphere models produces good fits to the ICC, with R^2 values between 0.65 and 0.83. The best fitting models are for obliquities of 25° and 35°.

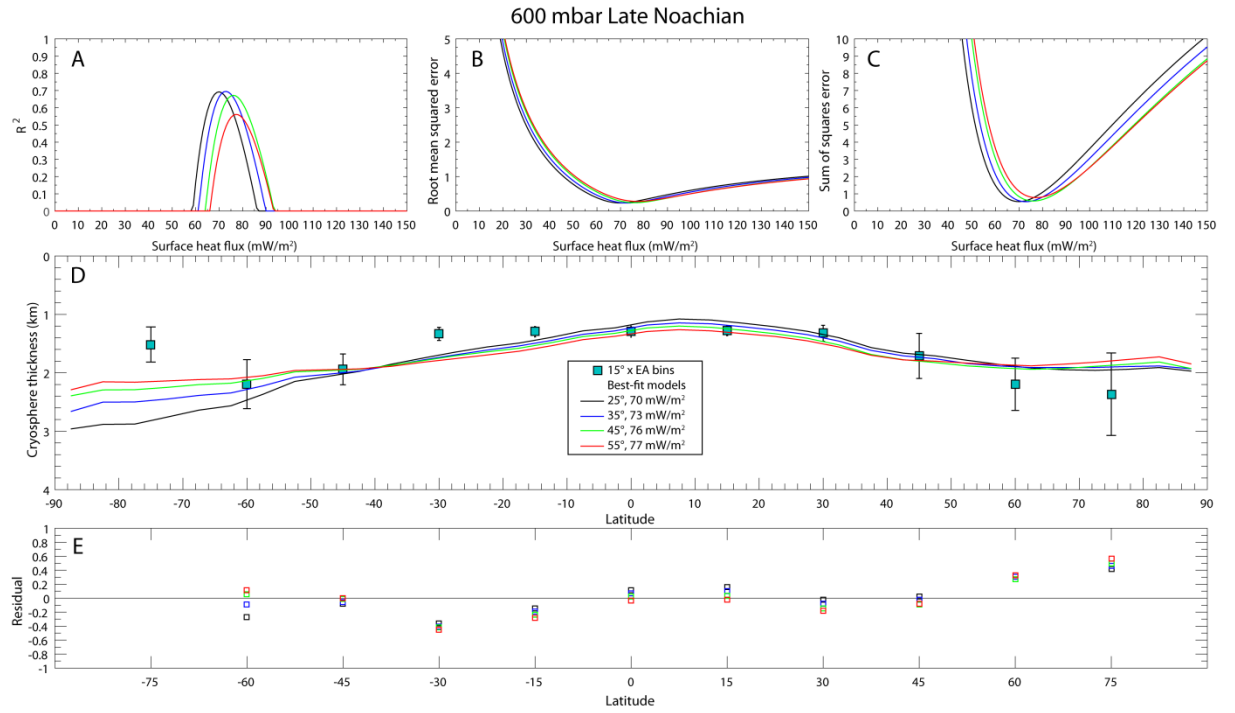


Figure 10. Same as Fig. 8 but for a 600 mbar atmosphere. The 600 mbar atmosphere models produces fair fits to the ICC, with R^2 values between 0.56 and 0.66. The best fitting models are for obliquities of 25° and 35°.

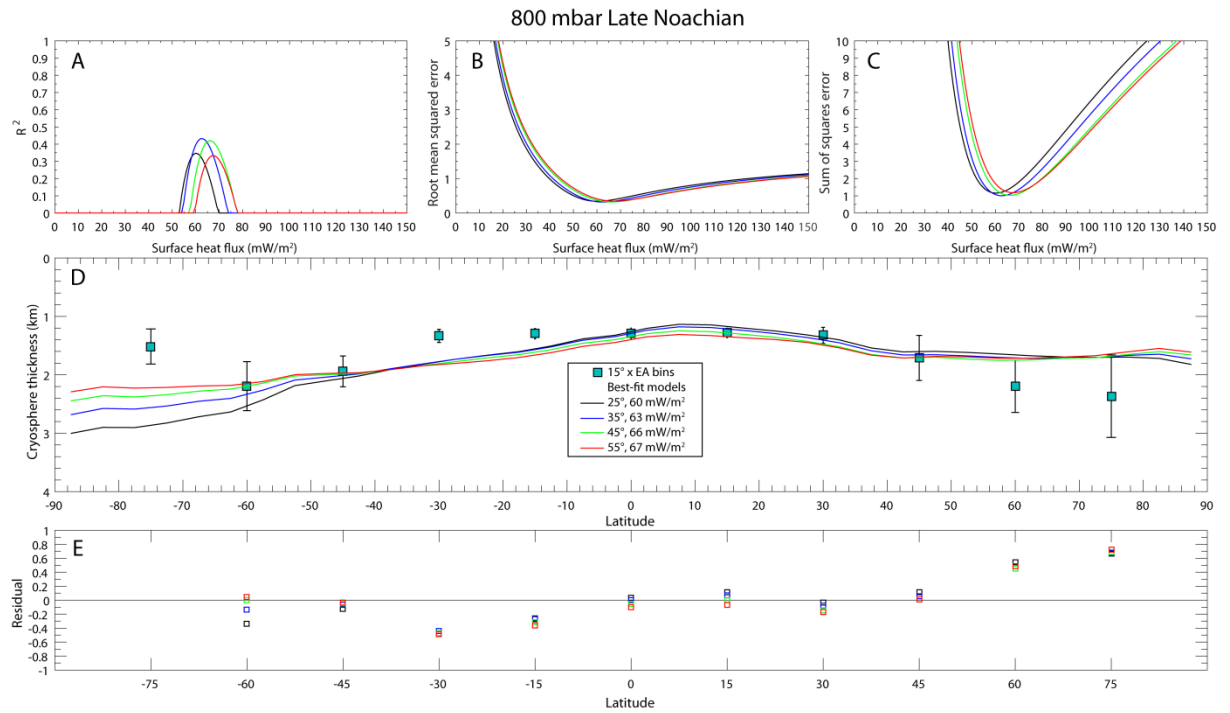


Figure 11. Same as Fig. 8 but for an 800 mbar atmosphere. The 800 mbar atmosphere models produces poor fits to the ICC, with R^2 values between 0.33 and 0.43. The best fitting models are for obliquities of 35° and 45°.

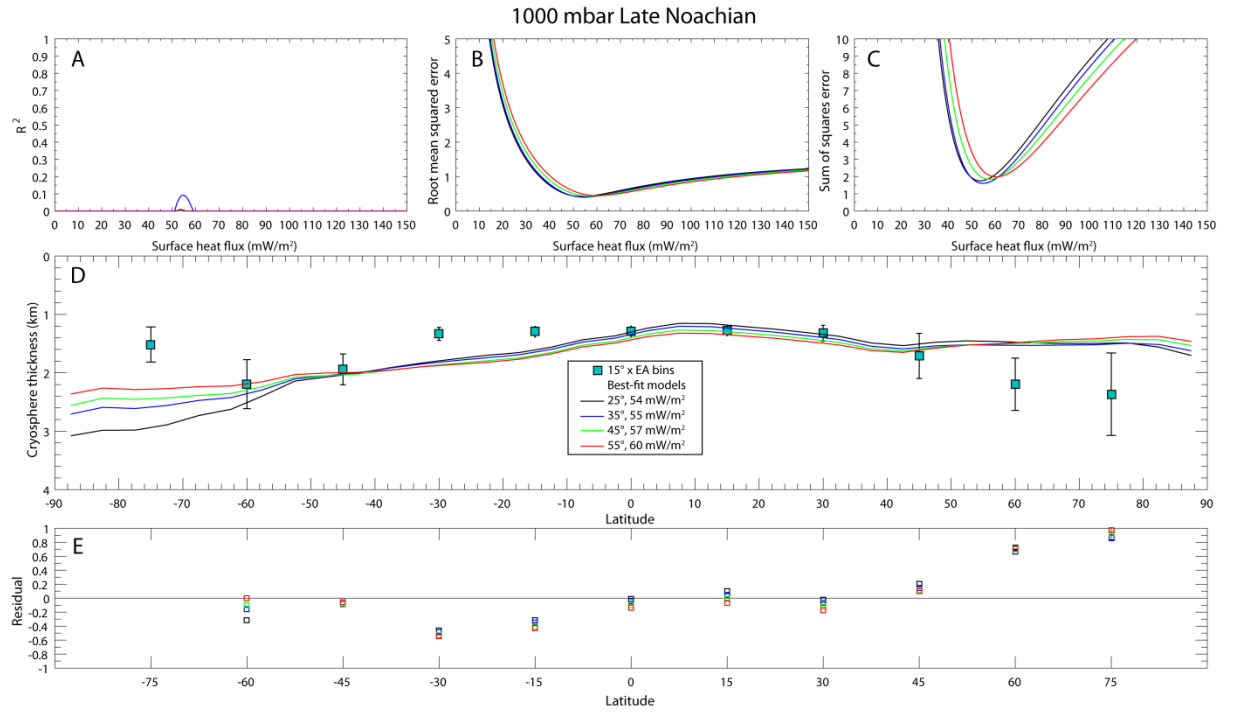


Figure 12. Same as Fig. 8 but for a 1000 mbar atmosphere. The 1000 mbar atmosphere models produces extremely poor fits to the ICC, with R^2 values between 0.00 and 0.09. The best fitting models are for obliquities of 25° and 35°.

P_F (mbar)	Θ (°)	MAST	273 K isotherm model				252 K isotherm model			
			Q_F	R^2	RMSE	SSE	Q_F	R^2	RMSE	SSE
7 (Amazonian)	0	205	105	0.346	0.340	1.156	82	0.000	0.521	2.717
7 (Amazonian)	15	204	105	0.477	0.304	0.925	82	0.000	0.476	2.268
7 (Amazonian)	30	202	104	0.802	0.187	0.351	79	0.435	0.316	0.998
7 (Amazonian)	45	200	102	0.867	0.154	0.236	76	0.805	0.186	0.346
7 (Amazonian)	60	198	103	0.712	0.226	0.509	76	0.734	0.217	0.470
125	25	199	107	0.820	0.179	0.319	82	0.567	0.277	0.765
125	35	199	105	0.834	0.171	0.293	79	0.747	0.212	0.448
125	45	197	106	0.757	0.207	0.429	80	0.743	0.213	0.454
125	55	195	108	0.660	0.245	0.601	82	0.667	0.243	0.589
400	25	214	81	0.833	0.172	0.295	56	0.579	0.273	0.745
400	35	213	82	0.809	0.184	0.338	56	0.732	0.218	0.475
400	45	211	84	0.738	0.215	0.463	58	0.722	0.222	0.492
400	55	209	87	0.654	0.247	0.611	61	0.657	0.246	0.606
600	25	221	70	0.692	0.233	0.544	44	0.383	0.330	1.091
600	35	219	73	0.695	0.232	0.540	47	0.577	0.274	0.749
600	45	216	76	0.672	0.241	0.580	50	0.649	0.249	0.622
600	55	215	77	0.561	0.279	0.777	51	0.514	0.293	0.860
800	25	228	60	0.348	0.340	1.154	35	0.000	0.509	2.588
800	35	226	63	0.432	0.317	1.005	37	0.000	0.421	1.768
800	45	223	66	0.421	0.320	1.023	40	0.160	0.385	1.485
800	55	222	67	0.333	0.343	1.179	41	0.040	0.412	1.698
1000	25	232	54	0.008	0.419	1.755	29	0.000	0.683	4.661
1000	35	231	55	0.091	0.401	1.606	29	0.000	0.607	3.682
1000	45	230	57	0.000	0.430	1.846	32	0.000	0.622	3.864
1000	55	227	60	0.000	0.545	2.968	35	0.000	0.615	3.783

Table 3. Best-fit atmospheric pressure (P_F), mean annual surface temperature (MAST, K), and heat flow (Q_F , mW/m²) configurations between the inferred ice-cemented cryosphere (ICC) and the cryosphere thermal models for both the 273 K isotherm and 252 K isotherm models. Statistics are shown for the case of a 300 m equatorial zone of $\kappa_{eq}=1$ W/mK. Shown are the coefficient of determination (R^2), root-mean-squared error (RMSE, km), and sum of squares error (SSE, km) for the least squares fit between the thermal models and the inferred ICC thickness. R^2 , RMSE, and RSS values were calculated excluding data at 75°S, due to its interpreted modification by an expanded south-polar cap (Section 6).

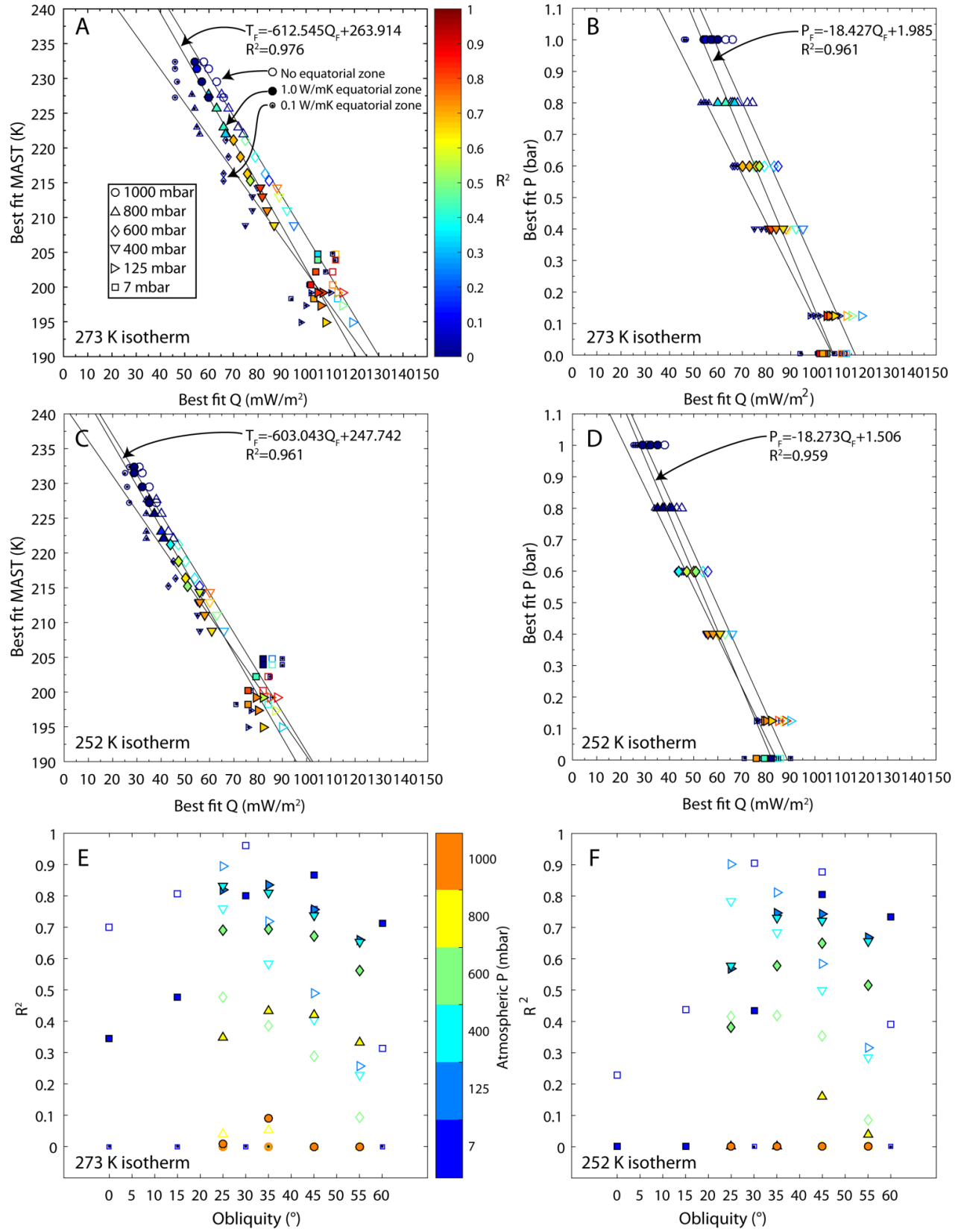


Figure 13. (A) Mean annual surface temperature (MAST) of the least squares fit to the different cryosphere 273 K isotherm models for the three different thermal conductivity configurations derived from a total of $N=22,500$ model runs. Open markers are for the case with no equatorial zone of low thermal conductivity. Filled markers are with a 300 m equatorial zone of $\kappa_{eq}=1.0$ W/mK. Small dotted markers are with a 300 m equatorial zone of $\kappa_{eq}=0.1$ W/mK. 1000 mbar Late Noachian atmosphere (circles), 800 mbar (triangles), 600 mbar (diamonds), 400 mbar (down-facing triangles), 125 mbar (squares), and 7 mbar Amazonian (right-facing triangles). The color of the markers corresponds to the R^2 value of the model fit. (B) Same as (A) but showing the best-fitting atmospheric pressures. (C) Same as (A) but for the 252 K isotherm model. (D) Same as (B) but for the 252 K isotherm model. (E) Obliquity versus R^2 value for the best-fit 273 K isotherm model runs; marker colors correspond to atmospheric pressure. (F) Same as (E) but for the 252 K isotherm model.

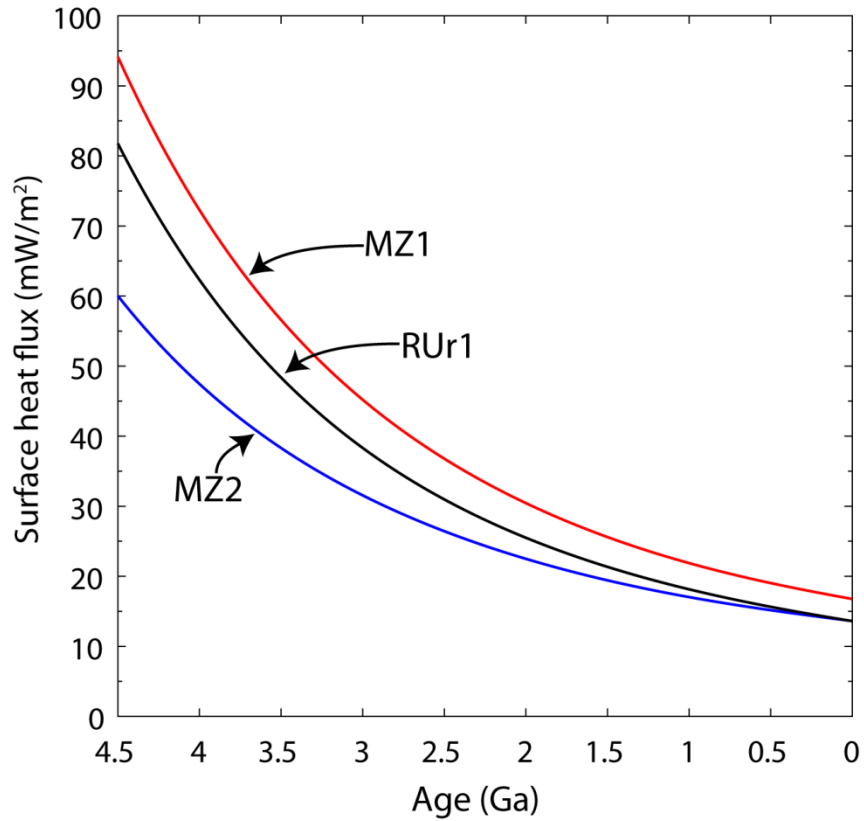


Figure 14. Global average surface heat flux over time derived from martian interior heat balance models of Montési and Zuber (2003) for an upper heat flow (red line; MZ1), a lower heat flow (blue line; MZ2), and a heat flow model from Ruiz et al. (2011) with a Urey ratio of 1 (black line; RUr1).

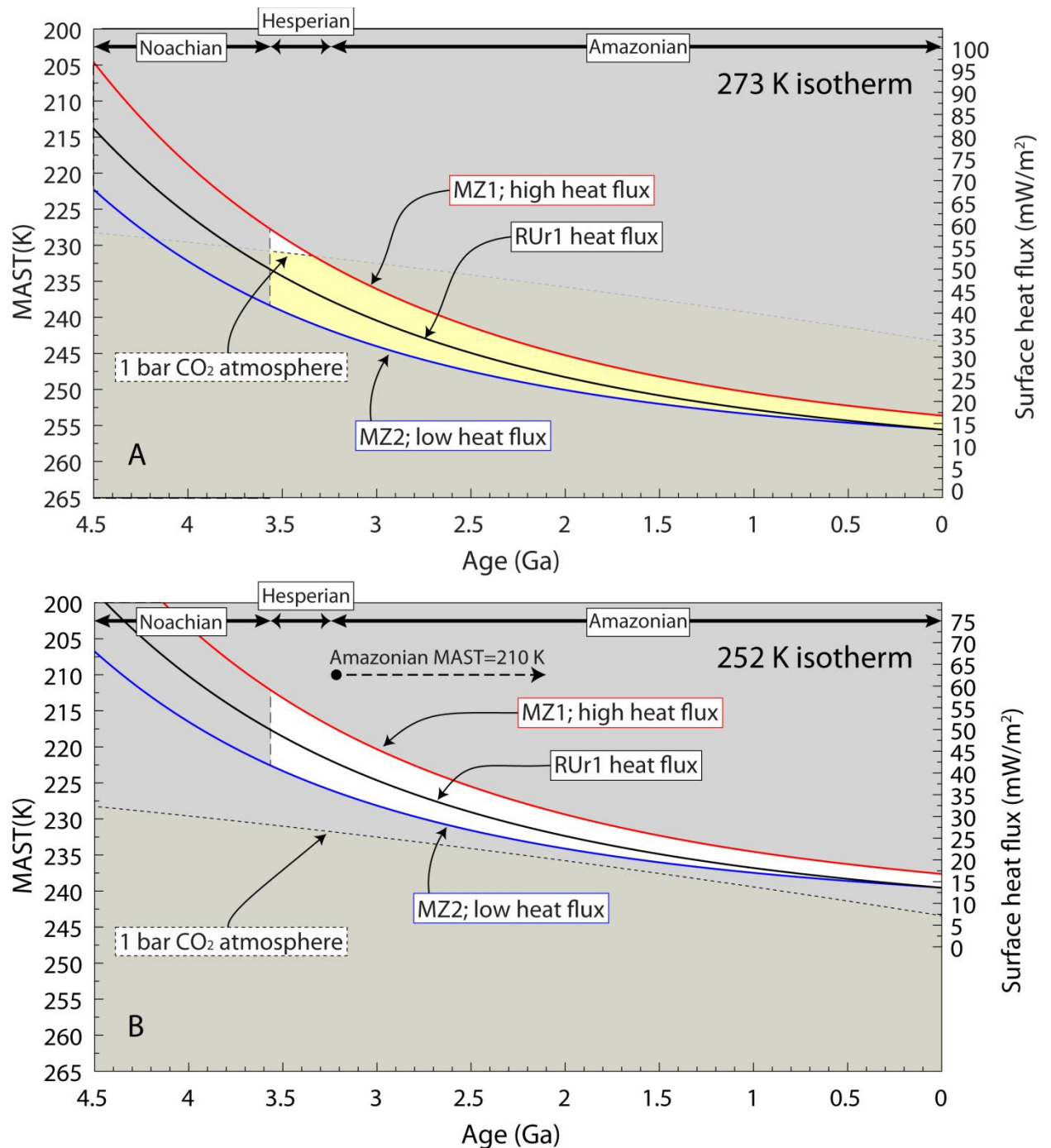


Figure 15. Best-fit mean annual surface temperature and surface heat flux relationships over time which allow the ICC to stabilize; for MZ1 heat flux (red line), MZ2 heat flux (blue line), and RUr1 heat flux (black line). (A) 273 K isotherm model. (B) 252 K isotherm model. These lines depict the MAST and heat fluxes required for the cryosphere freezing front to reach base of the ice-cemented cryosphere (ICC) (i.e., the time at which the ICC reaches the subsurface ice supply-limit). Greyed areas within the plot can be ruled out (see Section 5.1). The shaded yellow region depicts the area that can be ruled

out if the martian atmosphere at 3.6 Ga was at most a 1 bar (Kite et al., 2014) CO₂ atmosphere (the temperature of the 1 bar atmosphere increases with time due to the increasing solar luminosity; Gough, 1981). These relationships constrain the MAST, surface heat flux, and time relationships under which the ice-cemented cryosphere could have stabilized. Under MZ1 heat flow conditions (red line), the minimum MAST at 3.6 Ga is 227 K and minimum P_F is 850 mbar CO₂ atmosphere (273 K isotherm model) or 212 K and 390 mbar (252 K isotherm model). If the martian atmosphere at 3.6 Ga had at most a 1 bar CO₂ atmosphere (Kite et al., 2014), the maximum age of cryosphere stabilization occurs at ~3.3 Ga (273 K isotherm model). In the 252 K isotherm model, ICC stabilization is predicted to occur at the age in which MAST decreases to any point above the red line (likely near the Amazonian-Hesperian boundary based on the relatively cold climate believed to characterize the Amazonian period). Ages from Michael (2013) and Hartmann (2005).

273 K isotherm		Minimum	Minimum		
Heat flow limit	Q_F (mW/m ²)	MAST (K)	P_F (bar CO ₂)	ICC stabilization age	
MZ1	60*	227*	0.85*	3.6 Ga	<i>If ICC stabilized after</i>
RUr1	51	233	1.01		<i>Late Noachian-</i>
MZ2	42	238	1.16		<i>Hesperian boundary</i>
MZ1	53	Max 231	Max 1.00	3.3 Ga	<i>Latest age assuming 1 bar CO₂ atmosphere</i>
252 K isotherm		Minimum	Minimum		
Heat flow limit	Q_F (mW/m ²)	MAST (K)	P_F (bar CO ₂)	ICC stabilization age	
MZ1	60*	212*	0.39*	3.6 Ga	<i>If ICC stabilized after</i>
RUr1	51	217	0.56		<i>Late Noachian-</i>
MZ2	42	222	0.70		<i>Hesperian boundary</i>
ICC stabilization for the 252 K isotherm model occurs when the MAST falls below red line in Fig. 15. For example, if MAST at 3 Ga were less than 220 K (and CO ₂ atmospheric pressures less than 600 mbar), ICC stabilization would occur at 3 Ga.				3.0 Ga?	<i>Latest age assuming Amazonian MAST < 220 K</i>

Table 4. Best fit heat flow (Q_F), mean annual surface temperature (MAST), and atmospheric pressure (P_F) configurations for the MAST- Q_F least-squares fit temperature model (Fig. 15; from Eqs. 4-7) which allow the ICC to stabilize. The top three rows for both the 273 K isotherm model and the 252 K isotherm model show the minimum bound temperature and atmospheric pressure at 3.6 Ga, assuming the cryosphere freezing front reached the base of the ice-cemented cryosphere after 3.6 Ga. The bottom row shows the minimum bound age (and maximum temperature/pressure configuration) for ICC stabilization from Fig. 15. Ages from Michael (2013) and Hartmann (2005). * denotes the minimum bound Late Noachian temperature, pressure and heat flow configurations.

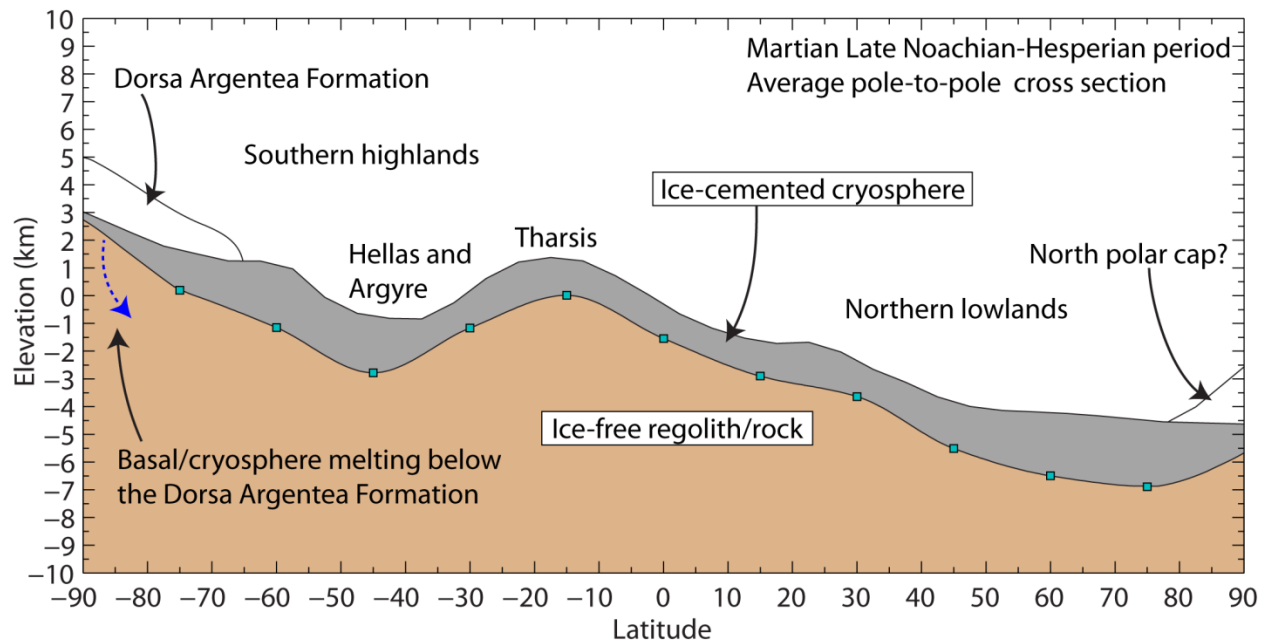


Figure 16. Generalized latitudinal relations for the ice-cemented cryosphere configuration between the Late-Noachian and Hesperian period when the Dorsa Argentea Formation was present and Mars may have had a higher atmospheric pressure. Elevation is from Fig. 5G. Green squares illustrate inferred ICC thicknesses from Fig. 3B. In the high southern latitudes the Dorsa Argentea Formation is predicted to raises the melting isotherm within the crust and produce melting at the base of the ICC (Section 6).

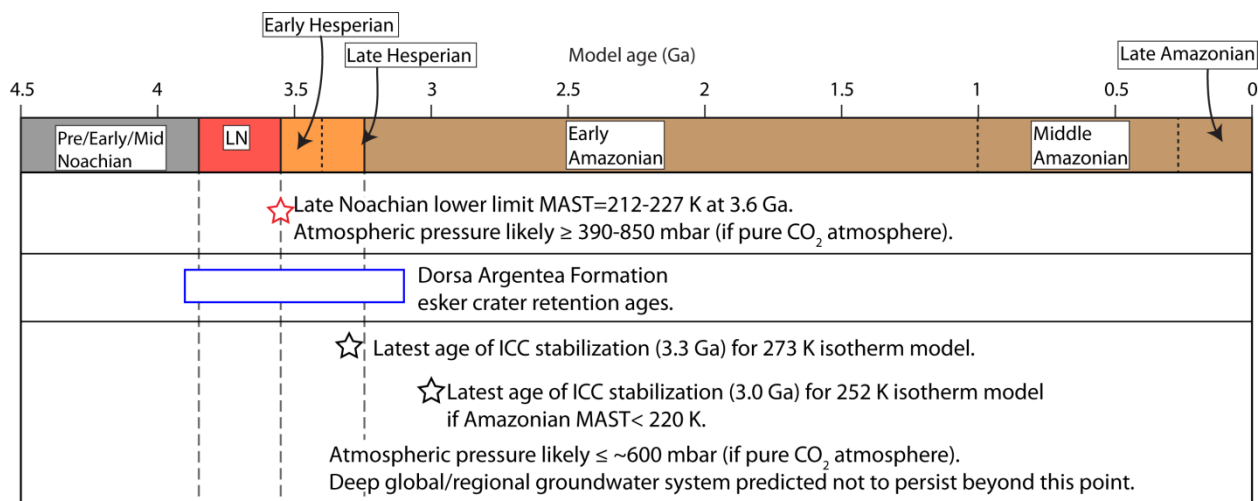


Figure 17. Geologic timeline illustrating the model results and chronology. Shown is the Late Noachian (LN) minimum MAST estimate from this study, the age of the Dorsa Argentea Formation crater retention ages from Kress and Head (2015), and the latest age of ice-cemented cryosphere stabilization from this study for the 273 K isotherm model (Fig. 15A) and the 252 K isotherm model (Fig. 15B). Model age is from Hartmann (2005) and Michael (2013).

Chapter 5:
Salt or ice diapirism origin for the honeycomb terrain in Hellas basin, Mars?: Implications for the early martian climate

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Abstract

The “honeycomb” terrain is a Noachian-aged cluster of ~7 km wide linear cell-like depressions located on the northwestern floor of Hellas basin, Mars. A variety of origins have been proposed for the honeycomb terrain, including deformation rings of subglacial sediment, frozen convection cells from a Hellas impact melt sheet, a swarm of igneous batholiths, salt diapirism, and ice diapirism. Recent work has shown that the salt or ice diapirism scenarios appear to be most consistent with the morphology and morphometry of the honeycomb terrain. The salt and ice diapirism scenarios have different implications for the ancient martian climate and hydrological cycle, and so distinguishing between the two scenarios is critical. In this study, we specifically test whether the honeycomb terrain is consistent with a salt or ice diapir origin. We use thermal modeling to assess the stability limits on the thickness of an ice or salt diapir-forming layer at depth within the Hellas basin. We also apply analytical models for diapir formation to evaluate the predicted diapir wavelengths in order to compare with observations. Ice diapirism is generally predicted to reproduce the observed honeycomb wavelengths for ~100 m to ~1 km thick ice deposits. Gypsum and kieserite diapirism is generally predicted to reproduce the observed honeycomb wavelengths for ≥ 600 -1000 m thick salt deposits, but only with a basaltic overburden. Halite diapirism generally requires $\geq \sim 1$ km thick halite deposits in order to reproduce the observed honeycomb wavelengths. Hellas basin is a distinctive environment for diapirism on Mars due to its thin crust (which reduces surface heat flux), low elevation (which allows Hellas to act as a water/ice/sediment sink and increases the surface temperature), and location within the southern highlands (which may provide proximity to inflowing saline water or glacial ice). The plausibility of an ice diapir mechanism generally requires temperatures ≤ 250 K within Hellas in order to reproduce

the observed diapir wavelength. Conversely, the viability of the salt diapir mechanism requires sufficiently thick evaporite deposits to accumulate in Hellas (generally ≥ 1 km), which requires the emplacement and evaporation within Hellas of a 14 to 2045 m global equivalent layer (GEL) of saline water ($\sim 2 \times 10^6 \text{ km}^3$ to $\sim 3 \times 10^8 \text{ km}^3$). On the basis of our analysis, we conclude that ice diapirism is more likely due to the thin deposits (~ 0.1 -1 km thick) and low water volumes required (only 0.3-24 m GEL water), and the potential for either glacial deposits or a frozen ocean to supply the necessary ice. Salt diapirism requires thick evaporite deposits and high water volumes by comparison, and thus appears less likely.

1. Introduction

The “honeycomb” terrain is a unique landscape on Mars that occurs on the floor of the Hellas Basin within the northwestern section of Hellas Planitia (Fig. 1A) and covers an area of $\sim 36,000 \text{ km}^2$ (Bernhardt et al., 2016a) (Fig. 1B). The honeycomb terrain is characterized by a dense cluster of oval to elongated cell-like depressions up to ~ 170 m deep (Fig. 1C) that each extend up to ~ 14 km on their long axis, and ~ 6 km on their short axis (Bernhardt et al., 2016a). The honeycomb terrain is interpreted to be Noachian in age (pre- ~ 3.7 Ga) on the basis of stratigraphic relationships with the adjacent wrinkle-ridged plains (Fig. 2), which appear to superpose the honeycomb terrain in certain locations and have a Hartmann (2005) model age of ~ 3.7 Ga (Bernhardt et al., 2016a, 2016b) (Fig. 2). The origin of the honeycomb terrain, however, remains enigmatic (e.g., Kerber et al., 2017). This terrain has been interpreted by different researchers to have formed through a variety of processes, summarized by Bernhardt et al. (2016a):

(1) **Deformation of subglacial sediment:** The honeycomb terrain has been proposed to be the imprints of grounded icebergs (comparable to terrestrial wallow pits, which form when an iceberg displaces/deforms the underlying sediment; e.g., Bigg, 2016) (Moore and Wilhelms, 2001). Bernhardt et al. (2016a) find this origin unlikely due to the wide distribution of sizes expected and smaller dimensions (widths less than ~100 m, depths less than 25 m) of these features compared to the relatively consistent (and larger) dimensions of the honeycomb cells.

(2) **Thermokarst origin:** Bernhardt et al. (2016a) explored a thermokarst origin for the honeycomb terrain, wherein loss of pore- or massive-ice in the subsurface by melting or sublimation generates shallow scallop-shaped depressions (which frequently overlap). While the widths of thermokarst features (up to ~15 km in diameter; e.g., Pewe and Journaux, 1983) appear to be consistent with the cells of the honeycomb terrain, Bernhardt et al. (2016a) do not favor a thermokarst origin based on the shallow depths (few tens of meters) of thermokarst holes and lack of overlap observed for the cells within the honeycomb terrain.

(3) **Impact melt convection** has been proposed to form the honeycomb terrain (Mangold and Allemand, 2003; Kite et al., 2009), wherein convection cells in an immediately post-impact Hellas basin impact melt sea formed and produced the honeycomb terrain when the convection cells cooled and froze. Bernhardt et al. (2016a) find this hypothesis unlikely because thick melt sheets are predicted to form a conductive lid that would prevent the observation of frozen convection cells at the surface (e.g., Cassanelli and Head, 2016a). Bernhardt et al. (2016a) also point out that preservation of convection patterns in impact melt is unlikely to occur because convection ceases before

a melt completely solidifies (e.g., Cassanelli and Head, 2016a). Furthermore, no cell-like features have been observed in the melt deposits of comparably large lunar impact basins (e.g., Bernhardt et al., 2016a; Vaughan et al., 2013).

(4) **Igneous diapirism** has been proposed to form the honeycomb terrain (Mangold and Allemand, 2003), wherein swarms of batholiths are intruded into the martian crust and are later exposed at the surface by crustal uplift and erosion. This hypothesis is not favored due to the lack of associated tectonic features in Hellas (Diot et al., 2016), and the observation by Bernhardt et al. (2016a) that terrestrial batholith swarms with pronounced topographic surface expressions have not been observed to occur in such regular dense assemblages.

(5) **Salt diapirism** has been previously explored (but not favored) as the origin for the honeycomb terrain (Mangold and Allemand, 2003; Kite et al., 2009; Diot et al., 2016). Although the dimensions and morphology of the honeycomb terrain are remarkably similar to examples of terrestrial salt diapirism (Fig. 1D and E) (e.g., see Fig. 9 in Bernhardt et al., 2016a and Fig. 8 in Fernandez and Kaus, 2015), these investigators suggested that the lack of brittle deformation features and the large volumes of water required to emplace the evaporitic deposits make a salt diapirism origin unlikely. Diot et al. (2016) noted, however, that brittle deformation features are not expected during “passive” downbuilding of diapirs, wherein diapirs propagate upwards at the same rate as they are buried by sediment (Jackson et al., 1994), and Bernhardt et al. (2016a) note that some terrestrial salt diapirism is associated with more ductile, rather than brittle surface deformation, and so the lack of brittle deformation features may not explicitly preclude a salt diapir origin. Bernhardt et al. (2016a) further suggest that the volume of water

necessary to produce the thick salt deposits may be as low as a ~ 3.5 m global equivalent layer (GEL) of water ($\sim 506,000 \text{ km}^3$), assuming fully saturated saline water, and that such a water volume could have been present and recycled throughout the Noachian period (e.g., Rosenberg and Head, 2015). Bernhardt et al. (2016a) conclude that a salt diapir origin for the honeycomb terrain remains a viable candidate formation hypothesis. They performed a preliminary assessment of the diapir-forming layer thicknesses required to produce the observed diapir wavelengths and found that a salt layer must be at least ~ 2 km thick and superposed by an overburden between ~ 2 to ~ 4 km thick to produce the observed diapir wavelengths.

We also note that the elongate morphology of the honeycomb terrain (Fig. 1C) is consistent with a diapir origin. Terrestrial salt diapirs are commonly elongate (Fails et al., 1995, pp. 27; Hudec and Jackson, 2007). Elongation can be caused by (1) the local stress field generated by pre-existing faults, (2) a specific tectonic regime, (3) underlying bed slope, or (4) salt thickness variations (Jackson and Talbot, 1986; Harding and Huuse, 2015), or alternatively (5) by variations in the sedimentation rate forming the overlying layer (Fernandez and Kaus, 2015).

(6) **Ice diapirism:** Diot et al. (2016) and Bernhardt et al. (2016a) assessed whether the honeycomb terrain could alternatively be formed by ice diapirism. While not observed on Earth (likely due to lack of ice confinement; Kite et al., 2009), ice diapirism has been interpreted to have occurred on Europa (Pappalardo et al., 1998; Rathbun et al., 1998; Pappalardo and Barr, 2004) and Triton (Schenk and Jackson, 1993), and could plausibly occur under martian conditions (Brand et al., 2008; Kite et al., 2009). Bernhardt et al. (2016a) suggested that an ice diapirism mechanism would require an ice layer ~ 1

km thick superposed by overburden deposits up to ~1 km thick in order to produce diapirism with the observed cell wavelength, on the basis of the reduced density of ice compared with salt (Brand et al., 2008). Bernhardt et al. (2016a) then assessed the viability of ice diapirism by determining whether sufficiently thick layers of ice are thermally stable in the martian subsurface. These authors concluded that an ice diapir-forming layer may be stable to depths up to ~2 km in the Hellas basin subsurface, and that ice diapirism is thus a viable candidate process to form the honeycomb terrain.

In summary, Bernhardt et al. (2016a) concluded that the honeycomb terrain in Hellas may be plausibly formed by either salt or ice diapirism, but they were unable to distinguish between the two scenarios. The salt diapirism hypothesis requires a climate with either temporary or prolonged warm conditions, and large volumes of saline water to flow into Hellas and then evaporate or freeze. The ice diapirism scenario, on the other hand, requires either a predominantly cold climate with intermittent warming periods to produce liquid water in Hellas (which would later freeze), or alternatively, a cold climate with a source of glacial ice to form massive ice to be buried and then eventually form the diapirs. The distinction between salt and ice diapirism in Hellas basin is important because all three origins have different (but major) implications for the ancient martian climate and hydrological cycle. In order to distinguish between a salt or ice diapirism origin for the honeycomb terrain, we expand upon the initial Bernhardt et al. (2016a) analysis by using updated values for (1) the ancient geothermal gradient within the Hellas basin, (2) the overburden thermal conductivity, as well as (3) a wider range of surface temperatures. We also reassess the relationships between diapir wavelength and diapir-forming layer/overburden thicknesses using updated numerical and semi-analytic models

(e.g., Fernandez and Kaus, 2015) to provide more specific constraints in combination with thermal modeling. Moreover, we consider the thermal stability of salt layers, due to their similarly low dehydration temperature in some cases (e.g., gypsum has a dehydration temperature of 363 K; Orstroff, 1964; Lager et al., 1984). Here, we reevaluate the thermal stability limits of ice and salt in the martian subsurface (as it relates to diapir cell wavelengths) in order to determine whether salt or ice diapirism is a more plausible candidate formation mechanism for the honeycomb terrain in the Hellas basin.

2. Thermal stability and diapir wavelength

In this study, we use thermal modeling of the martian subsurface to evaluate the thermal stability of salt and ice diapir-forming layers at depth on Mars. We then apply analytical models of diapir formation in order to evaluate the parameter space in which salt and ice diapirism is predicted to occur and is able to form cells with dimensions comparable to that observed for the honeycomb terrain (Fig. 1C).

The mechanism of salt or ice diapirism (Jackson et al., 1994) requires a density inversion, i.e., that a sufficient thickness of salt or ice is present beneath an overburden layer with a greater density. The density contrast between these two layers acts as a driving force within this gravitationally unstable configuration (Schultz-Ela et al., 1993), leading to upwelling of the less dense salt or ice layer (Fig. 3). Withdrawal basins (the topographic low between adjacent diapirs; Fig. 3) may then form as a result of downwelling sediment in between adjacent salt diapirs (e.g., Hudec et al., 2009; Peel, 2014). Because the melting temperature of ice/salt is reached at some depth below the

martian surface due to the geothermal gradient (e.g., Clifford et al., 2010), the depth of stability of ice/salt offers a constraint on the thickness of a diapir-forming layer (Fig. 3).

We assess whether sufficiently thick salt and ice layers are thermally stable in the martian subsurface. These results, in turn, can be used to determine the viability of a salt or ice diapir scenario for forming the diapirs interpreted to form the honeycomb terrain. If ice or salt is not stable in the martian subsurface to a depth sufficient to form the observed cell wavelengths, then a salt and/or ice diapirism mechanism can be ruled out. In order to evaluate this, we first assess the melting temperatures of diapir-forming materials (Section 2.1) used in the thermal profile (Section 2.2) to assess the depth at which diapir-forming layers remain stable.

2.1. Salt species and melting temperature

Although a number of salt species are candidates for evaporite deposits based on theoretical predictions (Catling, 1999; Tosca and McLennan, 2006; Toner et al., 2015), we focus our study on gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$), halite (NaCl), and kieserite ($\text{MgSO}_4 \cdot \text{H}_2\text{O}$) due to their repeated documentation on the martian surface (e.g., Langevin et al., 2005; Jensen and Glotch, 2011; Squyres et al., 2012; Weitz et al., 2013; Nachon et al., 2014; Ehlmann and Edwards, 2014 and references therein) and robust theoretical prediction as a volumetrically significant martian brine precipitate (Catling, 1999; Tosca and McLennan, 2006; Toner et al., 2015). Although some of these mineral assemblages have been detected on Mars in exposures up to several kilometers thick with no evidence of diapirism (e.g., in Valles Marineris; Mangold et al., 2008; Murchie et al., 2009), the lack of denser overburden and impurity of the deposits would have prevented any diapirism

from occurring in these cases. Of the common precipitates found by the former studies (Catling, 1999; Tosca and McLennan, 2006; Toner et al., 2015), gypsum has a relatively low dehydration temperature (363 K; Orstroff, 1964; Lager et al., 1984) and low thermal conductivity (Horai, 1971; Robertson, 1988), which may limit its stability at depth. Halite is also predicted to be a major component of any martian evaporite assemblage (Catling, 1999; Tosca and McLennan, 2006; Toner et al., 2015), but has a high melting temperature (1074 K; Akella et al., 1969) and a high thermal conductivity (Urquhart and Bauer, 2015), and so is predicted to remain stable at great depths in the martian subsurface. Hydrated magnesium sulfates such as kieserite have moderate dehydration temperatures (Chipera et al., 2006) and low thermal conductivities (Prieto-Ballesteros and Kargel, 2005), which may limit their stability at depth.

2.2. Thermal profile

In order to assess the thickness of thermally stable salt and ice layers in the martian subsurface, we find the temperature (T) distribution as a function of depth (Z) in the subsurface using the one-dimensional steady-state heat equation:

$$T_{(Z)} = T_{(Z-1)} + \frac{Q\Delta Z}{\kappa_{(Z)}} \quad (1)$$

where surface temperature, $T_s = T_{(Z=0)}$, Q is the geothermal heat flux (in W/m^2), and thermal conductivity is κ . The thickness of thermally stable ice or salt is defined as the zone in the subsurface where $T_{(Z)}$ remains below the melting isotherm; we use a melting isotherm of 273 K for pure water ice, 363 K for gypsum (Orstroff, 1964; Lager et al.,

1984), 1074 K for halite (Akella et al., 1969), and 600 K for kieserite (Chipera et al., 2006). We adopt the pure melting isotherm for ice (273 K), noting that the inclusion of salt will have only a minor effect on the depth of ice stability. For example, a melting isotherm of 270 K, which is only ~130 m shallower than the pure melting isotherm (for $Q=45 \text{ mW/m}^2$ and $\kappa=2 \text{ W/m K}$), is reached at 5 wt% NaCl (or 12 wt% of either NaClO_4 or $\text{Mg}(\text{ClO}_4)_2$; Chevrier et al., 2009); terrestrial seawater is 3.5 wt% by comparison. For the overburden thermal conductivity, we adopt the thermal conductivity model of Smoluchowski (1981), which relates the κ of the porous material to the κ of the solid rock (κ_0) and its porosity (Φ):

$$\kappa_{(z)} = \kappa_0(1 - \Phi_{(z)}^{\frac{2}{3}}) \quad (2)$$

We assume the sedimentary overburden deposit is produced from a basaltic protolith, and so we set κ_0 to the temperature-dependent thermal conductivity of basalt, which can be approximated by the following thermal conductivity relationship for ice (Clifford, 1993; Clifford et al., 2010), given by (Hobbs, 1974):

$$\kappa_{\text{basalt}(z)} = \frac{488.19}{T_{(z)}} + 0.4685 \quad (3)$$

The porosity as a function of depth ($\Phi_{(z)}$) is then found as (Clifford, 1993):

$$\Phi_{(z)} = \Phi_0 \exp\left(\frac{-z}{K}\right) \quad (4)$$

where Φ_0 is the porosity at the surface (we use $\Phi_0=0.3$), and K is the decay constant. Clifford (1993) adjusted the lunar porosity decay constant ($K_{Lunar}=6.5$ km) to martian gravity (g) ($K_{Mars} = K_{Lunar} \frac{g_{Lunar}}{g_{Mars}}$), which yielded a K value of 2.82 km. New results from the GRAIL mission suggest $K_{Lunar}=9.8$ km (Besserer et al., 2014), which, when adjusted for martian gravity, yields a value of $K_{Mars} = 4.28$ km. The thermal conductivity of a diapir-forming ice layer is found as (Yen, 1981):

$$\kappa_{(Z)} = 9.828 \exp(-0.0057 T_{(Z)}) \quad (5)$$

We use a constant value of $\kappa=1.25$ for gypsum (Horai, 1971; Robertson, 1988), and the temperature-dependent function for halite from Urquhart and Bauer (2015). For the thermal conductivity of kieserite, we adopt the temperature-dependent function of another hydrated magnesium sulfate, epsomite ($\text{MgSO}_4 \cdot 7\text{H}_2\text{O}$), from Prieto-Ballesteros and Kargel (2005). We find the thermally stable diapir-forming layer thickness (i.e., the depth below the overburden where $T_{(Z)}$ is less than the melting isotherm) as a function of overburden thickness for T_s ranging from 200 to 250 K. This temperature range encompasses the T_s range within Hellas basin for atmospheric pressures ranging from 8 mbar to 1 bar in recent Late Noachian 3D general circulation models with a pure CO_2 atmosphere and 100% humidity (Wordsworth et al., 2013). Plesa et al. (2016) used 3D numerical thermal models to evaluate the present-day heat flux variations across Mars, and found that Hellas is predicted to be a zone of low surface heat flux due to its thin crust (Zuber et al., 2000). We therefore adopt a Q of 45 mW/m^2 , which is a lower

estimate for the martian heat flux at 3.8 Ga (Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005). This value is ~70% of the upper estimate of the global average Q of ~65 mW/m² at 3.8 Ga (Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005), which is also the ratio between the present day Hellas basin surface heat flux and the present day global average of Plesa et al. (2016), and so is likely to be a valid Q estimate for Hellas during the Late Noachian period.

2.3. Diapir wavelength

Next, we explore the parameter space in which diapirism may produce the observed diapir dimensions in order to narrow down the candidate materials and parameter range. In order to establish the layer thickness relationships that reproduce the observed cell dimensions in the honeycomb terrain, we adopt recent scaling laws derived from a 2-D semi-analytical thick plate model (Kaus and Becker, 2007; Fernandez and Kaus, 2015), which relate diapir wavelength (λ) (Fig. 3) to the diapir-forming layer thickness (H_{diap} ; Fig. 3) and the overburden thickness (H_{overb} ; Fig. 3). Fernandez and Kaus (2015) find that this relationship varies depending upon the ratio of $\frac{H_{overb}}{H_{diap}}$, and define the parameter space in which each relationship is valid as its domain (D):

$$\frac{\lambda}{1.155H_{diap}} = 2.81 \quad (D1) \text{ for } \frac{H_{overb}}{H_{diap}} < DT_{1-2} \quad (6)$$

$$\frac{\lambda}{1.155H_{diap}} = 3.86 \left(\frac{H_{overb}}{H_{diap}} \right)^{\frac{1}{2}} \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{\frac{1}{6}} \quad (D2) \text{ for } DT_{2-3} > \frac{H_{overb}}{H_{diap}} > DT_{1-2} \quad (7)$$

$$\frac{\lambda}{1.155H_{diap}} = 3.05 \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{\frac{1}{3}} \quad (D3) \text{ for } \frac{H_{overb}}{H_{diap}} > DT_{2-3} \quad (8)$$

where η_{overb} is the viscosity of the overburden layer, and η_{diap} is the viscosity of the convecting layer (Fig. 3). We find the domain transition (DT) between the three domains as: $DT_{1-2} = 1.89938 \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{-0.36230}$ and $DT_{2-3} = 2.02338 \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{0.27120}$ (determined from the model results shown in Fig. 6 of Fernandez and Kaus, 2015). These relationships illustrate how diapir wavelength is linearly related to the diapir-forming layer thickness (eq. 6), as discussed in Bernhardt et al. (2016a) based on Turcotte and Schubert (2014). This relationship, however, holds only under certain conditions (under domain 1; D1), when the ratio of the overburden thickness to the diapir-forming layer thickness is less than the value of DT_{1-2} , which depends upon the ratio of the overburden viscosity to the diapir-forming layer viscosity. At lower values of $\frac{\eta_{overb}}{\eta_{diap}}$, the relationships relating λ to H_{diap} are more complex (eq. 7 and 8). We solve eq. 6-8 for λ at each combination of H_{overb} and H_{diap} (values between 0 and 5 km) for a temperature-dependent (and thus H_{overb} - and H_{diap} -dependant) value of η_{diap} (discussed in Section 2.5). An example of calculated domain transitions is shown in Fig. 5D-F.

These relationships place constraints on the layer thicknesses that will reproduce the observed cell wavelength in the honeycomb terrain, which can then be used to compare against the layer thicknesses predicted to be thermally stable in the subsurface. For comparison with these relationships, we next assess the layer thickness relationships that will allow diapirism to initiate.

2.4. Diapir initiation

We now seek to constrain the parameter-space further by establishing the layer thickness relationships in which diapirism is predicted to occur (i.e., what ratio of overburden thickness to salt/ice layer thickness is required to initiate diapirism?). We adopt additional scaling laws derived from the thick plate 2-D semi-analytical model (Fernandez and Kaus, 2015), which show that diapirism occurs when the amplification velocity (V_{amp} ; the speed at which a perturbation grows) at the transition time between D1 and D2 (t_{trans}) is greater than the sedimentation rate (V_{sed}): $V_{amp}(t_{trans}) > V_{sed}$, where V_{amp} and t_{trans} are given by:

$$V_{amp} = \frac{1}{3} H_{diap} \exp(q t_{trans}) q \quad (9)$$

$$t_{trans} = 1.067 \frac{H_{diap}}{V_{sed}} \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{-\frac{1}{3}} \quad (10)$$

The diapir growth rate (q), is given by (Fernandez and Kaus, 2015):

$$q \propto 0.16 \frac{H_{diap} \Delta \rho g}{\eta_{diap}} \quad (D1) \text{ for } \frac{H_{overb}}{H_{diap}} < DT_{1-2} \quad (11)$$

$$q \propto 0.15 \frac{H_{diap} \Delta \rho g}{\eta_{diap}} \left(\frac{H_{overb}}{H_{diap}} \right)^{-1} \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{-\frac{1}{3}} \quad (D2) \text{ for } DT_{2-3} > \frac{H_{overb}}{H_{diap}} > DT_{1-2} \quad (12)$$

$$q \propto 0.14 \frac{H_{diap} \Delta \rho g}{\eta_{diap}} \left(\frac{\eta_{overb}}{\eta_{diap}} \right)^{-\frac{2}{3}} \quad (D3) \text{ for } \frac{H_{overb}}{H_{diap}} > DT_{2-3} \quad (13)$$

where $\Delta \rho$ is the density difference between the overburden and the diapir-forming layer.

The density (ρ) of halite salt is taken as 2200 kg/m³ (e.g., Landolt-Boernstein, 1982;

Schultz-Ela et al., 1993; Fernandez and Kaus, 2015), gypsum salt as 2300 kg/m^3 (Williams-Stroud and Paul, 1997; Urai et al., 2008), kieserite as 2600 kg/m^3 (Urai et al., 2008), and the density of ice as 917 kg/m^3 (e.g., Cassanelli and Head, 2015). Fernandez and Kaus (2015) explored V_{sed} ranging from 0.1 cm/yr to 0.01 cm/yr , and found that $V_{sed}=0.01 \text{ cm/yr}$ produced diapirs with a more cellular and elongated morphology (see Fig. 8 in Fernandez and Kaus, 2015 for comparison with our Fig. 1C), in contrast to the more isolated, circular diapirs exposed at the surface under higher sedimentation velocities. On the basis of the cellular and elongated morphology observed within the Hellas honeycomb terrain (Fig. 1C), we adopt $V_{sed}=0.01 \text{ cm/yr}$.

We explore two overburden models in order to encompass the range of possible overburden configurations. The first model assumes an overburden composed entirely of sediment, where the viscosity is a constant 10^{19} Pa s (van Keken et al., 1993), and the density of the sediment overburden (ρ_{overb}) as a function of depth is found as $\rho_{overb(z)} = \rho_0(1 - \Phi(z))$, where ρ_0 is the density of the solid rock (we use $\rho_0=3000 \text{ kg/m}^3$ for a basaltic protolith). In this model, the thermal conductivity of the entire thickness of overburden is derived from eq. 2-4. We also evaluate a model where the overburden is composed of basalt on the basis of the superposing relationships between the wrinkle-ridged plains and the honeycomb terrain in Hellas (Fig. 2) reported by Bernhardt et al. (2016a, 2016b). Although wrinkle-ridged plains are formed by compressional stresses acting on a cohesive layer (and are thus not necessarily lavas), we proceed with the interpretation by Bernhardt et al. (2016b) that the wrinkle-ridged plains are basaltic in origin due to their concurrent ages with other circum-Hellas volcanic features (Williams et al., 2010). We adopt a typical terrestrial upper crustal viscosity value for the effective

viscosity of the basaltic layer (10^{22} Pa s; e.g., Wdowinski and Axen, 1992; Karner, 2004; Kaus et al., 2008; Meyer and van Wijk, 2015). This model adopts the thermal conductivity for basalt from eq. 3 and the density as ρ_0 for the basaltic layer. The model schematic illustrating the major variables is shown in Fig. 3. Next, we assess the viscosities of the salt and ice diapir-forming layers (η_{diap}) for input into eq. 6-8 and 10-13.

2.5. Viscosity of diapir-forming layer

We now find the viscosity (η_{diap}) of the salt diapir-forming layer as (van Keken et al., 1993):

$$\eta_{salt} = \frac{\rho T(Z) d^3}{20.85 \cdot M \cdot 10^{15} \cdot \exp\left(-\frac{24530}{RT(Z)}\right)} \quad (14)$$

where d , the grain size, is primarily dependent upon the evaporation rate of the overlying standing body of water (e.g., Dai et al., 2015), which is largely temperature-dependent (e.g., Linacre, 1977; Valiantzas, 2006) (larger grain sizes precipitate under lower rates of evaporation). In this case, in order for an evaporite layer of a given thickness to form, the standing body of water must remain above freezing for the duration of evaporation until the evaporite thickness is reached. Thus, in the case of the formation of thick evaporite deposits, Earth-like temperature conditions (where mean annual surface temperature is greater than 273 K) are required, and so we use $d=1$ cm, which is typical of terrestrial salt diapirs (van Keken et al., 1993; Davison et al., 1996; Miralles et al., 2000; Urai et al.,

2008). Alternatively, salt may precipitate out of solution during the progressive freezing of the water layer, which concentrates salt into the underlying unfrozen water. In this case, salt will precipitate if the water becomes saturated, and the grain size will depend on the freezing rate of the water. For simplicity, we consider the end-member scenario of warm Earth-like conditions and proceed using $d = 1$ cm. For the molar mass (M), we use 136.14 g/mol for gypsum, 58.44 g/mol for halite, and 138.38 g/mol for kieserite; R is the gas constant (8.314 J K⁻¹ mol⁻¹). This relationship (eq. 14) represents the shear viscosity of salt due to pressure solution (van Keken et al., 1993; Williams-Stroud and Paul, 1997; Urai et al., 2008), which is applicable for strain rates lower than $\sim 10^{-12}$ s⁻¹ (corresponding to typical diapiric strain rates in non-tectonic settings on the Earth; Heard, 1972; Jackson and Talbot, 1986), and provides results consistent with viscosity estimates from terrestrial diapirs (e.g., Table 3 in Mukherjee et al., 2010). For the viscosity of ice (η_{diap}), we adopt the Newtonian temperature dependent viscosity relationship:

$$\eta_{ice} = \eta_0 \exp \left[A \left(\frac{T_m}{T_{(Z)}} - 1 \right) \right] \quad (15)$$

where η_0 is the viscosity at the melting temperature (T_m) (we use $\eta_0 = 10^{14}$ Pa s, equivalent to a grain size of 0.5 mm; e.g., Barr and Showman, 2009), and A is a constant (we use 26, equivalent to an activation energy of 60 kJ mol⁻¹; Goldsby and Kohlstedt, 2001). We approximate the ice viscosity as Newtonian due to the low strain-rates expected during diapir formation (10^{-15} - 10^{-12} s⁻¹) (Heard, 1972; Jackson and Talbot, 1986), which cause the stress-dependence on η to be low compared with the temperature dependence. Under these circumstances, the viscosity of ice is known to behave as Newtonian, an assumption

that has widely been applied in studies of the icy satellites (e.g., Barr and Showman, 2009). In reality, ice behaves in a non-Newtonian manner, and so some external forcing may be required to initiate diapirism (e.g., Solomatov and Barr, 2007); this could include variations in ice or overburden layer thickness due to non-homogenous emplacement, basin loading-induced faulting, or alternatively erosion of the overburden through impact or aeolian processes. These various processes may not necessarily occur immediately following overburden deposition; therefore, the timing of diapir piercement to the surface would be dictated by the timing of the external forcing event in tandem with the sedimentation rate. We find the average viscosity of each layer (η_{diap}) as the harmonic mean (e.g., Schmeling et al., 2008; Cheria et al., 2009; Fuchs and Schmeling, 2013) of the entire thickness of the salt/ice layer.

2.6. Observed Diapir wavelength

In order to determine diapir wavelength (λ), we measured the distance between the center of individual cells within the honeycomb terrain (Fig. 4A and B shows a section of the measured honeycomb terrain). In total we measured $N=604$ diapir λ values; Fig. 4C shows a histogram of the data. We find the mean diapir λ to be 6.9 km, the median to be 6.6 km, and the standard deviation (σ) to be 2.3 km (Fig. 4C). The diapir λ range within 1σ of the mean is 4.6 km to 9.2 km. These values are in agreement with the diapir wavelengths measured by Bernhardt et al. (2016a), who found that the cell wavelengths were typically 5-10 km. In order to encompass the range of observed diapir wavelengths, we proceed in our analysis using λ values of 4 km to 10 km, and use these as bounding values for λ in eq. 6-8 to establish parameter fields in which diapirism would produce the

observed λ as a function of H_{overb} and H_{diap} .

On the basis of (1) the thermal stability constraints on the thickness of diapir-forming and overburden layers, (2) the layer thickness relationships which can produce the observed diapir wavelengths, and (3) the layer thicknesses which allow diapirism to initiate, we now have all the parameters necessary to assess the viability of a salt and/or ice diapir mechanism to form the Hellas honeycomb terrain.

3. Results and Discussion

Our model results for ice diapirism with a surface temperature of 200 K are shown in Fig. 5A. Ice is not thermally stable in the yellow region (ice thicknesses above 2.5 km and overburden thicknesses above 1.6 km) and will melt for increased ice or overburden thicknesses, and so the overburden and ice layer thicknesses in this field can be ruled out. In the upper gray region the predicted diapir wavelength is too large ($\lambda > 10$ km), and in the lower gray region the predicted diapir wavelength is too small ($\lambda < 4$ km), and so layer thicknesses in these fields can also be ruled out. The green region shows the zones in which diapirism is not predicted to initiate for a sedimentation rate of 0.01 cm/yr and can thus be ruled out. The identical case for gypsum diapirism is shown in Fig. 5B, and Fig. 5C shows the case for halite diapirism. The white field shows the “zone of consistency”, in which the salt/ice thickness is (1) thermally stable, (2) predicted to initiate diapirs, and (3) will produce diapirs of the correct wavelength. These relationships illustrate the point that by considering the layer thicknesses that are (1) thermally stable and (2) can reproduce the observed diapir wavelengths, we have drastically reduced the parameter space that is consistent with a diapirism origin for the honeycomb terrain in Hellas.

The increased viscosity, density, and higher melting isotherm of gypsum and halite relative to ice shifts the zone of consistency to higher diapir-forming layer thicknesses and higher overburden thicknesses (Fig. 5B and C). Thicker overburden layers decrease the thickness of salt/ice layers that remain stable for both ice and gypsum because the diapir-forming layer is present at a greater depth where temperatures are higher. This increased temperature with depth also causes thicker overburden layers to decrease the viscosity of the underlying salt/ice layer and thus causes the diapir-forming layer to generate the observed wavelength at reduced salt/ice thicknesses.

We find that the viscosity and thickness relationships (i.e., $\left(\frac{H_{overb}}{H_{diap}}\right)$ and $\left(\frac{\eta_{overb}}{\eta_{diap}}\right)$ in eq. 6-8) cause ice diapirism to largely remain in domain 1 and 2 (D1 and D2 in Fig. 5D). Gypsum diapirism appears to be divided between domain 1 and 3 (D1 and D3) within the parameter space explored in this study (Fig. 5E) due to its high density, which causes the diapir growth rate to be small relative to the sedimentation rate and generally eliminates D2 solutions. The extremely high melting isotherm of halite allows it to be thermally stable to melting throughout all depths explored in this study (Fig. 5C), but its relatively higher viscosity causes the wavelength relationships (i.e., $\left(\frac{H_{overb}}{H_{diap}}\right)$ and $\left(\frac{\eta_{overb}}{\eta_{diap}}\right)$ in eq. 6-8) to remain largely within domains 1 and 3 (Fig. 5C and F) (i.e., eq. 6 and 8 are used to solve for λ instead of eq. 7). Kieserite diapirism (not shown in Fig. 5) exhibits similar wavelength and domain relationships to gypsum. Its higher density (2600 kg/m³; Urai et al., 2008), however, causes the zone in which diapirism is not predicted (i.e., green zone in Fig. 5) to encompass the entire parameter space, therefore preventing kieserite diapirism from being viable under the conditions shown in Fig. 5.

In summary, our model results (Fig. 6) for ice diapirism (dotted black region), gypsum diapirism (red region), halite diapirism (cross-hatched blue region), and kieserite diapirism (zigzag region) are shown for surface temperatures of 200 K (left panels), 225 K (middle panels), and 250 K (right panels) for both the sediment overburden model (top panels) and the basaltic overburden model (bottom panels). The zones of consistency (from Fig. 5) are depicted as the shaded/textured regions within Fig. 6. We can now discuss in the following sections the results for these different models.

3.1. Ice diapir model results

The gray shaded/dotted regions in Fig. 6 show the layer thickness relationships where ice diapirism may plausibly form the honeycomb terrain. For $T_s=200$ K and the sediment overburden model (Fig. 6A), the thickness of ice is generally predicted to be inversely correlated with the overburden thickness. For example, this model would predict that an overburden thickness of 500 m would allow ice diapirism for ice thicknesses between ~300 m and ~1.3 km. Alternatively, an overburden thickness of 1500 m would allow ice thicknesses between ~100-200 m. In total, for ice diapirism to occur and produce the correct cell wavelength, overburden thicknesses must be less than ~3 km, and ice thicknesses must be between ~100 m and ~3.1 km (but generally below ~1.7 km for overburden thicknesses greater than ~350 m). For an increased surface temperature of $T_s=225$ K (Fig. 6B), the allowable overburden and ice thickness is reduced: the overburden thickness must generally be between ~50-100 m and ~1.5 km, and ice thicknesses must be between ~50 m and ~2.5 km, but generally below ~1.3 km for overburden thicknesses greater than ~200 m. Additionally, increasing the surface

temperature to $T_s=250$ K (Fig. 6C) further decreases the ice and overburden thicknesses that may produce the honeycomb terrain: the overburden thickness must generally be between ~ 50 m and ~ 700 m, and ice thicknesses must be between ~ 50 m and ~ 1 km. Mean annual surface temperatures greater than 250 K (corresponding to the mean annual surface temperature in Hellas for a 1 bar pure CO_2 atmosphere in the 3D general circulation models of Forget et al., 2013; Wordsworth et al., 2013; Horan and Head, 2016) (Fig. 7E and F) generally preclude ice diapirism in these models because sufficiently thick ice layers would not be thermally stable in the subsurface.

The models in which the overburden is composed of basalt predict similar results (Fig. 6D-F), but the increased density and viscosity of the basalt layer has the effect of decreasing the maximum ice and overburden thicknesses required to produce the correct wavelength, and increasing the minimum ice thickness required. Overburden thicknesses beyond ~ 500 m thick are not predicted to produce the observed diapirism. For $T_s=200$ K (Fig. 6D), overburden thicknesses are required to be between ~ 50 m and ~ 450 m, and ice thicknesses between ~ 300 m and ~ 1.8 km, but generally below ~ 1 km for overburden thicknesses greater than 150 m. For higher temperatures, $T_s=225$ K (Fig. 6E), overburden thicknesses are required to be between ~ 50 m and ~ 350 m, and ice thicknesses between ~ 100 and ~ 900 m. For a further increased surface temperature of $T_s=250$ K (Fig. 6F), overburden thicknesses are required to be between ~ 50 m and ~ 300 m, and ice thicknesses between ~ 100 m and ~ 400 m. Surface temperatures greater than 260 K prevent ice layers of the necessary thickness from being thermally stable in the subsurface. We also note that the emplacement of basaltic lava flows on surface ice deposits will melt some thickness of the ice deposit. Recent work has shown, for

example, that the emplacement of 500 m of lava (in 100 m increments) would melt a cumulative ~430 m of the underlying ice; 100 m of lava would melt ~190 m of ice, 200 m of lava would melt ~280 m ice, 300 m of lava would melt ~340 m ice, and 400 m of lava would melt ~390 m ice (eq. 5 in Cassanelli and Head, 2016b). Consequently, in order to produce diapirism with a basaltic overburden, ice thicknesses are required to be up to ~200-400 m greater than discussed above (Fig. 6D-F) to account for contact melting of the ice deposit following lava emplacement.

In summary, these results show that there is a wide range of parameter space in which ice diapirism is both thermally stable and able to reproduce the observed cell wavelength. Surface temperatures are generally required to be below 260 K in order to meet the ice stability and cell wavelength constraints. Ice thicknesses are generally required to be between ~100 m to ~1000 m (although this is highly temperature dependent), except in the cases of overburden thicknesses less than ~100-400 m, where thicker ice (up to ~3.1 km thick for $T_s=200$ K) may also reproduce the observed cell wavelength. The sedimentary overburden adopted in our models is denser and orders of magnitude more viscous than terrestrial glacial till (e.g., Paterson, 1981), and thus our models do not predict diapirism to occur for the relatively thin ($<\sim 1$ km) Amazonian-aged mid-latitude debris-covered ($<\sim 100$ m debris) ice deposits (e.g., Head et al., 2010) (which also have a lower $Q=\sim 20$ mW/m²; Montési and Zuber, 2003; McGovern et al., 2004; Solomon et al., 2005).

3.2. Gypsum diapir model results

We now show gypsum diapir model results (the shaded red regions in Fig. 6) for

comparison with the ice diapir models. The increased density and viscosity of gypsum compared with ice increases the required layer thicknesses that are predicted to reproduce the cell wavelength. The higher melting isotherm of gypsum (363 K; Orstroff, 1964; Lager et al., 1984) increases the maximum gypsum thickness that is thermally stable in the subsurface and reduces the temperature-dependence (and overburden thickness dependence) of a gypsum diapir-forming layer.

For the sediment overburden model and $T_s=200$ K (Fig. 6A), overburden thicknesses may range from ~ 2.3 km to ~ 5 km (generally $>\sim 3.2$ km), and gypsum thicknesses from ~ 600 m to 2.7 km (but generally between 0.6-1.2 km for overburden thicknesses greater than ~ 3.2 km). For an increased surface temperature of $T_s=225$ K (Fig. 6B), gypsum diapirism is viable for a much smaller parameter range: the overburden thickness ranges between ~ 3.3 km and ~ 3.6 km, and gypsum thicknesses range from ~ 700 m and ~ 900 m. The surface temperature of $T_s=250$ K (Fig. 6C) entirely eliminates gypsum as a viable diapirism candidate. Gypsum is highly temperature dependent in our models because its high density causes the zone in which diapirism is not predicted to occur to be large (e.g., green region in Fig. 5B); at high temperatures, this zone intersects with the zone in which gypsum is not thermally stable in the subsurface (e.g., yellow region in Fig. 5B).

The models in which the overburden is composed of basalt predicts a relatively larger parameter space in which the observed diapirism can occur due to the high density of basalt relative to gypsum. As in the case of ice diapirism, the increased density and viscosity of the basalt layer decreases the gypsum and overburden thicknesses required to produce the correct wavelength. For $T_s=200$ K (Fig. 6D), overburden thicknesses are required to be between ~ 50 -600 m, and gypsum thicknesses between ~ 1.2 km and ~ 3.1

km. For $T_s=225$ K (Fig. 6E), overburden thicknesses are required to be between ~50 m and ~550 m, and gypsum required thicknesses to be between ~900 m and ~3.1 km. For an increased surface temperature of $T_s=250$ K (Fig. 6F), overburden thicknesses are required to be between ~50 m and ~500 m, and gypsum required thicknesses to be between ~700 m and ~3.1 km.

In summary, these results show that there is a modest range of parameter space in which gypsum diapirism is both thermally stable and able to reproduce the observed cell widths, but only for a basaltic overburden, or $T_s=200$ K with a sedimentary overburden. Under the conditions where gypsum is a viable diapirism candidate, the parameter space is shifted relative to that of ice diapirism for increased gypsum and overburden thicknesses. Gypsum thicknesses are generally required to be between ~0.6-1.0 km and ~3 km. Due to the high density and low dehydration temperature of gypsum, increasing the surface temperature above 200 K generally prevents the thermal stability of sufficiently thick gypsum layers in the subsurface for a sedimentary overburden, and therefore gypsum diapirism is a candidate process only in the scenario of a cold early Mars climate unless the overburden is composed of higher density basalt.

3.3. Kieserite diapir model results

The kieserite diapir model results are shown as the zigzag-textured region in Fig. 6 for comparison with the ice, gypsum, and halite diapir models. The increased density of kieserite compared with both ice and the other salts restricts the viable parameter space for kieserite diapirism because the sedimentation rate commonly exceeds the diapir growth rate. The higher melting isotherm (600 K; Chipera et al., 2006) and lower thermal

conductivity (Prieto-Ballesteros and Kargel, 2005)) of kieserite have the net effect of yielding similar thermal stability results to gypsum.

For the sediment overburden model, kieserite is not a viable diapir-forming material under any range of parameters or surface temperatures due to its exceptionally high density, which causes the sedimentation rate to exceed the diapir growth rate. For the models in which the overburden is composed of basalt, the results are similar to gypsum diapirism but the maximum overburden and minimum kieserite thicknesses are restricted to slightly lower values. For $T_s=200$ K (Fig. 6D), overburden thicknesses are required to be between ~50-450 m, and kieserite thicknesses between ~950 m and ~3.1 km. For $T_s=225$ K (Fig. 6E), overburden thicknesses are required to be between ~50 m and ~520 m, and kieserite requires thicknesses to be between ~800 m and ~3.1 km. For an increased surface temperature of $T_s=250$ K (Fig. 6F), overburden thicknesses are required to be between ~50 m and ~400 m, and kieserite requires thicknesses to be between ~600 m and ~3.1 km.

In summary, these results show that there is a modest range of parameter space in which kieserite diapirism is both thermally stable and able to reproduce the observed cell widths, but only for a basaltic overburden. Under these conditions, kieserite diapirism behaves similarly to gypsum diapirism; kieserite thicknesses are generally required to be between ~0.6-1.0 km and ~3 km. Due to the high dehydration temperature of kieserite, it is relatively insensitive to the temperature conditions, and therefore (with gypsum) is a candidate process in both cold and warm early Mars climate scenarios if the overburden is composed of basalt.

3.4. Halite diapir model results

The halite diapirism model results are shown in the shaded blue cross-hatched regions (Fig. 6) for comparison with the ice, gypsum, and kieserite diapir models. Halite has a higher viscosity than ice, gypsum, and kieserite, which increases the required halite layer thicknesses that are predicted to reproduce the cell wavelength. The high thermal conductivity and melting isotherm of halite (1074 K) drastically reduce the temperature-dependence of the halite layer thicknesses on the cell wavelength. The high melting isotherm of halite also allows halite to be thermally stable across all of the depths explored in this study, and so its viability as a diapir-forming layer to form the honeycomb terrain in Hellas depends only on its ability to initiate diapirism and form the correct cell wavelength rather than thermal constraints.

For the sediment overburden model and $T_s=200$ K (Fig. 6A), overburden thicknesses may range as low as ~ 1.4 to 2 km (for halite thicknesses from 1.4-3 km) and above 5 km, depending on the halite thickness. Halite thicknesses generally ranging from ~ 1.4 km to ~ 3.1 km; the minimum halite thickness is reduced to ~ 750 m at overburden thicknesses approaching 5 km. Increasing the surface temperature to 225 K (Fig. 6B) has the minor effect of decreasing the minimum halite thicknesses to ~ 1.2 km. In this case, thinner halite layers (down to ~ 600 m) are viable at high overburden thicknesses greater than ~ 2.5 km. The surface temperature of $T_s=250$ K predicts a similar result (Fig. 6C), where halite diapirism occurs for overburden thicknesses between ~ 1.1 km and above 5 km, and halite thicknesses generally between ~ 1 km and ~ 3.1 km. In this case, overburdens thicker than ~ 2.5 km allow halite layers as thin as ~ 500 m.

For models in which the overburden is composed of basalt, a much smaller range is

predicted where halite diapirism can produce the correct cell wavelength. For $T_s=200$ K, halite diapirism requires overburden thicknesses to be between ~ 50 m and ~ 1.3 km, and halite thicknesses between ~ 1.8 and 3.1 km (Fig. 6D). Increasing the surface temperature to $T_s=225$ K (Fig. 6E) allows halite diapirism to occur for overburden thicknesses between ~ 50 - 900 m and halite thicknesses between ~ 1.2 km and ~ 3.1 km. For the higher surface temperature of $T_s=250$ K (Fig. 6F), overburden thicknesses are required to be between ~ 50 m and ~ 650 m, and halite thicknesses between ~ 1 km and ~ 3.1 km.

In summary, these results show that there is a wide range of parameter space in which halite diapirism is able to reproduce the observed cell wavelengths, but the parameter space is shifted relative to that of ice, gypsum, and kieserite diapirism to increased thicknesses of both halite and overburden. Due to its high melting temperature, halite is stable within all depths explored in this study and is a candidate diapir-forming material in the scenario of a much warmer ($T_s \geq 273$ K) early Mars climate. Halite thicknesses are generally required to be between ~ 1 to ~ 3.1 km, with overburden thicknesses ranging from ~ 50 m up to 5 km.

3.5. Summary of model results

Our models suggest that a wide range of layer thickness relationships permit salt and ice diapirism, but make several predictions on the basis of the varying densities, viscosities, and melting temperatures of the different candidate diapir-forming materials (Fig. 6). Ice diapirism is predicted to require ice thicknesses between ~ 100 m and ~ 1 km (and a comparably thick overburden), and requires mean annual surface temperatures in Hellas to remain at or below ~ 250 K. Gypsum and kieserite diapirism generally require

salt thicknesses of ~0.6-1.0 km to ~3 km (with an overburden generally between than ~50-600 m), and only for the basaltic overburden case. Halite diapirism generally requires halite thicknesses between ~1 km and ~3 km (with an overburden thicker than ~50 m to ~1 km). The high melting isotherm of halite allows diapirism to occur at much warmer surface temperatures (i.e., >273 K) compared to ice. Gypsum and kieserite are also relatively insensitive to the surface temperature as long as the overburden is composed of basalt. The basaltic overburden models (Fig. 6D-F) generally serve to drastically reduce the layer thickness relationships that are predicted to form diapirism with the correct wavelength. A basaltic overburden is favored by Bernhardt et al. (2016a), but it remains uncertain whether the diapirs interpreted to compose the honeycomb terrain formed before the emplacement of the volcanic plains, or after (and in response to) the emplacement and superposition of the volcanic plains (Fig. 2). The lack of brittle deformation features within the honeycomb terrain suggest that a basaltic overburden may be unlikely unless the lava flows composing the overburden were emplaced periodically (with a deposition rate comparable to the diapir growth rate) to enable the diapirs to grow through downbuilding (i.e., the diapir is propagating upwards at a rate comparable to overburden deposition).

4. Location exclusively within Hellas

There is abundant evidence for surface and subsurface ice deposits as well as fluvial activity throughout the geologic history of Mars (e.g., Carr and Head, 2010). Why then might evidence for diapirism be located exclusively within Hellas basin and nowhere else on the surface of Mars? We note that the Hellas basin floor is likely to be a unique

environment for diapirism due to:

(1) Its thin crust resulting from basin excavation (Zuber et al., 2000), which is predicted to reduce its surface heat flux (Plesa et al., 2016) due to radiogenic elements sequestered in the crust having been removed. This allows thick ice or gypsum layers to remain thermally stable in the subsurface at greater depths and surface temperatures.

(2) Its low elevation (Fig 1A), which could allow Hellas to act as a sedimentary basin which would be able to accumulate salt deposits derived from the evaporation or freezing of standing water. Furthermore, standing water could also freeze to form an ice diapir-forming layer. Finally, the basin is a preferred location for accumulation of sediment/lava cover to act as the overburden. Several studies have suggested that the Hellas basin contained standing water from inflowing groundwater (Andrews-Hannah et al., 2010; Andrews-Hannah and Lewis, 2011) or an ocean during the Noachian period (Malin and Edgett, 2000; Clifford and Parker, 2001; Moore and Wilhelms, 2001; Crown et al., 2005; Wilson et al., 2007; Di Achille and Hynek, 2010).

The low elevation also increases its surface temperature for atmospheric pressures greater than a few tens of millibars (Forget et al., 2013; Wordsworth et al., 2013), which allows the Hellas basin floor to be among the warmest regions on early Mars (Fig. 7). This limits ice and gypsum stability at depth, but would also serve to prevent subsurface temperature-sensitive diapir-forming layers (namely ice) from being too thick (or too viscous) to reproduce the observed cell wavelength.

(3) Its location within the southern highlands, which is proximal to Late Noachian-aged fluvial activity (e.g., Fassett and Head, 2008; Hynek et al., 2010) that could supply water into the basin, and nearby snow and ice deposits in a cold and icy climate scenario

(Fig. 14 in Head and Marchant, 2014; Wordsworth et al., 2013; 2015; Fastook and Head, 2015). Snow is not predicted to accumulate within Hellas in such a cold climate (Wordsworth et al., 2013; 2015), but could be transported into the basin by glacial flow from the Hellas basin rim (Fastook and Head, 2016).

(4) Its proximity to an ancient south circumpolar ice sheet, the Dorsa Argentea Formation, which could have flowed down the basin wall topographic slope and into Hellas (Scanlon et al., 2016; Fastook and Head, 2016), providing a source of relatively pure ice for a diapir-forming layer. Recent 3D global climate modeling shows that this scenario may occur only under certain climatic conditions, with a pure CO₂ atmosphere between 0.6-1 bar and 42° obliquity (Scanlon et al., 2016) (Fig. 7D and F). Scanlon et al. (2016) found that warmer atmospheres prevent ice from being stable in Hellas, and lower obliquities (Fig. 7A-E) prevent ice layers from accumulating to sufficient thicknesses to flow glacially into Hellas (Fastook and Head, 2016).

4.1. Water volume required for diapirism

How much water is required to generate a sufficiently thick salt layer? Bernhardt et al. (2016a) argued that an absolute minimum of ~3.5 m GEL of hypersaline water (assuming ~30-35 wt% salt) would be required to form 2 km thick evaporite deposits based on the ~36,000 km² area of the honeycomb terrain. We note that this approach will underestimate the minimum volume of water required to form salt deposits because the saline water must fill an equipotential surface to at least the elevation of the honeycomb terrain. For example, the shaded black region in Fig. 8A shows an equipotential surface of -6300 m below the datum. This region (~1,570,000 km²) represents the minimum

water level required to deposit salt across the entire area of the honeycomb terrain (a body of standing water at least $\sim 623,000 \text{ km}^3$ or 0.004 m GEL). Adopting our salt thickness range estimate of ~ 1 to ~ 3 km (Fig. 6), the minimum required cumulative GEL of water to deposit the salt is 682-2045 m (3.5 wt% salt, similar to terrestrial seawater) or 119-358 m (for a brine with 20 wt% salt, close to the 23.3 wt% NaCl required for a eutectic solution of NaCl).

We also present an alternative scenario, which assumes that the modern-day topography is not representative of the Hellas basin during the deposition of salt on ancient Mars. The shaded black region in Fig. 8B shows an equipotential surface of -7000 m below the datum. Although this area has incomplete coverage ($\sim 85\%$) over the honeycomb terrain mapped by Bernhardt et al. (2016b) (white region in Fig. 8), we consider this case due to the uncertain lateral boundaries of the honeycomb terrain (which may or may not extend beneath the surrounding plains units), and because the emplacement of the wrinkle-ridged plains within Hellas could have altered the basin floor topography during loading and subsidence (e.g., Solomon and Head, 1980). This region (shaded black zone in Fig. 8B; $\sim 188,000 \text{ km}^2$) represents the minimum water level required to deposit salt across $\sim 85\%$ of the honeycomb terrain (a body of standing water at least $\sim 44,200 \text{ km}^3$ or 0.0003 m GEL). Assuming salt deposits ~ 1 -3 km thick (Fig. 6), the minimum required cumulative GEL of water to deposit the salt is 82-245 m (for 3.5 wt% salt) or 14-43 m (for 20 wt% salt).

Note that these water volumes represent the cumulative volume of water required to deposit the salt layers, and not the volume of water required on the surface of Mars at any given time (i.e., a much smaller volume of water could be recycled numerous times). For

frame of reference, Rosenberg and Head (2015) estimated that a cumulative water volume of 3-100 m GEL was required to form the valley networks during the Late Noachian period, and Carr and Head (2015) estimated the current surface and near-surface water budget of Mars to be ~34 m GEL, and ~24 m GEL during the Late Noachian to Early Hesperian.

For comparison, an ice diapir layer between 100 m and 1 km thick would require only 0.05-0.5 m GEL of water within the mapped area of 36,000 km². If the ice was a remnant from a frozen ocean it would also be predicted to follow an equipotential surface: for the -6300 equipotential surface, the minimum required GEL of water required is 2.4-24 m, or 0.29-2.9 m for the -7000 m equipotential surface.

5. Conclusions

The honeycomb terrain located in the northwestern floor of Hellas Basin on Mars (Fig. 1) has recently been proposed to result from either salt or ice diapirism (Fig. 3). Hellas basin is a unique environment for diapirism on Mars due to its thin crust (which reduces surface heat flux), low elevation (which increases the surface temperature and allows Hellas to act as a water/ice/sediment sink), and location within the southern highlands (which could provide proximity to inflowing saline water or glacial deposits). The ice diapirism scenario implies that the ancient martian climate was cold ($\leq \sim 250$ K in Hellas at the time of diapir formation), and had a source of massive ice that was subsequently buried by a sediment and/or lava layer. The salt diapirism hypothesis, on the other hand, requires large volumes (on the order of 14 to 2045 m GEL water) of moderate to hypersaline waters to flow into Hellas and evaporate or freeze. Both models

have significant implications for the ancient martian climate and hydrological cycle, and so determining whether the honeycomb terrain is formed by salt or ice diapirism is an important distinction.

We find that both salt and ice diapir origins for the Hellas basin honeycomb terrain remain consistent with our current knowledge of the parameter range of salt/ice/overburden thicknesses and diapir wavelengths (Fig. 6). Halite, gypsum, and kieserite remain viable options to generate salt diapirism with the observed cell wavelength, but their layer thickness are required to be substantially thicker than an ice diapir-forming layer due to their increased viscosity and density. Furthermore, the high densities of gypsum and kieserite generally require a higher density, basaltic overburden to remain viable candidates for diapir formation. The consistency of salt diapirism in our analysis is more dependent upon the ability for sufficiently thick evaporite deposits to accumulate in Hellas (generally $>\sim 1$ km) and the composition of the overburden, than on specific climatic conditions (i.e., surface temperature) (Fig. 6). The increased thicknesses of halite, gypsum, and kieserite ($>\sim 1$ km) required relative to that of ice (~ 100 m to ~ 1 km) (Fig. 6) necessitate the deposition of thick sequences of evaporite deposits from a voluminous source of water (14 to 2045 m GEL water). This water volume estimate overlaps (but generally greatly exceeds) the range of recent estimates of water required to form the valley networks in the Late Noachian period (3-100 m GEL; Rosenberg and Head, 2015). We consider ice diapirism as a more attractive mechanism to form the honeycomb terrain in the Hellas basin because it requires only 0.3-24 m GEL of water (similar to the global estimate range of Car and Head, 2015). On this basis, a sedimentary overburden may be more likely because it offers a wider parameter space to remain a

viable candidate for ice diapirism. Although the parameter space for ice diapirism appears to be more restricted than salt (Fig. 6), the layer of ice required to form an ice diapir-forming layer could have accumulated from a greater number of sources. For example, a diapir-forming ice layer on early Mars could be sourced from either a frozen ocean in Hellas (Fig. 8), or from downslope glacial flow of regional surface ice (e.g., Fastook et al., 2012, Wordsworth et al., 2013, 2015; Fastook and Head, 2015, 2016; Scanlon et al., 2016).

Future work is required to assess the climatic implications of salt or ice diapirism within Hellas. Specifically, the physical plausibility of glacial flow of either regional surface ice deposits (Fastook and Head, 2015, 2016) or the Dorsa Argentea Formation (Scanlon et al., 2016) into Hellas requires further consideration. For example, if glacial flow into Hellas is found to be plausible (Fastook and Head, 2016), an ice diapir origin in a cold and icy climate would seem more likely. If glacial flow into Hellas is found to be difficult to achieve, the hypothesis of a Hellas ocean to generate ice diapirism (if frozen in a cold climate) or salt diapirism (if liquid in a warm climate) is supported.

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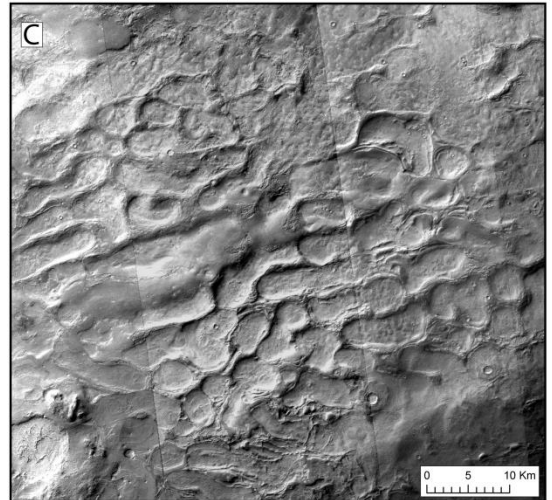
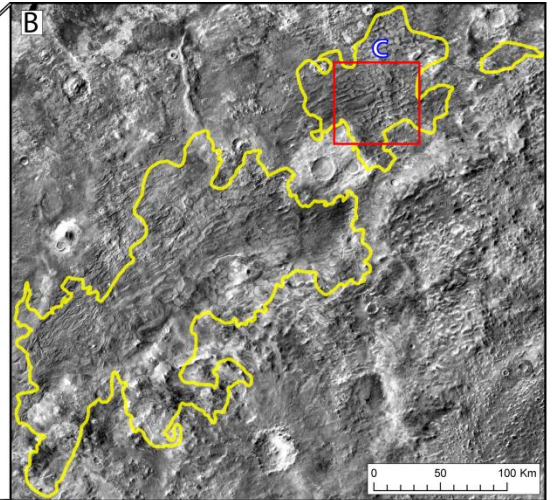
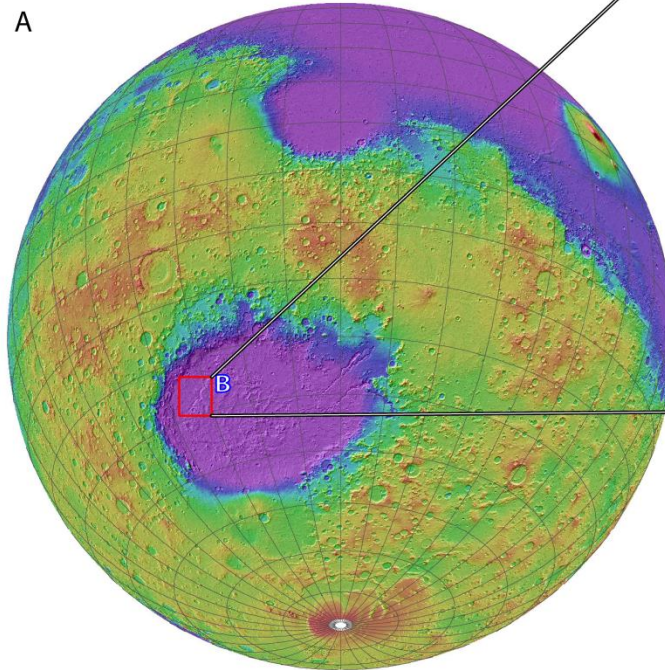
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Figures, tables, and captions:

Mars, honeycomb terrain (58.0°E, 34.7°S)

A



Earth, Sigsbee Nappe (92.2°W, 26.9°S)

D

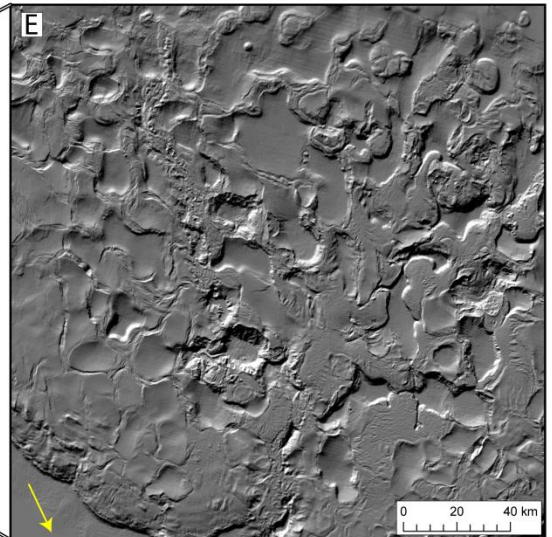
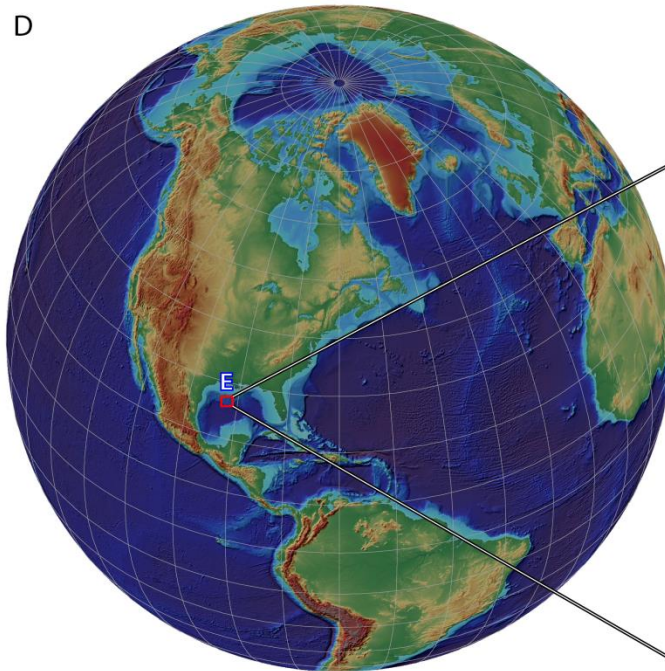


Figure 1. A) The honeycomb terrain is located within the northwestern section of the

Hellas Basin, Mars. The red box shows the location of panel (B). B) The yellow outline shows distribution of the honeycomb terrain reproduced from Bernhardt et al. (2016a) overlain on a THEMIS global daytime mosaic. The red box shows the location of panel (C). C) Enlargement of the honeycomb terrain, showing detailed morphology. D) The location of the Sigsbee Nappe (an allocthononous unit formed by the gradual advance of salt from its landward edge; Rowan et al., 1997) off the coast of the Gulf of Mexico on Earth. The red box shows the location of panel (E). E) Enlargement of the southern Sigsbee Nappe, a terrestrial terrain formed by salt diapirism; bathymetric shaded relief map from the U.S. Coastal Relief Model (CRM) by the National Oceanic and Atmospheric Administration (NOAA). Illumination direction is depicted as the yellow arrow in the bottom-left corner. CTX images in (C): P19_008361_1443, P18_008216_1443, P17_007860_1462, P18_008005_1460.

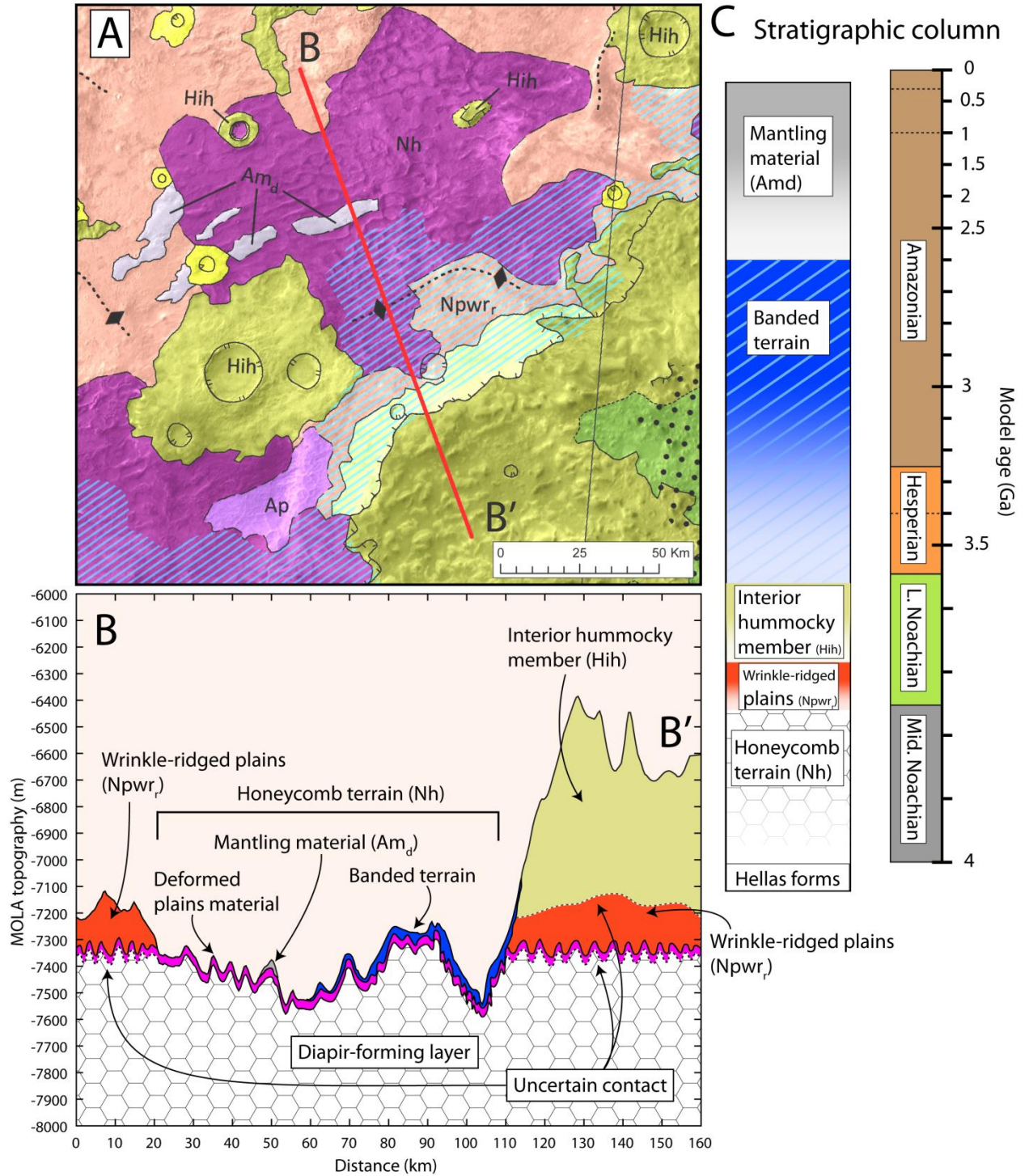


Figure 2. Geologic context of the honeycomb terrain. (A) Portion of the northeastern section of the honeycomb terrain from the geologic map of Bernhardt et al. (2016b). Units shown are the honeycomb terrain (Nh), the banded terrain (shown as diagonal blue lines), mantling material (Amd), the interior hummocky member (Hih), and wrinkle-ridged plains (Npwr). See Bernhardt et al. (2016b) for descriptions and interpretations of

these different units. Bernhardt et al. (2016a, b) interpret the wrinkle-ridged plains as basaltic in origin and favor this material as the overburden for the diapir-forming layer; this unit has alternatively been interpreted as sediment deposited in standing water (Moore and Wilhelms, 2001). (B) Mars Orbiter Laser Altimeter (MOLA) (Smith et al., 1999) topographic profile in (A) from B to B' with interpreted stratigraphy from Bernhardt et al., (2016a); dashed lines indicate an uncertain contact. Panels (A) and (B) are adapted from Fig. 10 of Bernhardt et al. (2016a). (C) Stratigraphic column depicting the superposition relationships and model ages of the units shown in panel (A) and (B), adapted from Fig. 6 of Bernhardt et al. (2016b). Dashed lines refer to boundaries between Late Amazonian, Middle Amazonian, Early Amazonian Early Hesperian, and Late Hesperian. Model ages are from Bernhardt et al. (2016b); banded terrain model age is from Diot et al. (2014); Hellas age from Fassett and Head (2011). All model ages and epoch boundaries use the Hartmann (2005) production function from Michael (2013).

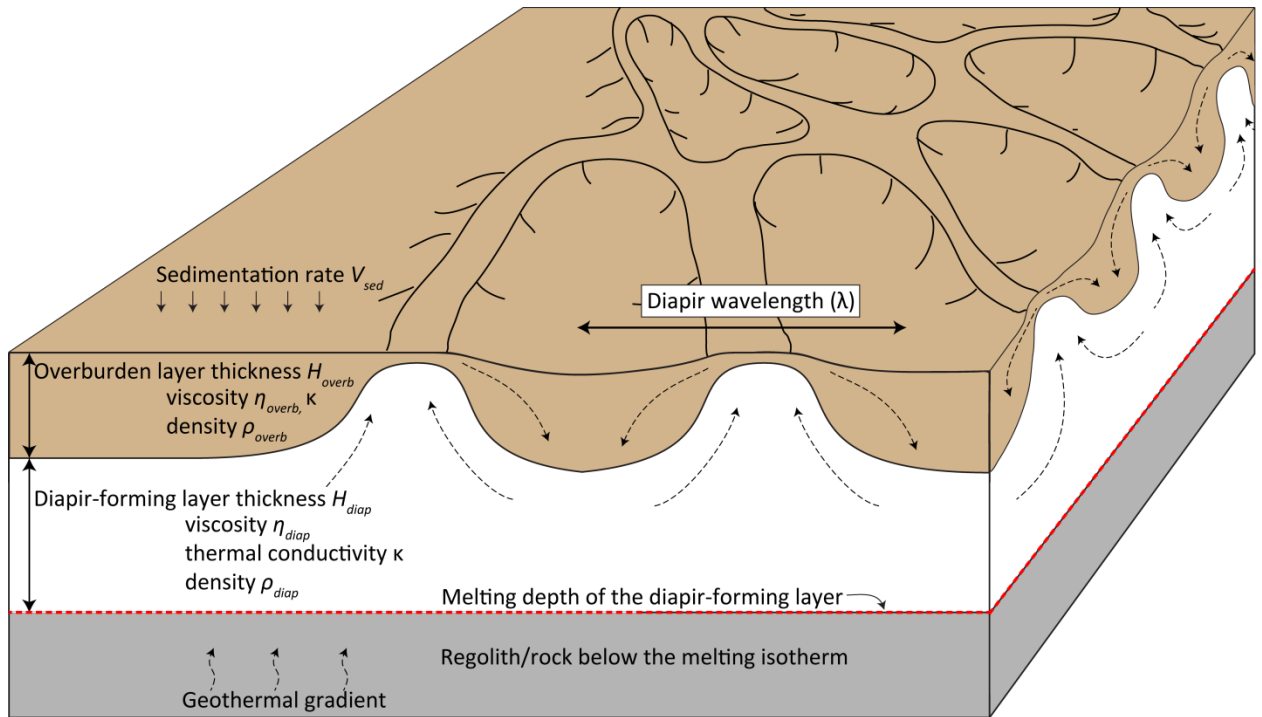


Figure 3. Schematic diagram showing the geometry of diapirism and the thermal/diapir wavelength model configuration and variables. The upper tan layer is the overburden, the middle white layer is the salt or ice diapir-forming layer, and the bottom gray layer is the regolith/rock below the diapir-forming layer. The decreased density of the diapir-forming layer relative to its overlying overburden causes the diapir-forming layer to upwell and form diapirs; the negative diapirs (withdrawal basins) between individual diapirs are formed by the denser downwelling overburden (dashed arrows). In our models, the maximum thickness of the diapir-forming layer is determined by its predicted depth of melting (dashed red line), which is governed by the local geothermal gradient.

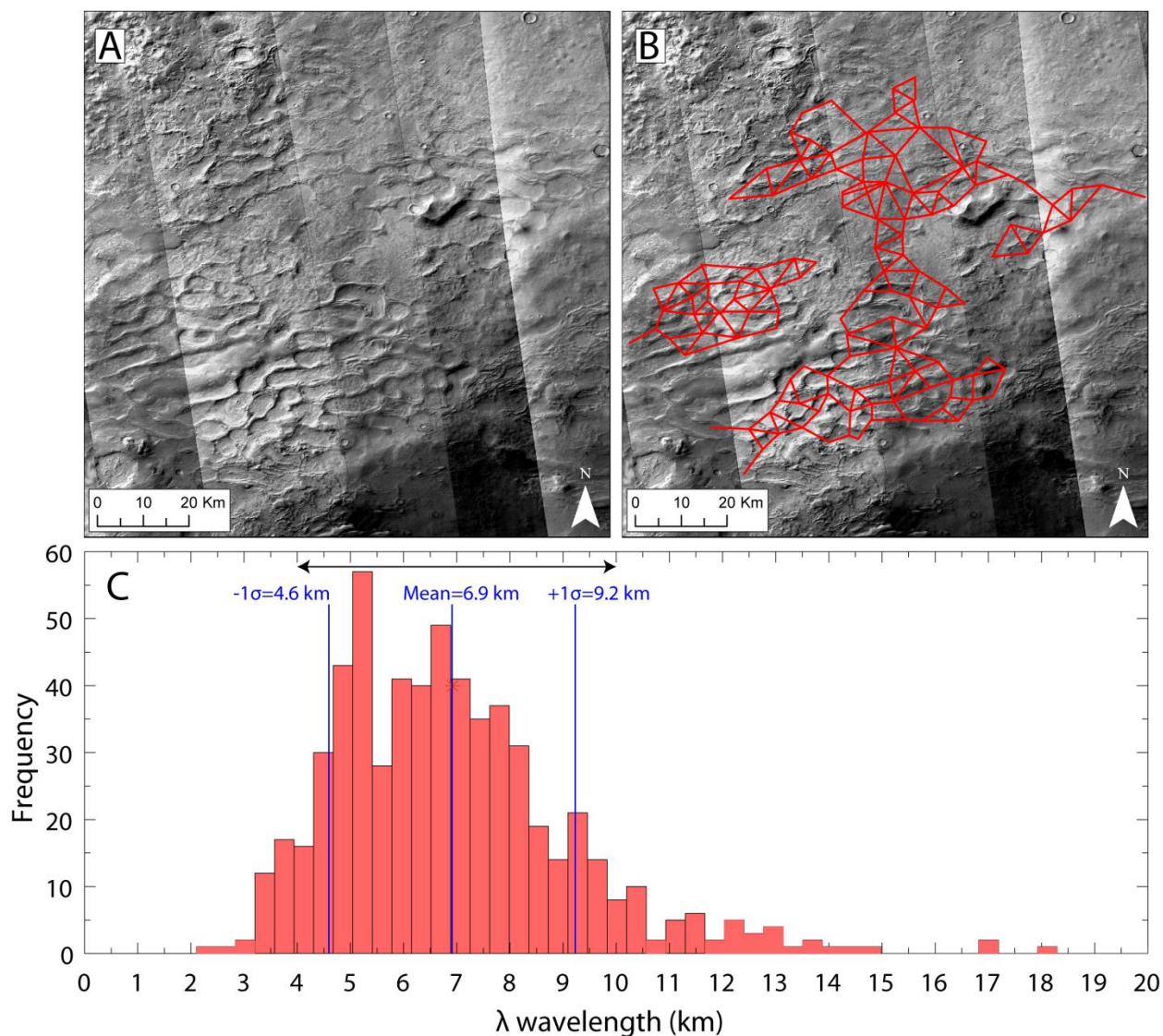


Figure 4. (A) A sample section of the honeycomb terrain in Hellas basin (50.65° E, 34.78° S). (B) The wavelength of the cells within the honeycomb terrain are measured as the distance between the centroid of adjacent cells (distance depicted as red lines). (C) Histogram of the diapir wavelength (λ) measurements made in this study (N=604). The mean is 6.9 km, the median is 6.6 km, and the standard deviation (σ) is 2.3 km. The black arrow shows the wavelength range explored in our study (4 to 10 km). CTX images in (A) and (B): P19_008361_1443, P18_008216_1443, P17_007860_1462, P18_008005_1460, P17_007504_1434.

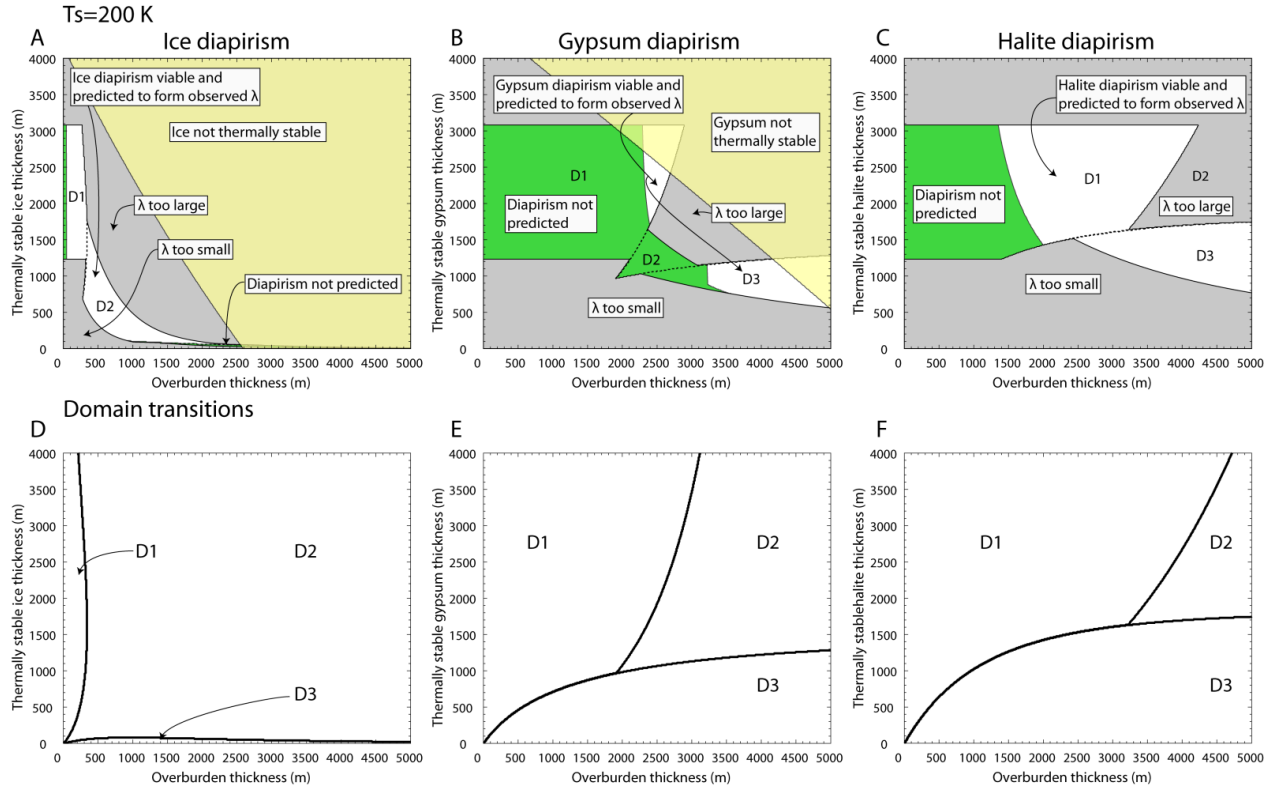


Figure 5. Model results showing the thermally stable diapir-forming layer thickness as a function of overburden thickness for a surface temperature of 200 K in the sediment overburden model. (A) Ice diapirism. (B) Gypsum diapirism. (C) Halite diapirism. The yellow regions show the zones in which melting occurs and the diapir-forming layer is not thermally stable in the subsurface, and thus diapirism is not predicted. Gray regions are zones in which diapirism is not predicted to produce the correct cell wavelength (λ) (eq. 6-8). The green regions are zones in which diapirism is not predicted to initiate (eq. 9-13) for a sedimentation rate (V_{sed}) of 0.01 cm/yr. The white fields thus define the “zone of consistency” (the layer thickness relationships which are able to produce diapirism with the correct wavelength). Note that the zone of consistency is smaller and encompasses smaller layer thicknesses for the case of ice diapirism due to the decreased viscosity of ice relative to salt. Gypsum has a relatively higher viscosity and melting temperature, and so both the zone of melting (yellow region in B) and the zone of consistency comprise higher layer thicknesses. For comparison, halite has the highest viscosity and melting temperature, and so is thermally stable across the entire parameter range explored and requires greater layer thicknesses for diapirism to occur. Kieserite diapirism is not shown but exhibits similar wavelength and domain relationships to gypsum. Its higher density causes the green zone in which diapirism is not predicted to encompass the entire field, therefore preventing kieserite diapirism from being viable under these conditions. Parameter space where domain 1 (eq. 6) (D1), domain 2 (eq. 7) (D2), and domain 3 (eq. 7) (D3) are applicable are separated by dashed black lines. (D) Domain transitions for ice diapirism from (A) across the entire parameter space of H_{overb} and H_{diap} . (E) Same as (D) but for gypsum diapirism from (B). (F) Same as (D) and (E) but for halite diapirism from (C).

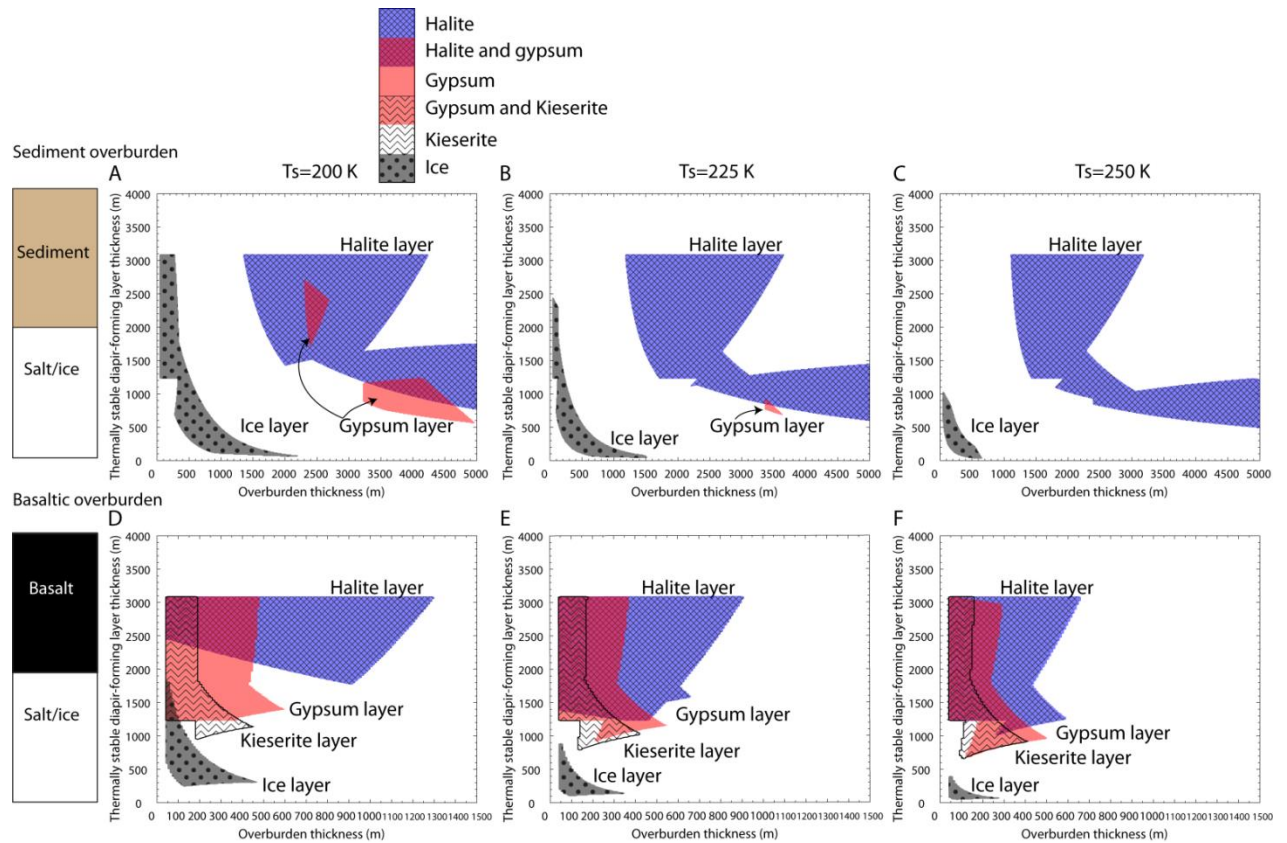


Figure 6. Model results showing the thermally stable diapir-forming layer thickness as a function of overburden thickness for surface temperatures of 200 K (left panels; A, D), 225 K (middle panels; B, E), and 250 K (right panels; C, F). Permissible regions of ice diapirism are shown in shaded black/dotted regions, gypsum diapirism in red, halite diapirism in blue cross-hatched region, and kieserite diapirism in zigzag region. Where the ice, gypsum, or halite diapirism fields overlap, the gypsum (red) field is shown transparent and on top (see legend). The top panels (A-C) show the sediment overburden models, and the bottom panels (D-F) show the basaltic overburden models; the stratigraphic configuration is shown to the left of these panels. The textured/colored regions show the zones in which diapirism is predicted to (1) produce the correct wavelength, (2) allow the diapir-forming layers to be thermally stable in the subsurface, and (3) allow diapirism to initiate using a sedimentation rate (V_{sed}) of 0.01 cm/yr (i.e., the “zone of consistency” from Fig. 5).

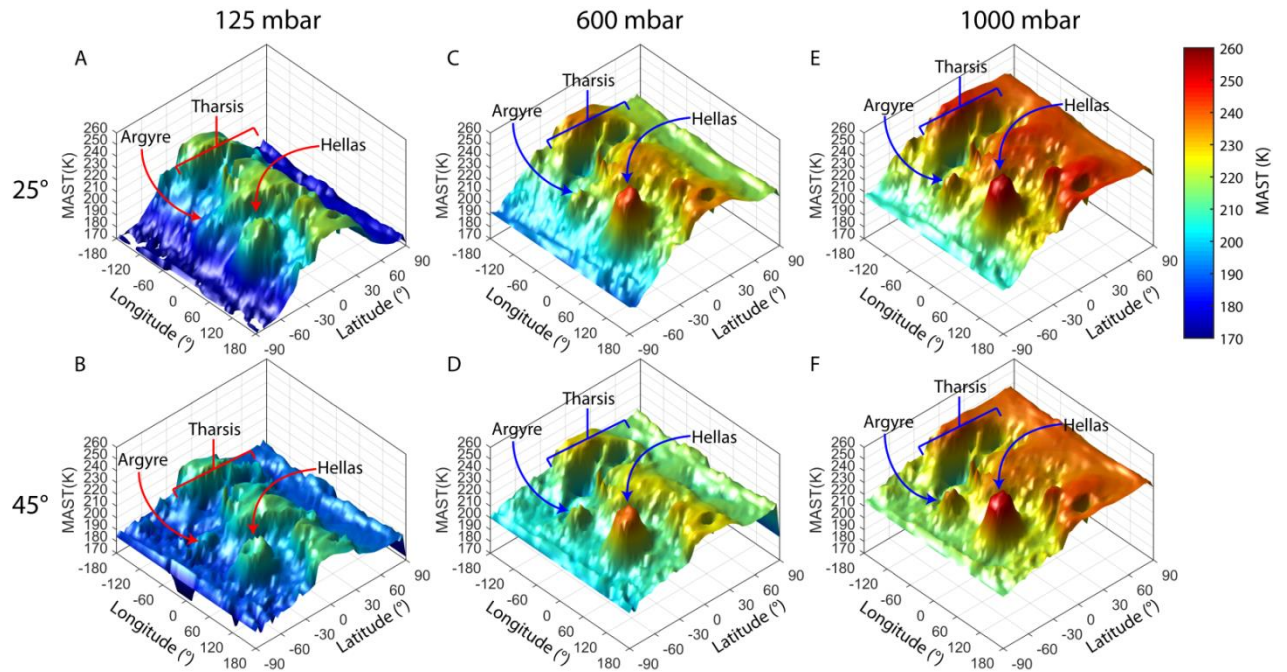


Figure 7. Maps of the mean annual surface temperature (MAST) of Mars derived from the general circulation models of Horan and Head (2016) for a pure CO₂ atmosphere and 100% humidity with a water cycle and eccentricity of 0 (Forget et al., 2013; Wordsworth et al., 2013). A 125 mbar atmosphere is shown in the left panels (A, B), 600 mbar atmosphere shown in the middle panels (C, D), and 1000 mbar atmosphere shown in the right panels (E, F). Model runs with 25° obliquity shown in the top panels, and 45° obliquity in the bottom panels. In each case, Hellas is shown to be among the warmest (if not the warmest) region predicted by early Mars climate models, making it a unique environment for diapirism (Section 4).

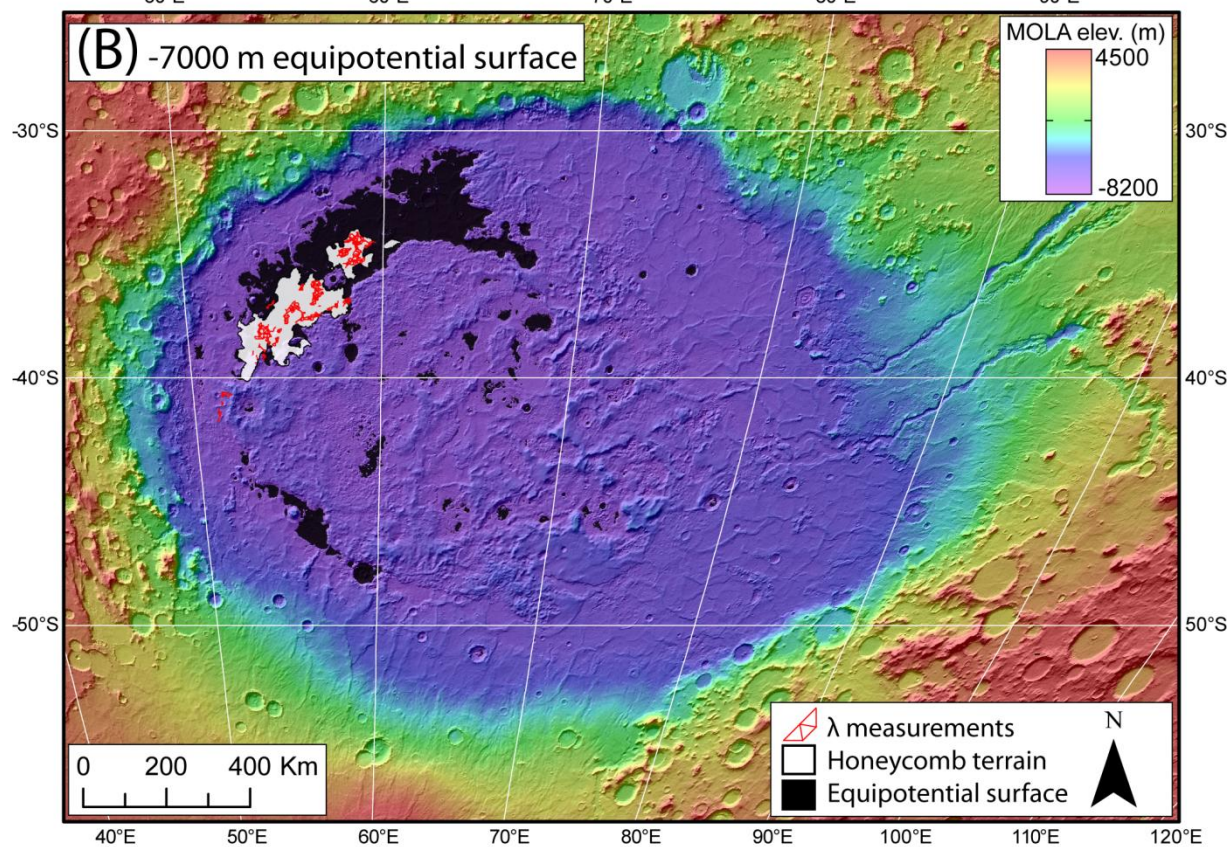
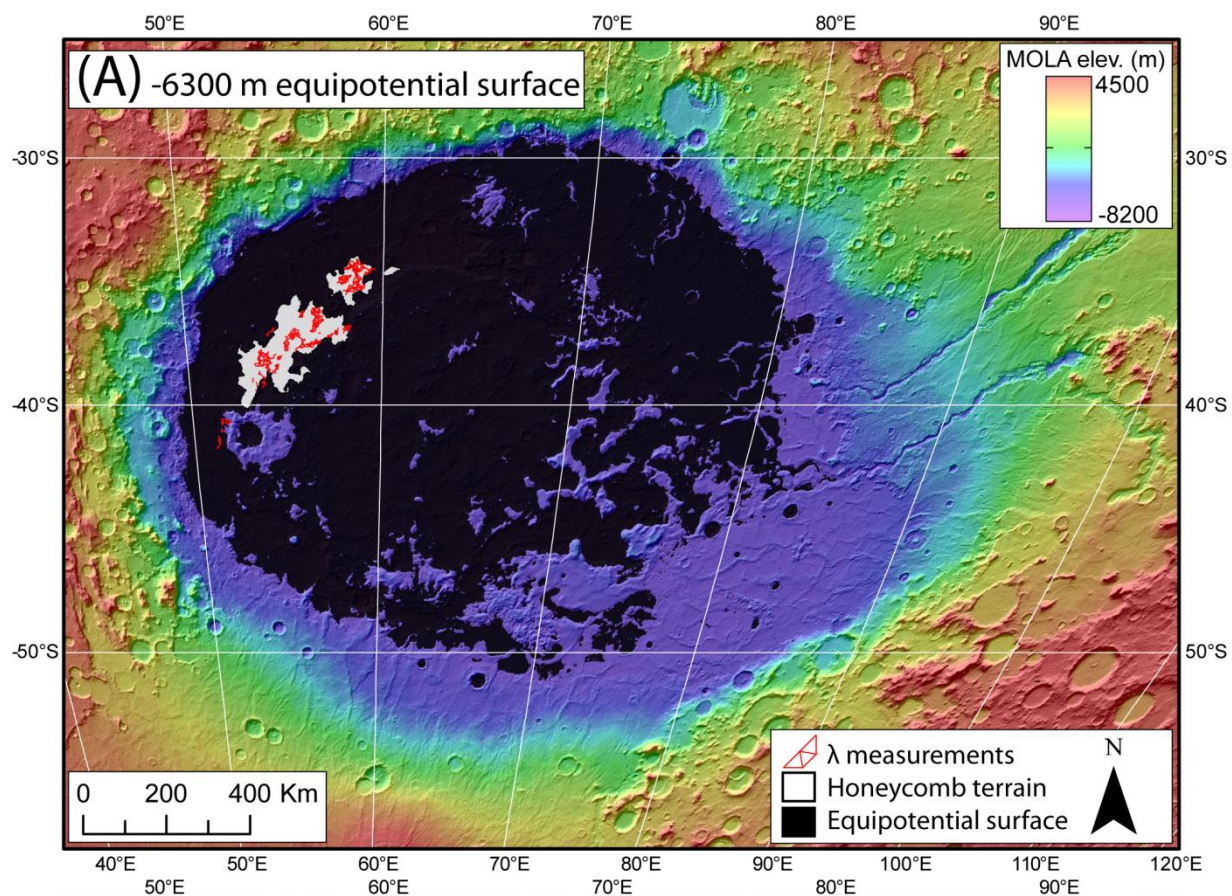


Figure 8. MOLA topography data (Smith et al., 1999) overlain on MOLA shaded relief map (463 m/pixel) showing Hellas basin and the extent of the honeycomb terrain mapped by Bernhardt et al. (2016b) (shaded white region) and our areas of wavelength measurements (red lines). (A) The shaded black region shows the -6300 m equipotential surface, which covers the entire honeycomb terrain. (B) The shaded black region shows the -7000 m equipotential surface, which covers 85% of the honeycomb terrain. These equipotential surfaces serve as lower bounds for the area in which a standing body of water could have emplaced evaporite deposits, and thus require 14-2044 m GEL water to produce the honeycomb terrain through evaporation and subsequent salt diapirism (Section 4.1).

Chapter 6:
Climate, hydrology, and impact cratering on Mars: A synthesis

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1. Introduction

This dissertation work has explored the ways in which climate effects the impact cratering process (Chapters 1, 2, and 3), and how the relationship between climatic variables and impact cratering in turn influences fluvial and landscape degradation processes on Mars (Chapters 2 and 3). This dissertation also explored how impact craters can be used to assess the portion of the martian water inventory that is sequestered in the subsurface as pore-ice, and the behavior and history of the martian cryosphere and groundwater system (Chapter 4). We also examined how large impact basins can serve as microenvironments for unusual geologic processes such as ice diapirism, which can provide a window into the early martian climate (Chapter 5). This work emphasizes that Mars is a planet that has experienced numerous “ice ages” throughout its geologic history, and illustrates how the interaction between these surface/subsurface ice deposits and impact cratering is strongly reflected in the martian geologic and hydrologic record. In this chapter, we summarize the results of each chapter in the dissertation, and then discuss important avenues for future research highlighted by this work.

2. Chapter 1: Testing landslide and atmospheric effects models for the formation of double-layered ejecta craters on Mars

There are several thousand double layered ejecta (DLE) craters present on Mars, generally concentrated throughout the mid-latitudes. This unique crater morphology was originally identified in the late 1970's (Carr et al., 1977); however, their unusual morphology compared with lunar and other martian impact craters made their specific formation mechanism an enigma. Since that time, several hypotheses have been proposed

to explain the ejecta emplacement mechanics and target characteristics unique to DLE craters, but all have had serious issues in explaining the landforms. There has been no consensus within the community on what unique target/impact conditions lead to the formation of these craters.

In Chapter 1 (Weiss and Head, 2017a), we provided morphologic evidence that the inner ejecta facies of DLE craters on Mars is consistent with formation involving a landslide off the structurally-uplifted rim-crest, and inconsistent with previously proposed atmospheric and base surge formation mechanisms. We evaluated evidence which provides strong support for an origin involving impact into and excavation beneath a surface ice-sheet; this includes the documentation of sublimation pits marginal to DLE crater ejecta facies, the association of ring-mold craters with DLE craters, the distribution of expanded secondary craters, and the relationship between the measured ejecta volume and the theoretically predicted ejecta volume.

We then assessed qualitatively how the specific morphology of DLE crater ejecta (grooves, ramparts) are consistent with an ejecta sliding-on-ice origin, and found that the relationship between the friction coefficient and Froude number for terrestrial landslides may offer insight into the conditions required for longitudinal groove formation. We also suggested that the subdued ramparts associated with DLE craters (relative to the other martian layered ejecta craters) may be the result of the underlying low friction ice sheet, which reduces the efficiency of kinetic sieving.

Ultimately, Chapter 1 demonstrated that numerous lines of evidence support the glacial substrate model for DLE crater formation (Weiss and Head, 2013), however, a number of avenues for future work remain, which we discuss below.

2.1. Is the mass of near-rim ejecta accounted for?

In the glacial substrate model for DLE crater formation, the ejecta that is emplaced on the rim is predicted to landslide off the structurally uplifted rim, producing the inner facies. In this scenario, the inner facies is predicted to have an identical volume to the ejecta that was initially emplaced on the rim. One method to further test whether the landslide of near-rim ejecta occurred to form the inner facies is to evaluate whether the volume of the inner facies is indeed consistent with that predicted to have been deposited on the rim during ejecta emplacement. While the observation that DLE craters appear to have an excess volume of ejecta (relative to that predicted) precludes meaningful volume measurements and comparison, one way to perform this test of the hypothesis is to measure the rim-crest heights of DLE craters for comparison with other layered ejecta craters. For example, in the glacial substrate model the ejecta deposited on the rim-crest is predicted to have slid down the slope of the rim. If this model is correct, DLE craters would therefore be predicted to have lower rim-crest heights relative to their diameter compared with other martian craters. Consequently, future work involves measuring the azimuthally averaged rim-crest height of DLE craters and other martian craters. Next, the ratio of the average rim-crest height to the crater diameter can be compared for a sample of these crater populations. If the ratio of rim-crest height to crater diameter is lower for DLE craters than for other martian craters, the hypothesis that the near-rim ejecta slid off in a landslide mode is supported. Alternatively, if the ratio of rim-crest height to crater diameter is equal to or greater for DLE craters than for other martian craters, this hypothesis is not supported. The results of this study will improve our understanding of

DLE crater formation by evaluating a critical prediction of the landslide hypothesis.

2.2. Relationship between groove wavelength and ejecta mobility

In Chapter 1, we showed that terrestrial landslides exhibit a trend wherein a landslide may or may not exhibit grooves depending upon the relationship between the Froude number and the coefficient of friction. We found that lower friction coefficients allowed landslides with lower Froude numbers to exhibit grooves, while higher Froude numbers allowed landslides with higher friction coefficients to exhibit grooves. If this theory is correct, the efficiency of groove formation on the martian layered ejecta craters should correlate positively with ejecta mobility because increased ejecta mobility (i.e., an increased ejecta runout distance) would require either an enhanced Froude number (i.e., from higher speed ejecta) or a lower friction coefficient. We propose that measurements of ejecta mobility should be compared with measurements of groove wavelength and ejecta thickness to explore these potential correlations: the general model outlined in Chapter 1 would suggest that groove wavelength (i.e., the spacing between grooves) should decrease as flow thickness decreases (and Froude number increases), and as ejecta mobility increases (either because friction decreases or Froude number increases). The results of this test would have implications not only for martian layered ejecta craters, but will also improve our understanding of the process of longitudinal groove formation (Dufresne and Davies, 2009) on free surface flows throughout the solar system (e.g., landslides, debris avalanches, granular flow experiments).

2.3. Testing the effects of target substrate on kinetic sieving

In Chapter 1, we showed initial results which support the suggestion that the distal ejecta ramparts of DLE craters are more subdued and exhibit a lower concentration of larger particles compared with other martian layered ejecta craters (Baratoux et al., 2005; Mouginis-Mark and Baloga, 2006; Barlow, 2006; Osinski et al., 2011). We suggested that this observation could be explained by the low friction of the underlying ice sheet, which we predicted to reduce the efficiency of kinetic sieving. Further work is required to test the robustness of this result in a larger sample size using thermal inertia data. Although recent work has attempted such a test (Jones et al., 2016), the sample of DLE craters used in that study used an outdated classification scheme. Our careful review of the craters used in that study showed that none of the craters should in fact be classified as DLE craters.

Because the glacial substrate model predicts that DLE craters should exhibit a lower concentration of large particles at the margins of the ejecta (relative to the other martian layered ejecta craters), finding lower values of thermal inertia (corresponding to smaller grain sizes) along the ejecta margins of DLE craters (compared with other layered ejecta craters) would support the hypothesis that the low friction icy substrate reduced the efficiency of kinetic sieving. Alternatively, this hypothesis is not supported if the thermal inertia of the ejecta margins of DLE craters is equal to or greater than that observed for the other martian layered ejecta craters. This study will improve our understanding of the free surface granular flow and ejecta emplacement/flow/sliding processes.

2.4. DLE craters as a probe into the obliquity history of Mars

Double-layered ejecta (DLE) craters are a particularly enigmatic type of impact crater

found on Mars and are located primarily in the mid-latitudes. In Chapter 1, we showed that these craters are likely to have formed through impact into a thick surface ice sheet which superposes the underlying regolith/rock. In this scenario, the lubricating icy substrate allows the ejecta to slide farther and the near-rim ejecta landslides off of the rim, thereby producing the distinctive inner layer. This continues to be a controversial research area, but there is mounting evidence from outside groups (e.g., Wulf and Kenkmann, 2015; Viola et al., 2016; Boyce et al., 2016) which suggests that the scenario described above appears to be the most consistent with the characteristics of DLE craters. In the context of this hypothesis, can DLE craters be used as a proxy for the timing of ice ages (and thus the climate history) of Mars? Currently, our knowledge of the obliquity (and thus specific ice age) history of Mars is limited to the past ~20 Ma, beyond which the simulation results by other investigators become chaotic (Laskar et al., 2004). Understanding the timing of obliquity variations is important in order to understand both the history of ice-related geomorphological features (e.g., Head and Marchant, 2009), as well as candidate snow-melting events which may contribute to the mid-latitude fluvial geomorphologic record (e.g., Dickson and Head, 2009; Dickson et al., 2009). Could DLE craters offer a method to probe the obliquity history of Mars into the deep past (e.g., Kadish and Head, 2014)?

DLE craters represent one of the numerous surface-ice related features found in the mid-latitudes of Mars (e.g., Kadish et al., 2010; Baker et al., 2010; Fastook and Head, 2014), but their relatively large size compared with other debris-covered glacial features allows statistically significant crater counts to be performed on their ejecta facies to estimate their age. Furthermore, their association with the entire martian crater population

enables the application of global cratering statistical techniques to better understand their ages (e.g., Hartmann, 2005; Kadish and Head, 2014). DLE craters are thus important proxies for the timing and presence of obliquity-driven climate change and ice ages on Mars. In the future, this concept may be extended to measure the ages of a large sample of the global DLE crater population for use as bounding values of the timing of ice ages on Mars. Further work can then develop Monte Carlo simulations that seek to reproduce the timing and global spatial patterns of ice ages on Mars. By assigning spatially random impact events that follow the known martian crater production function (and recording each model impact as forming in an ice-covered substrate or a rock-substrate), the Monte Carlo simulations can produce a variety of simulated crater size-frequency distributions for impact craters that form in surface ice (e.g., double-layered ejecta craters) and those that are predicted to form in a target not covered in surface ice (e.g., the other layered ejecta crater morphologies). By comparing the simulation results with the observed crater size-frequency distribution from global martian crater databases (e.g., Barlow, 2015), statistical predictions can be produced for the history of obliquity variations and surface ice deposition events on Mars over the past ~3.5 billion years (the oldest ages of the layered ejecta crater population). DLE craters can thus be used as a proxy for the timing of surface ice sheets on Mars and may allow us to better understand the history of obliquity shifts and ice ages through time.

3. Chapter 2: Crater degradation in the Noachian highlands of Mars

The presence of Noachian-aged valley networks (Hynek et al., 2010) and highly degraded craters in the southern highlands of Mars (Mangold et al., 2012) points to a

more fluviially active early Mars, and has led to the suggestion that the early Mars climate was much different than observed today (e.g., Craddock and Howard, 2002).

Canonically, it has been believed that the early Mars climate was generally warm and arid, with minimal amounts of rainfall leading to diffusive rain splash dominated landscape erosion (and crater degradation) and complete infiltration of water in the subsurface (Craddock and Howard, 2002), followed by a climatic optimum in the Late Noachian, which produced substantial amounts of rainfall precipitation and surface runoff to form the valley networks (e.g., Irwin et al., 2005).

Recent work, however, has shown that 3D general circulation models with updated physics parameterizations are unable to produce martian mean annual surface temperatures above the freezing point of water during the Late Noachian (Forget et al., 2013; Wordsworth et al., 2013, 2015). These models showed, however, that an increased atmospheric pressure leads to thermal coupling between the atmosphere and the surface, resulting in the higher elevation southern highlands acting as a cold trap for regional surface snow and ice deposits. In this model, hundreds of meters of surface ice could accumulate in the southern highlands. This surface environment (Head and Marchant, 2014) thus offers considerably different predictions for landscape and crater degradation processes than in the warmer environment considered previously (e.g., Craddock and Howard, 2002).

In Chapter 2 (Weiss and Head, 2015), we evaluated whether the degraded state of the Noachian Highlands craters, long believed to be the result of modification by rainfall in a warm climate, is consistent with impact crater formation and modification in a cold and icy ancient martian climate. In this scenario, immediately following an impact event a

substantial portion of the rim-crest may be composed of surface ice, which would produce a subdued rim following ice removal. Continued degradation is then predicted to occur due to backwasting of the rim-crest, and fluvial erosion from top-down melting from peak temperatures on seasonal or longer timescales (e.g., Palumbo et al., 2017), basal melting of the surface ice sheet caused by the elevated geotherm from the overlying low thermal conductivity ejecta, and contact melting between the hot ejecta overlying surface ice. Additionally, we found that the paucity of craters <32 km in diameter and the lack of Noachian-aged craters ≤ 5 km in diameter (Irwin et al., 2013) can be plausibly explained by their formation in thick regional surface ice deposits. Small craters forming entirely in surface ice would be removed following the removal of the ice deposit, and larger craters would appear to be smaller following ice removal due to bowl shrinkage of the paraboloidal crater cavity. We concluded that the characteristics of these craters (e.g., degraded rims, lack of ejecta, fluvial features, paucity of small craters) is at least equally consistent with a cold and icy climate scenario as it is with a warm and wet early Mars climate. Future work is required to further evaluate the characteristics of Noachian landscape degradation to test against the different early Mars climate scenarios, which we review below.

3.1. Evaluating the relationship between terrain age and elevation

Previous work has shown that Noachian terrains appear to have more superposed craters with increasing elevation, which was suggested to indicate that lower elevation Noachian terrains were resurfaced more recently (Craddock and Maxwell, 1993). Later, Irwin et al. (2013) showed that Early Noachian terrains were present at higher elevations

than Middle and Late Noachian terrains, which was also interpreted to suggest that lower elevation Noachian terrains were resurfaced more recently. Craddock and Maxwell (1993) interpreted their results to suggest that rainfall precipitation occurred at progressively lower altitudes as atmospheric pressure declined with time. In Chapter 2, we alternatively interpreted these results to indicate that the lower elevations were resurfaced more extensively through time simply due to the gravity-driven process of downslope fluvial runoff. Importantly, however, the strength of the age-elevation trend appears to be relatively minor, and absent at the highest elevations. How robust is this relationship, and what does it mean for Late Noachian landscape degradation processes?

Interestingly, Bouley et al. (2010) conducted crater counts within the drainage basin of the Parana Valles valley network, and found a similar trend of increasing crater retention age with increasing elevation. This result is remarkable considering that the crater counts were performed at a much smaller scale than those of Craddock and Maxwell (1993) and Irwin et al. (2013). We highlight the importance of reproducing the results of Bouley et al. (2010) over a larger number of drainage basins to further test the robustness of the age-elevation trend. Based on these results, we predict drainage basins associated with valley networks to exhibit a stronger age-elevation trend than other terrains because they are locations where fluvial erosion has occurred. In the Craddock and Maxwell (1993) model, the relationship between terrain and elevation relative to datum would be predicted to remain relatively constant across the surface of Mars since their model predicts that the elevation of clouds and precipitation across the entire southern highlands is decreasing as a function of decreasing atmospheric pressure through time. In the alternative model suggested in Chapter 2, however, the relationship

between terrain age and absolute elevation may be allowed to vary across the surface of Mars since the elevation relative to datum of the fluvial erosion is unimportant. In this model, only the relative elevation between the valley network source region and outlet is important, since downslope fluvial runoff (rather than cloud elevation) would be predicted to produce the age-elevation trend. Evaluating these relationships for a variety of drainage basins associated with valley networks would allow this hypothesis to be tested.

A further avenue for this work would be to compare terrain-age/elevation data from the Moon and Mars. By nature of the volcanic mare flooding of the large lunar basins, the Moon should also be predicted to have some (weak?) trend of increasing terrain age with elevation. By comparing lunar data with martian data, the background age-elevation trend (i.e., the trend in the absence of an atmosphere and fluvial erosion) can be determined. This will allow a better understanding of the role that climate and fluvial processes have played in producing the age-elevation trend on Mars.

3.2. Exploring predicted crater size-frequency distributions in an icy early Mars

Previous work has shown that Noachian-aged terrains appear to be deficient in craters ≤ 32 km in diameter, and entirely absent in craters ≤ 5 km in diameter (e.g., Irwin et al., 2013). In Chapter 2, we tested whether the paucity of small Noachian craters could be related to formation in surface ice deposits. We developed synthetic crater size-frequency distributions (CSFDs) for comparison with the observed the Late Noachian terrain CSFD, and showed that the removal of surface ice deposits could plausibly contribute to the observed paucity of small craters due to terrain deflation (i.e., surface ice removal),

which results in bowl shrinkage of the paraboloidal crater cavity. The concept that the diameter of a degraded crater diameter might be smaller than the original crater diameter is at odds with the canonical view that degradation increases crater diameter through backwasting (Forsberg-Taylor et al., 2004), but it is nonetheless a robust prediction of crater degradation in thick surface ice deposits due to subsequent surface ice removal. In Chapter 2 we evaluated the predicted CSFD of the Late Noachian terrains under the scenario where surface ice deposits between 300 m and 1.5 km were present starting at 4 Ga and then removed at 3.5 Ga (other degradation processes such as fluvial erosion and airfall mantling were not considered). This work highlighted the effect that removal of surface ice deposits can have on CSFDs, but future work is required to fully assess the parameter space. We propose to produce more synthetic CSFDs under a wider age range for the existence of surface ice sheets due to the uncertain duration and thickness (if any) of surface ice in the Noachian. Because of the non-linear cratering flux through time, the timing of ice sheet emplacement and removal will have a large effect on the resulting CSFDs. Although in Chapter 2 we evaluated CSFDs for surface ice between 4.0 and 3.5 Ga, future CSFDs exploring ranges between 4.4 Ga and 3.0 Ga should be considered to evaluate the entire parameter range of viable ages in which surface ice could have been present in the southern highlands. This will allow us to evaluate the viability of different climate/water inventory/surface ice deposition scenarios for Late Noachian Mars.

4. Chapter 3: Impact ejecta-induced melting of surface ice deposits on Mars

In Chapter 2 we showed that crater degradation in a cold and icy early Mars climate will progress in a very different manner than in a warm and wet climate scenario. We

suggested that in a cold early Mars, ice-melting processes may be important in crater degradation, including basal melting and contact melting of surface ice deposits, but the degree to which these processes could operate remained unclear. We addressed these knowledge gaps in Chapter 3 (Weiss and Head, 2016), where we used thermal modeling to assess quantitatively the end-to-end-processes of contact-melting of surface ice sheets by hot ejecta and subsequent basal-melting of these ice deposits by insulating ejecta. In this work, we provided the first estimate of martian ejecta temperature as a function of crater diameter. We then provided predictions for how these melting processes may have operated on Mars through time as a function of crater diameter and specific climatic conditions, and assessed morphologic evidence that these processes have occurred on impact craters throughout the history of Mars. Now that a theoretical framework is in place, we highlight the importance of testing these predictions with detailed geomorphological analyses.

4.1. Testing theoretical framework with geomorphologic analyses

In Chapter 3, we provided a theoretical framework which predicts locations of basal and contact melting, as well as melt fluxes and total volumes, if surface ice is present at the time of crater formation. Armed with these predictions, it is now possible to evaluate individual impact craters to determine if their associated fluvial channels are better explained by rainfall precipitation or impact-induced snowmelt. Such geomorphologic analyses may include (1) assessing the location (e.g., interior, near-rim, or distal) and direction of channels (towards the crater interior or exterior) as a function of crater diameter for comparison with model predictions; and (2) evaluating degraded impact

craters and any associations between adjacent pre-existing craters and fluvial channels to test whether the pre-existing crater may have hosted surface ice fill which was subject to contact melting following the emplacement of hot ejecta from the younger impact. Additionally, melt fluxes and volumes required for channel formation (based on morphometric considerations) can be compared with results predicted by the numerical models developed in Chapter 3 to test whether the morphometry of the fluvial channels is consistent with a snowmelt origin. This will enable a broader understanding of the temporal and spatial distribution of surface ice on Mars.

4.2. Evaluating the formation of the Lyot crater fluvial channels

Lyot crater is a ~225 km diameter impact basin located in the northern lowlands of Mars (50.5°N, 29.3°E), just north of the dichotomy boundary and Deuteronilus Mensae. The floor of Lyot is at an elevation of ~-7000 m below datum, and is thus the lowest point in the northern hemisphere. Dickson et al. (2009) performed crater counting statistics on Lyot and reported an Amazonian age: 1.6 Ga in the Hartmann (2005) system, and 3.3 Ga in the Neukum system (Ivanov, 2001). Secondaries from Lyot crater have been documented to the southwest, southeast, and east, but are generally absent north and northwest of Lyot (Robbins and Hynek, 2011). Lyot ejecta is generally concentrated to the northeast, while the southwest exhibits a paucity of ejecta, suggesting that Lyot was formed by an oblique impact (Russell and Head, 2002) trending from southwest to northeast.

Lyot has historically been of interest due to its large size, relatively young age, and location in the northern hemisphere. These factors make Lyot a unique location to probe

the existence of Amazonian groundwater at depth on Mars (Russell and Head, 2002; Harrison et al., 2010). Because hydrologic models for the martian cryosphere and subsurface typically assume a globally integrated and voluminous groundwater system (Clifford, 1993), the low elevation of Lyot crater is predicted to provide substantial hydraulic head for any underlying groundwater (Russell and Head, 2002). Russell and Head (2002) pointed out that the Lyot impact event represents a transient “drill-hole” into the martian hydrosphere/cryosphere system, and should have penetrated beneath the cryosphere and released groundwater into the crater interior. Russell and Head (2002) reported no evidence within the crater interior for groundwater inflow following the impact, and raised the possibility that substantial groundwater was not present in the martian crust beneath Lyot at the time of impact. More recently, Harrison et al. (2010) documented the existence of an extensive system of fluvial channels present to the west, north, and east of Lyot.

Can Lyot offer insight into the martian groundwater system and the impact cratering process? How did these fluvial channels form? Future work is required to address these questions (Head et al., 2016). In the most recent model, the fluvial channels are proposed to form by groundwater discharge generated by impact-induced seismic liquefaction of shallow, unconfined groundwater (Harrison et al. 2010). This model can be tested by evaluating the thermal constraints of groundwater stability in the shallow subsurface during the Amazonian. It is also possible that vapor plume effects may be responsible: Lyot is a large impact basin, and its well preserved morphology (in comparison with the degraded Noachian highland craters; Craddock et al., 1997; Mangold et al., 2012; Weiss and Head, 2015) offers a unique opportunity to assess models for impact-induced rainfall

(Segura et al., 2002, 2008; Toon et al., 2010; Palumbo and Head, 2017). This model can be tested by evaluating the distribution of fluvial channels associated with Lyot: this model predicts numerous fluvial channels should originate on top of the ejecta due to rainfall precipitation. We also propose to evaluate models where the fluvial features are formed through dewatering of pore-water contained within the ejecta (Harrison et al., 2010). This model can be tested by evaluating the volume of pore water predicted to be present within the Lyot ejecta for comparison with the water volumes predicted to carve the fluvial channels. Noting that Lyot is present at a location and latitude known to have accumulated regional surface ice deposits in the past during periods of higher martian obliquity (e.g., Head et al., 2003; Laskar et al., 2004; Head et al., 2006; Morgan et al., 2009; Madeleine et al., 2009), we additionally highlight the importance of exploring the possibility that the formation of fluvial channels may be related to contact melting from hot ejecta (Weiss and Head, 2016) superposed on regional surface ice deposits. Chapter 3 provides the quantitative framework with which this hypothesis can be tested. The numerical models developed in Chapter 3 can be directly applied to Lyot to determine locations predicted for contact and basal melting, as well as predicted melt fluxes and volumes. This model can be tested by comparing the model results with the required volumes and melt fluxes derived from geomorphometric analysis, and a detailed evaluation of whether regional ice deposits may have been present during the formation of Lyot crater. These analyses will provide a better understanding of impact-related fluvial processes for large basins in the relatively recent Amazonian period of Mars history.

5. Chapter 4: Evidence for stabilization of the ice-cemented cryosphere in earlier martian history

The latitudinal distribution of pore-ice as a function of depth has been the subject of much consideration in the past 40 years due to its importance in the global water budget and the hydrologic history of Mars (e.g., Fanale, 1976; Clifford, 1993). Theoretical estimates on the thermal stability of water in the martian subsurface have shown that pore-ice may be stable within the shallow martian crust from up to several kilometers to several tens of kilometers deep (Clifford et al., 2010). Beneath this zone (known as “permafrost” in the terrestrial literature but as the “cryosphere” in the martian literature), groundwater is thermally stable if it is present. As the geothermal heat flux decays over time and the temperature in the subsurface decreases, the zone in which pore-ice is thermally stable advances deeper into the martian crust. If groundwater is present beneath the cryosphere and is in contact with the advancing freezing front, it can freeze directly onto the cryosphere as it grows, thickening the ice-cemented portion of the martian cryosphere. In places where groundwater is present beneath the cryosphere but remains deeper than the advancing freezing front, groundwater is predicted to be transported upwards through the vadose zone by thermally induced vapor diffusion, and freeze onto the base cryosphere.

The morphology of impact craters on Mars has previously been interpreted to reflect the properties of the target in which they form (e.g., Barlow and Bradley, 1990; Weiss and Head, 2014). In Chapter 4 (Weiss and Head, 2017b), we used a global database of impact crater morphologies that are interpreted to form in the martian ice-cemented cryosphere, and provided an estimate of the zonally averaged thickness of the ice-

cemented cryosphere on Mars today. This enabled us to provide an estimate of the volume of water within the shallow martian crust, the most significant potential reservoir of the total martian water inventory (Carr and Head, 2015). We then examined how these ice-cemented cryosphere thickness measurements compared with thermal model predictions. By varying model parameters such as atmospheric pressure, heat flux, and obliquity, we provided insight into how the deep subsurface ice reservoir has evolved through time on Mars. Our results showed that the martian cryosphere is likely to be supply-limited, and that the supply of subcryospheric groundwater was likely to have been exhausted (frozen onto the cryosphere) in a more ancient period of Mars history.

5.1. Evaluating retention of the ice-cemented cryosphere during vapor diffusion

In Chapter 4, we proposed that the martian ice-cemented cryosphere is likely to be present at all latitudes but may be thinner than the maximum thickness predicted based solely on the thermal stability of ice in the subsurface. If indeed the ice-cemented portion of the cryosphere comprises only a thin upper section of the whole cryosphere, it suggests that there may be a supply-limit of ice (and groundwater) in the martian subsurface, such that insufficient groundwater was present beneath the ice-cemented cryosphere to fully saturate the pores of the cryosphere. This in turn would imply that groundwater is not extant or abundant in the deep martian subsurface today. Recent vapor diffusion modeling work by Grimm et al. (2016) has shown, however, that the equatorial zone of the ice-cemented cryosphere should entirely desiccate to the atmosphere (e.g., Fanale, 1976; Clifford, 1993) on the order of hundreds of millions of years unless there was continuous resupply of pore-ice from the upward vapor diffusion of underlying

groundwater. Therefore, the suggestion that pore-ice remains present within the cryosphere at the equator might imply the presence of an underlying groundwater source to resupply the pore-ice through upward vapor diffusion within the vadose zone.

These works thus provide two end-member scenarios for the current abundance of groundwater on Mars: (1) An anomalously thin ice-cemented cryosphere caused by a supply limit of subsurface water/ice, suggesting the absence of abundant underlying groundwater in the present-day, and (2) a cryosphere that is fully saturated with pore-ice, with abundant underlying groundwater that continuously replenishes the equatorial ice-cemented cryosphere during desiccation. The distribution of pore-ice in the martian crust is thus critical in determining the hydrologic history of groundwater on Mars and whether groundwater remains extant in the martian crust today. Distinguishing between these two scenarios is critical in understanding the current water budget of Mars and the hydrologic history of the planet.

Future work is required to test these two end-member models to further evaluate the abundance of groundwater on Mars in the past and present. One example is to use vapor diffusion models and explore a wide parameter space of climatic (*e.g.*, obliquity, atmospheric pressure) and target characteristics (*e.g.*, presence of shallow ice-cement, sediment compaction, and pore radius) to assess the parameter space in which equatorial pore-ice requires replenishment to avoid complete desiccation. By assessing the viability of vapor diffusion as an equatorial pore-ice replenishment process, we may better understand which of the two scenarios described above is consistent with realistic vapor diffusion parameters and geomorphological constraints. This will allow us to better understand the potential for past and present-day groundwater on Mars, and offer insight

into the hydrologic history of the planet.

5.2. Origin of martian outflow channels in a supply-limited cryosphere

In Chapter 4, we noted that our observation of a supply-limited cryosphere were inconsistent with any model of outflow channel formation involving the pressurization of a saturated global groundwater system. Because outflow channels are the primary line of evidence for abundant groundwater on Mars (e.g., Carr, 1979), it is imperative to understand how they formed. Critically, each of the published models for the groundwater recharge pressurization required to form the outflow channels, including South polar recharge (Clifford, 1993), Tharsis recharge (Harrison and Grimm, 2004; Russell and Head, 2007; Cassanelli et al., 2015), and pressurization from a growing cryosphere (Carr, 1979) suffer from major difficulties (e.g., Carr, 2002; Harrison and Grimm, 2009; Cassanelli et al., 2015), and it remains unclear whether groundwater discharge is capable of providing the necessary fluid fluxes (Wilson et al., 2004, 2009, Kleinhans, 2005; Harrison and Grimm, 2008). Our surprising result, which raised the possibility of a supply-limited cryosphere, should motivate a second look at the formation of outflow channels. Specifically, models which do not require a global pressurized aquifer should be considered and rigorously tested, including those which involve the breaching of surface reservoirs of water (e.g., Harrison and Grimm, 2009), or volcanic activity (e.g., through melting of surface ice; Cassanelli and Head, 2016; subsurface ice; McKenzie and Nimmo, 1999; or direct erosion by lavas; Leverington, 2011). This will provide a critical analysis of robustness of the evidence for abundant groundwater on Mars, and thus provide further insight into the hydrologic history of the planet.

6. Chapter 5: Salt or ice diapirism origin for the honeycomb terrain in Hellas basin, Mars?

The “honeycomb” terrain is a Noachian-aged cluster of elongated cell-like depressions (typically ~7 km in width and ~170 m deep) present within the northwestern floor of Hellas basin (Bernhardt et al., 2016). The honeycomb terrain has recently been interpreted to be the result of salt or ice diapirism (Bernhardt et al., 2016), wherein deposition of sediments or volcanic material on top of a pre-existing salt or ice layer resulted in a gravitationally unstable density inversion which led to diapirism. We noted in Chapter 5 that distinguishing between a salt or ice diapirism origin is critical in understanding the nature of the ancient martian climate because each origin provides different predictions for the climatic conditions within Hellas. For example, a salt layer could accumulate through the evaporation of a standing body of water derived from groundwater flow or surface runoff, whereas an ice layer could be derived from a frozen body of water or an inflowing glacier. In Chapter 5 (Weiss and Head, 2017c), we specifically tested whether the honeycomb terrain is consistent with a salt or ice diapirism origin. We used thermal modeling to assess the stability limits on the thickness of an ice or salt diapir-forming layer at depth within Hellas, and applied analytical models for diapir formation (Fernandez and Kaus, 2015) in order to evaluate the diapir wavelengths predicted for these ice and salt thicknesses for comparison with observation.

We found that both salt and ice diapir scenarios are viable within the explored parameter range, but salt diapirism generally requires ~1-3 km of salt underlying either a basaltic or sediment overburden, and ice diapirism requires ~100 m to ~1 km of ice,

generally with a sediment overburden. Because salt would require deposition in a standing body of water, the volume of water required for salt diapirism is governed by the minimum volume of water required to fill an equipotential surface within Hellas to at least the elevation of the honeycomb terrain. Using this approach, we found that salt diapirism requires a cumulative volume of ~14 to 2045 m global equivalent layer (GEL) of water, compared with only ~0.3-24 m GEL water for ice diapirism. Based on the lower cumulative volumes of water required and the possibility that an ice diapir-forming layer can be formed through a wider variety of scenarios (frozen body of water or inflowing glacier; Fastook et al., 2017), we favored the origin of the honeycomb terrain through ice diapirism. Future work is required to determine if such an ice layer could have formed through freezing of a standing body of water or by an inflowing glacier.

6.1. Assessing the plausibility of glacial flow into Hellas

In Chapter 5, we suggested that the ice-diapir forming layer which may have formed the honeycomb terrain could have plausibly formed through either a frozen body of water or an inflowing glacier. Distinguishing between these two origins will provide further insight into the nature of the early martian climate. As discussed in Chapter 5, Scanlon et al. (2016) and Fastook and Head (2016) found that under certain atmospheric pressure and obliquity conditions, the Late Noachian-Early Hesperian expanded south polar ice cap, the Dorsa Argentea Formation, could have glacially flowed down the topographic slope into Hellas basin, which could have provided a source of glacial ice to form the diapir-forming layer.

Future work is required to determine the entire range of parameter space under which

highlands glaciers are predicted to flow into Hellas (Fastook et al., 2017). Is glacial flow restricted only to the Dorsa Argentea Formation, or would snow and ice be predicted to accumulate and flow along the entire rim of Hellas? Could top-down melting or basal melting of these ice deposits occur and generate surface runoff to produce a standing body of (frozen?) water in Hellas? What early martian water inventory (Carr and Head, 2015) would be required to allow any of the above processes to operate? Is the smaller Argyre basin predicted to behave similarly, and thus predicted to exhibit evidence for diapirism? We highlight the importance of future work coupling the results of 3D general circulation models to glacial flow modeling (Fastook et al., 2017) in order to address these questions. The results of such models will provide further insight into the origin of the honeycomb terrain, the Hellas floor environment, and the nature of the early martian climate.

7. Conclusions

In this dissertation, we highlighted the relationship between the climate, hydrology, and the impact cratering process on Mars. In Chapter 1, we showed that a wide variety of evidence supports the suggestion that the enigmatic martian double-layered ejecta craters form through impact into (and excavation through) decameters-thick regional surface ice deposits. Future work involves (1) evaluating whether the ejecta volumes are consistent with the landsliding hypothesis; (2) testing the hypothesized relationship between the groove wavelength and ejecta mobility for martian craters; (3) analysis of ejecta facies thermal inertia to evaluate the effects of substrate friction on the kinetic sieving process; and (4) using the spatial and temporal distribution of DLE craters as a probe into the

timing of ancient martian obliquity variations.

In Chapter 2, we showed that the degradation state of the Late Noachian highland craters, long believed to result from degradation in a warm and arid climate, could also plausibly be accounted for in a cold and icy climate. Future work involves (1) evaluating the relationship between terrain age and elevation within individual drainage basins of valley networks; and (2) exploring the effects of the presence of Late Noachian hectometers-thick surface ice deposits on synthetic crater size-frequency distributions over a wider parameter range of ice sheet durations and thicknesses.

In Chapter 3, we provided an end-to-end framework for the thermal effects of ejecta emplacement on surface ice deposits. Future work involves (1) detailed morphologic analysis of fluvial channels associated with individual impact craters to evaluate evidence (or lack thereof) of surface ice; and (2) application of model results to the young, large Lyot crater to evaluate the origin of the extensive network of fluvial channels marginal to the Lyot ejecta facies.

In Chapter 4, we suggested that the martian ice-cemented cryosphere is supply-limited on the basis of its anomalously thin global distribution inferred from the excavation depths of martian impact craters. We suggested that the martian groundwater system was assimilated into the growing cryosphere and depleted in a more ancient period of martian history, perhaps as late as ~ 3 Ga. Future work is required to (1) determine subsurface ice loss rates to assess the plausibility of a thin, extant ice-cemented cryosphere in the absence of continued replenishment by vapor diffusion from underlying groundwater; and (2) reevaluate the formation of outflow channels to determine if the predicted absence of abundant groundwater beyond the Hesperian period can be

reconciled with viable outflow channel formation mechanisms.

In Chapter 5, we suggested that the honeycomb terrain in Hellas basin is likely to have formed through ice diapirism on the basis of analytic diapir formation models and hydrologic/topographic/water inventory considerations. Future work coupling the results of 3D general circulation models and glacial flow models is required to assess whether a potential ice diapir-forming layer on the floor of Hellas basin could be sourced by a standing body of water or inflowing glacial ice.

Ultimately, the results of this dissertation emphasize the relationship between climate and impact cratering (e.g., DLE crater formation during periods of high obliquity, the fluvial effects of emplacement of heated impact ejecta, using impact craters to infer the abundance of groundwater through time, and large impact basins as optimal microenvironments for diapiric processes), and highlight future avenues of work that may provide further insight into the nature of the impact cratering process and the martian climate.

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