CLIMATE CHANGE ON MARS AND THE FORMATION OF GULLIES, LOBATE DEBRIS APRONS, AND SOFTENED CRATERS

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DOCTOR OF PHILOSOPHY

in

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by

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Abstract

Climate Change on Mars and the formation of Gullies, Lobate Debris Aprons, and Softened Craters

by

Reid Allen Parsons

Recent data acquired from spacecraft missions has bettered our understanding of the nature and distribution of ice- and water-related features formed during recent periods of climate change on Mars. Numerical modeling of physical processes constrained by these newly acquired observations is an important tool with which hypotheses relating to the Martian climate can be tested. This work describes the development and implementation of a set of these models focused on the formation of a few young, ice- and water-related features. The subjects of this research are gullies, lobate debris aprons (LDAs), and craters with subdued topography known as “softened” craters. Flow of liquid water and ice over and/or within the Martian surface has been invoked in the formation of these features. Quantifying processes such as fluvial erosion and ice deformation using laboratory experiments is a Rosetta stone with which we can read into the climate history of Mars that is written on its surface.

We test the hypothesis that sediment transport on gully slopes occurs via fluvial transport processes by developing a numerical sediment transport model based on steep flume experiments performed by Smart [1984]. At 20° slopes, channels 1 m deep by 8 m wide and 0.1 m deep by 3 m wide transport a sediment volume equal to the alcove volume of $6 \times 10^5$ m$^3$ in 10 hours and 40 days, respectively, under constant flow conditions. Snowpack melting cannot produce the water discharge rates necessary for fluvial sediment transport, unless long-term
(kyr) storage of the resulting meltwater occurs. If these volumes of water are discharged as groundwater, the required aquifer thicknesses and aquifer drawdown lengths would be unrealistically large for a single discharge event. More plausibly, the water volume required by the fluvial transport model could be discharged in \( \sim 10 \) episodes for an aquifer 30 m thick, with a recurrence interval similar to that of Martian obliquity cycles (\( \sim 0.1 \) My).

Radar observations in the Deuteronilus Mensae region by Mars Reconnaissance Orbiter have constrained the thickness and dust concentration found within mid-latitude ice deposits, providing an opportunity to more accurately estimate the rheology of the ice within lobate debris aprons based on their apparent age of \( \sim 100 \) My. We developed a numerical model simulating ice flow under Martian conditions using results from ice deformation experiments, theory of ice grain growth based on terrestrial ice cores, and observational constraints from radar profiles and laser altimetry. We find that an ice temperature of 205 K, an ice grain size of 5 mm, and a flat subsurface slope give reasonable ages for many LDAs in the northern mid-latitudes of Mars. Assuming that the ice grain size is limited by the grain boundary pinning effect of incorporated dust, these results limit the dust volume concentration to less than 4%. However, assuming all LDAs were emplaced by a single event, we find that there is no single combination of grain size, temperature, and subsurface slope which can give realistic ages for all LDAs, suggesting that some or all of these variables are spatially heterogeneous. Based on our model we conclude that the majority of northern mid-latitude LDAs are composed of clean (\( \leq 4 \)vol\%), coarse (\( \geq 1 \) mm) grained ice, but regional differences in either the amount of dust mixed in with the ice, or in the presence of a basal slope below the LDA ice must be invoked. Alternatively, the ice temperature and/or timing of ice deposition may vary significantly between different mid-latitude regions.
The presence of an extensive ice-rich layer in the near subsurface of the Martian regolith could result in viscous creep responsible for softening of craters at middle and high latitudes. The temperature of ground ice will vary both temporally and spatially due, respectively, to changes in Mars’ obliquity and due to the slope effect on the effective angle of insolation. Numerical simulations of viscous creep indicate that these temperature variations cause the pole-facing slopes of craters to be systematically steeper than those of equator-facing slopes. Crater slopes should be most asymmetric between $25^\circ$ and $40^\circ$ latitude, depending on the thickness of the creeping layer. On the basis of the lack of any systematic slope asymmetry observed in the craters, we can place a conservative upper limit of $\sim 150$ m on the thickness of the ice-rich creeping layer assuming a volumetric dust content of $\leq 70\%$ and an exponentially decreasing regolith porosity with depth. If the creeping layer contains relatively clean ice, then the thickness of ice-rich material is limited to $\sim 100$ m or less. The observations also suggest that the thickness of this creeping layer is reduced by $\sim 30\%$ toward the equator. These results imply a global ice-rich regolith water volume of $<\sim 10^7$ km$^3$, comparable to that proposed for a modest-sized northern plains ocean.
To my beloved wife, Ayumi and son, Ko, and to my friends and family.

To all the children who are curious about this world and what’s beyond ... your questions continue to inspire the greatest of scientific endeavors.
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Chapter 1

Introduction

1.1 Looking in from above

The most major revelation in recent history within the field of Earth Science is arguably the discovery of plate tectonics via sea-floor spreading and subduction in the early 1960’s. Its discovery required technically advanced data-collecting techniques involving magnetometers towed behind ships, ocean drilling vessels, and precise earthquake detection and locating through seismology. The global process of plate tectonics is fundamental to understanding how the Earth has been shaped by volcanism, earthquakes, and the splitting and coalescing of continents over time. At the time of its discovery, plate tectonics explained many outstanding mysteries in geology and paleontology, but collecting the supporting evidence required data collection, and observation at specific locations scattered around the globe. Understanding that we humans are standing on broken plates covering the Earth’s surface that migrate over time like giant conveyor belts necessitated looking at the Earth from a different perspective. Radial sections of the Earth’s interior viewed through seismic waves offered just such a perspective, as did data from ocean drilling and paleomagnetic survey campaigns.
If, however, we were to explore the Earth from orbit relying on instrumentation aboard spacecraft missions, we would probably only realize that the process of plate tectonics was operating if it was possible to: a) measure ocean bathymetry from orbit using radar and discover the inter-connected network of mid-ocean ridges, or, b) if we already had some prior knowledge of plate tectonics and knew of the symptoms to look for beforehand. Otherwise, we would be hard-pressed to discover that this fundamental, global process was operating on Earth without the necessary data, or the *a priori* knowledge of such a process (see Oreskes [2001] for more). This perspective of “looking in from above” on our own planet provides some insight into both the processes by which we explore and our level of understanding of other planetary objects in our solar system.

### 1.1.1 The cautious and the qualitative

During my short experience as a Mars scientist I have noticed a dichotomy among the researchers in the community that may well apply to other science communities. There are the skeptical scientists who require several independent observations and/or theoretical calculations in order to be convinced that a hypothesis is valid. On the other hand, there are researchers who are willing to submit hypotheses based on one or two observations without additional supporting evidence. Although these different approaches to conducting science promotes a healthy exchange of ideas within the community, the scientific method requires discriminating between the cautious and the qualitative when looking for answers to the many questions regarding Mars’ history. With the outpouring of data from the recent missions to Mars (for example, Mars Reconnaissance Orbiter with over 40 Terabytes of data expected by the end of its primary science phase) there are more images and data than scientist can look at. From this newly acquired
mountain of data our job as scientists is to find the evidence for processes that continue to shape, or have shaped the surface and interior of Mars.

In the past, astronomers could only rely on telescopic observations of the Moon and planets, and the observer had to rely on their experience (or imagination) to explain what they saw. For example, well into the twentieth century, the craters on the Moon were believed to be formed from volcanism and not from ballistic impacts from meteorites. Only after Meteor Crater in Arizona was identified as an impact structure did the impact theory gain momentum [Koberl, 2001]. It does seem only natural to attribute the formation of a feature on another planetary object to a process familiar to the observer. However, as planetary scientists, we must constantly remind ourselves that the place we are observing is not, in the least, familiar to our sensory experience. If we learn from the past, we must be cautious in drawing conclusions based solely on images of the surface because they do not represent a sufficient amount of evidence to support or refute a hypothesis. In truth, there are many processes that can result in similar features on planetary surfaces, just as craters can be formed both from impacts and volcanism. The defining principle of any science is the ability to test the predictions made by a hypothesis through further observations. In this regard planetary scientists can learn from the example provided by progress in the Earth Sciences over the years.

Our understanding of active processes on Earth is measured by our ability to quantify and model them after studying these processes in the field and in the laboratory. Although conducting lab experiments that accurately represent the conditions on other planetary bodies can be, at best, expensive, and most likely, impossible (not to mention going into the “field”), we must be cautious of basing conclusions on images alone. Further evidence must be gathered through experiments and numerical simulations to provide scientifically sound hypotheses that
explain how features on other planets form.

1.2 The Earth’s kin

Looking at the terrestrial planets in our Solar System - Mercury, Venus, Earth and Mars - Venus is often given the title of “Earth’s Twin” because the two planets have a similar size, composition, and distance from the Sun. However, as we have learned more about these planets through spacecraft exploration, it has become clear that the conditions on the surface of Venus and Earth are vastly different. Venus’ thick, CO\textsubscript{2} atmosphere (90 times as massive as Earth’s) absorbs outgoing thermal radiation in a devastating greenhouse effect that raises temperatures at the surface to 740 K (470°C). The photo-dissociation of water vapor in the upper atmosphere, followed by the excape of hydrogen to space and the reaction of oxygen with gases in the atmosphere or crustal minerals has left Venus bone-dry [Kasting and Pollack, 1983].

Mars, on the other hand, has a diameter only slightly more than half of Earth’s, and lies 50\% further from the Sun. The atmospheric pressure on the surface of Mars is less than one hundredth that of Earth’s and is composed predominantly of CO\textsubscript{2}. This reduced greenhouse effect and distance from the Sun leaves Mars, on average, about 100°C cooler than Earth. However, when compared to Venus, these temperatures are more similar to Earth-like temperatures. In fact, places like the Antarctic dry valleys are used as an analog study site for Mars due to the cold, dry conditions. However, in order to replicate the atmospheric pressure conditions on the surface of Mars, one would have to travel up to the stratosphere to 40 km altitude.

Despite its tenuous atmosphere, Mars has been able to keep much of its water. Therefore, in terms of surface and climate conditions, Mars is the terrestrial planet most similar to
Earth. In addition, the Martian surface holds evidence of processes that we see on Earth’s surface such as past catastrophic flood events, channel networks, and ice ages. Studying these features provides a means of probing its climate history - giving insight into the past environments that have existed on the Martian surface. Understanding how processes such as fluvial erosion and glacier flow operate on Earth can help us begin to read the climate history of Mars that is written on its surface. My dissertation attempts to translate a part of this history.

1.3 A brief on Martian geologic history

This dissertation addresses three features of different ages relating to climate change on Mars, so I will briefly introduce some nomenclature regarding the geography and geologic history of Mars that will be used in the subsequent chapters. First, I will discuss the Martian geologic timescale followed by a brief description of how the Martian climate changed in its early history. Next, I will provide some background on how recent climate change on Mars has taken place, and how the features I focus on in this work relate to these changes in Martian climate.

The Martian geologic timescale is broken into three epochs: the Noachian, Hesperian, and Amazonian (from oldest to youngest) based on the relative age of surfaces with respect to three prominent terrains. These three extensive terrains are Noachis Terra, Hesperia Planum, and Amazonis Planitia (Figure 1.1). However, the boundary between these epochs is actually defined by a characteristic crater density value set by Tanaka [1986]. A Noachian age surface has 25 or more 16 km diameter craters per $10^6 \text{ km}^2$ and 200 or more 5 km diameter in the same area. Hesperian surfaces have fewer craters than Noachian surfaces, but have a 67 or more 5 km diameter craters or 400 or more 2 km diameter craters. Finally, Amazonian surfaces have fewer...
craters than Hesperian surfaces. The absolute ages of the boundaries between these epochs is between 3.5 and 3.8 Gy and between 1.8 and 3.5 Gyr for the Noachian-Hesperian and the Hesperian-Amazonian boundaries, respectively [Ivanov, 2001; Hartmann and Neukum, 2001].

1.3.1 A note on dating the Martian surface

Because a sample return mission to Mars has yet to be performed, we have no way of knowing exactly how old a particular part of its surface is. Although the shergottites, nakhlites, and chassignites (SNC) meteorites are datable pieces of Mars that have landed on Earth, we have no way of knowing where they came from and they are therefore of little use in dating the surface. Therefore, we must rely on the statistics of impact cratering events to tell us the age of different surfaces. Determining the relative ages of different surfaces is, in theory, relatively straightforward using intersection (superposition) relationships and crater abundance. However, the degradation of craters, and ambiguous intersections of different surfaces introduces uncertainty into this practice.

To convert the density of impact craters within a given area to an age estimate requires even more effort. Because we have samples from the Moon that tell us how old a surface with a particular crater density is, a model can be calibrated that estimates the relative impactor flux between the Moon and Mars in order to determine the age of Martian surfaces. Initial models put forth by Neukum and Wise [1976] and Hartmann et al. [1981] use the crater densities from the Moon, and the age determined by the Apollo lunar samples and apply a correction due to differences in Martian gravity, impactor velocity, and target properties to give an age estimate for the Martian surface. Subsequent, updated versions of these models have been periodically released with the most recent by Hartmann [2005].
Using this technique absolute age estimates can be made for surfaces on Mars. However, dating young surfaces is particularly difficult because the flux of impactors in the inner solar system declined significantly after the (late) heavy bombardment, \( \sim 3.8 \) Ga. Therefore, small differences in crater densities represent large spans of time for surfaces younger than Hesperian in age. For example, the difference in crater densities between 3.7 and 3.6 Ga surfaces is greater than that of 3 and 2 Ga surfaces (e.g. Hartmann [2005]). Furthermore, secondary impacts generated from the ejecta of a primary comet or asteroid impact introduces more uncertainty into the practice of converting a crater density into an absolute age because it alters the crater population in different ways on different planetary objects. In addition, many features of interest only cover a small area and do not permit a statistically significant sample of craters to be counted. Thus, the ages suggested for many of the youngest Martian features are very approximate.

1.3.2 The observable record of early Mars

A global shaded relief map of Mars is shown in Figure 1.1. The highlands that cover nearly all of the southern hemisphere are the most heavily cratered, and oldest part of Mars. What actually formed the hemispheric dichotomy of the southern, cratered highlands, and the northern, smooth lowlands remains a mystery although a giant impact [Wilhelms and Squyres, 1984; Andrews-Hanna, J.C. et al., 2008; Marinova et al., 2008; Nimmo et al., 2008] or mantle convection [Sleep, 1994; Roberts and Zhong, 2006] are presently the leading hypotheses. Among the notable features within the southern highlands are valley networks, outflow channels, large impact basins, phyllosilicate detections, and a strong remnant crustal magnetization. These Noachian features speak to a very active early Mars where large impactors bombarded
the surface, a hot interior led to volcanic outpourings, convection in the core drove a magnetic dynamo, and liquid water episodically carved through the landscape. As the internal heat dwindled, the core dynamo shut down and volcanism waned after much of the Tharsis bulge had been constructed at the close of the Noachian $\sim 3.8$ Ga [Phillips et al., 2001].

All of the large (> 2000 km) impact basins were formed prior to the Noachian, with Hellas representing the last of these large impacts. The Hellas impact at roughly 4 Ga has been adopted as the event signaling the start of the Noachian and the start of the geologic record that is currently visible on the surface [Carr and Head, 2010]. Although the Martian crust retains features older than the Hellas impact such as magnetic anomalies in the southern highlands and partially-filled large impact basins that pre-date Hellas [Nimmo and Tanaka, 2005], smaller scale surface features relating to the Martian climate are only preserved following the formation of Hellas. Subsequent to these mega-impacts, the Martian surface continued to be bombarded
by large (∼ 10 km diameter) meteorites, on average, every million years [Ivanov, 2001]. The heat generated from these large impactors may have produced the transient warm temperatures necessary to generate the water-related features associated with the Noachian [Segura et al., 2002], although a massive greenhouse gas-laden atmosphere fed by volcanic outgassing may have also been responsible for generating the conditions necessary for these features to form [Stewart Johnson et al., 2008, 2009].

1.3.3 Climate conditions on early Mars

Although researchers agree that early Mars was a warmer and wetter place than it is today, there is not yet a consensus as to how much warmer it was, what mechanism warmed the climate, or how long these warmer conditions persisted. One viable hypothesis states that a thick atmosphere of greenhouse gases was responsible for a warm early Mars, but this hypothesis has to overcome some obstacles.

In the early Solar System, the faint young Sun would make greenhouse warming extremely difficult. Large amounts of strong infrared absorbing gases like SO$_2$ or CH$_4$ must be invoked [Carr and Head, 2010] because H$_2$O and CO$_2$ alone would not have been up to the task [Kasting, 1991]. Forget and Pierrehumbert [1997] suggest CO$_2$ clouds scattering infrared radiation back to the surface of Mars may have delivered the additional heat flux needed to maintain the conditions necessary for liquid water to be stable, but this process requires nearly 100 percent cloud cover [Lammer et al., 2008]. Furthermore, although some carbonate has been detected on Mars [Ehlmann et al., 2008], the lack of large carbonate deposits on Mars [Bibring et al., 2006] also presents a problem since these deposits should have formed on a wet Mars with a massive CO$_2$ atmosphere. The lack of carbonate deposits can be explained by volcanically-
injected SO$_2$ that brought on acidic conditions preventing carbonate formation [Fairén et al., 2004; Halevy et al., 2007].

By the beginning of the Noachian the Martian dynamo had collapsed (e.g. Arkani-Hamed [2004]; Johnson and Phillips [2005]), and CO$_2$ would have slowly leaked from the atmosphere via atomic collisions with O$^+$ generated by the solar wind and the interplanetary magnetic field [Kass and Yung, 1995]. Although Kass and Yung [1995] suggest that it would take 3 Gy to lose $\sim$ 3 bars of carbon dioxide and a 80 m thick global layer of water, the early Martian atmosphere may have reached a runaway point at which the loss of CO$_2$ to space cooled the planet enough to allow the freezing out of greenhouse gases onto the surface which then led to further cooling.

Although we do not yet fully understand how different the early Martian climate was from the current climate, the geomorphology of Noachian terrains tell us that large-scale water related activity peaked around 3.8 Ga. Precipitation of rain or snow, eruption of ground water, erosion of channel networks, deposition of fluvial deltas, and the formation of crater lakes all would have taken place during the Noachian and early Hesperian (e.g. Howard et al. [2005a,b]; Malin and Edgett [2003]; Fassett and Head [2005]; Fassett and Head [2008]). Still, the time-averaged erosion rates for Mars during the Noachian period are more than 10 times smaller than terrestrial rates [Carr and Head, 2010], and some Noachian surfaces appear to be nearly pristine [Carr, 1996] indicating that the early Martian hydrologic cycle was neither as widespread nor as continuous through time as Earth’s is today.
1.3.4 Where did the water go?

Following the early hydrologic activity during the Noachian, subsequent large outflows of water over the Martian surface are generally attributed to melting of ground ice from volcanic activity during the Hesperian and Amazonian [Fassett and Head, 2008]. Overall, the climate became colder and drier and water-related activity less wide-spread as volatiles froze onto the polar caps. However, the fate of the abundant water that was present during the Noachian remains unresolved. Although there was likely a significant amount (4-80 m global layer) lost to space [Lammer et al., 2008; Kass and Yung, 1995], the remaining water must have diffused into the regolith, or frozen onto the polar caps. Given the uncertainty in how much water Mars started with, and how much has been lost to space, the current inventory of retained Martian water remains unknown. Clifford [1993] has suggested Mars may still have an active water cycle in which water melted from the base of the polar caps feeds a global aquifer which delivers water equator-ward to replace the water diffusing through the regolith and into the atmosphere. Atmospheric deposition of frost and accumulation of ice on the poles completes this cycle. This model would invoke a nearly water-saturated Martian regolith extending to ~ 5 km depth. Assuming a surface porosity of 20% this would amount to a 540 m global layer of water. In the hopes of better constraining the evolution of the Martian water inventory, I investigated its role in shaping the landscape over the past ~ 1 Gy by studying gully formation, lobate debris aprons, and crater softening.

1.3.5 Recent climate change

Compared to the outflow channels and valley networks of the Noachian, gullies and lobate debris aprons represent more recent, although perhaps less dramatic, periods in which the
climate was different to what we see today. These late Amazonian (~ 100 Ma for LDAs and < 10 Ma for gullies) features record significant episodes of hydrologic activity at mid-latitudes on Mars. Episodic climate variations resulting from changes in the tilt of Mars’ rotation axis can explain the formation of these younger features due to the redistribution of water ice between the poles and lower latitudes over $10^5 - 10^6$ yr timescales [Laskar et al., 2002; Fanale et al., 1986; Mellon and Jakosky, 1995]. The history of the tilt of the rotation axis, or obliquity, of Mars is constrained for the past 10 My (Figure 1.2) based on gravitational interactions between the planets (mostly Jupiter in the case of Mars) as they orbit the Sun [Laskar et al., 2002], but is unknown at earlier times due to the chaotic dynamics of orbital interactions [Laskar et al., 2004]. During times of high obliquity (> 30°) ice sublimating from the poles can be deposited at lower latitudes (< 60°). Global climate models suggest that at very high obliquity (> 45°) precipitation of ice would occur on the equatorial Tharsis Montes, and east of Hellas basin [Forget et al., 2006]. Many cycles of these ice ages have likely occurred to form viscous flow features [Milliken et al., 2001], a mid-latitude ice-rich mantling unit [Mustard et al., 2001],
gullies, and LDAs.

1.4 My piece of the puzzle: Geomorphology of young Martian surfaces

This dissertation presents numerical models for three geologic processes related to water on Mars: fluvial erosion of gullies, flow of glacial ice in the form of lobate debris aprons, and crater softening via viscous creep of an ice-laden regolith. These geologic processes are listed in order of ascending time scale with gullies being the youngest features, and softened craters being the oldest. Examples of these features are shown in Figure 1.3. Key findings from these projects include: constraining the amount and potential sources of water involved in gully formation, determining the properties and purity of ice contained in LDA deposits, and constraining the amount of ice contained in the Martian regolith.

1.4.1 Gully formation: How wet did it get?

Gullies incised into the pole-facing slope of a crater in the southern mid-latitudes are shown in Figure 1.3a. The location of this gullied crater, which is the focus of the research presented in Chapter 2, is given in Figure 1.1. Since the discovery of gullies by Malin and Edgett [2000] using images from the Mars Orbiter Camera on board Mars Global Surveyor (MGS), the flow of liquid water has often been invoked to explain their formation. However, the current pressure and temperature conditions on the Martian surface do not permit water to persist as a liquid. Yet the majority of gullies are nearly free of impact craters suggesting they formed under conditions similar to those of the present-day. Subsequent observations unearthed a variety of gully morphologies and raised further questions about how much water, or if water
is required at all, in forming gullies. Focusing on a gully morphology most indicative of fluvial activity (gullies with sinuous channels), we developed a numerical model of sediment transport on the steep, gully-incised slopes in order to address the following outstanding questions: If flowing water is responsible for transporting sediment, then how much water is required to carve out a gully alcove and deposit a distal fan? What are the sediment transport rates, and how much time is needed to form gullies? What potential sources of water could produce the water discharge rates necessary to form gullies via surface runoff?

The answers to these questions help us determine the duration and quantity of recent water-related activity on Mars and help constrain which of the potential reservoirs of water and
ice are responsible for supplying water to gully alcoves in the recent past.

### 1.4.2 Lobate Debris Aprons and Martian climate change

Figure 1.3b provides an example of an LDA in the Deuteronilus Mensae province (Figure 1.1). Although LDAs are most abundant in this province (located in the northern mid-latitudes) [Kochel and Peak, 1984; Crown et al., 2006], a small population of LDAs are also found in the Argyre basin, and an abundant population of at least 90 debris apron complexes are found to the east of Hellas impact basin [Squyres and Carr, 1986; Squyres et al., 1992; Pierce and Crown, 2003]. LDAs have been identified as ice-related features since the Viking missions in the 1970’s, and although subsequent topography data collected by MGS in the late 1990’s provided further evidence for the presence of ice below a protective cover of regolith, the concentration of ice within these deposits remained a mystery. Most recently, the Mars Reconnaissance Orbiter (MRO) shallow radar (ShaRAD) instrument detected a large amplitude reflected signal from the base of several LDA deposits [Holt et al., 2008; Plaut et al., 2009] suggesting ice concentrations in excess of 70% by volume within these deposits. In total, these deposits may contain $\sim 10^5$ km$^3$ of ice, or about 10% of the volume of the current northern polar cap [Crown et al., 2006].

Because LDAs are found at latitudes lower than where ice is stable at the surface currently, they must have formed under different climatic conditions. Given the uncertainties in crater age dates on LDA surfaces, their age is only roughly constrained to the past few 100 Ma [Morgan et al., 2009; Mangold, 2003]. I developed a numerical model of ice flow in order address the following questions which will be addressed in Chapter 3: Given the age constraints on LDAs from crater counts, what rheological properties of the ice are required to give runout
distances comparable to the observed LDA runout lengths over $\sim 100$ My timescales? What do these rheological constraints say about the ice grain size, dust concentration, and temperature of LDAs?

### 1.4.3 Crater softening and the Martian water budget

One of the most likely places for storing water from the Noachian is within the regolith as groundwater. Determining how much water is held within the Martian regolith has important implications for the current hydrologic cycle and the evolution of the Martian climate. A crater with softened topography is shown in Figure 1.3c. The crater rim is smoother and the topography more muted than the fresh crater shown in Figure 1.3d. Gradual flow of an ice-laden regolith has been suggested in smoothing high latitude terrains [Kreslavsky and Head, 2000], and infilling craters on the polar layered deposits [Pathare et al., 2005].

We investigate the effect that viscous relaxation of an ice-laden regolith would have on crater topography using a numerical model in Chapter 4. Using this tool we investigate the question of how much frozen water is contained within the regolith.

### 1.5 Portions of this thesis for publication

The chapters constituting the body of this dissertation have been published or submitted for publication as follows.


Chapter 2

Gullies

2.1 Background

Martian gullies, first identified by Malin and Edgett [2000], are of critical importance for understanding the evolution of the Martian hydrosphere and climate. These features probably required liquid water to form and are apparently young, suggesting that liquid water has been present at or near the surface of Mars in the recent past. However, there is currently no consensus on the quantities of water involved in gully formation, or indeed if water is actually required at all. In this chapter we develop and apply a quantitative model of one proposed mechanism for gully formation - fluvial sediment transport - to estimate the fluid discharge rates required and thus to explore whether this mechanism is feasible.

The morphology of Martian gullies (Figure 2.1) consists of an eroded alcove, incised channel, and a depositional apron. The distribution of gullies on Mars is limited to mid to high latitudes, with the most well-developed occurring between 30 and 45 degrees [Bridges and Lackner, 2006]. Gullies are among the youngest features on Mars based on their superposition on relatively young features such as dunes and polygonal terrain, as well as a general scarcity...
of cratered gullies (e.g. Heldmann et al. [2007]).

The mechanism by which gullies form on Mars remains controversial although numerous models have been put forth. Gullies may have formed by the melting of surficial or near-surface ice deposited during periods of high obliquity [Costard et al., 2002; Christensen, 2003; Williams et al., 2009; Bridges and Lackner, 2006]. Alternatively, groundwater discharge (perhaps controlled by obliquity changes) may be a source for gully water [Mellon and Phillips, 2001; Gilmore and Phillips, 2002]. Others have suggested that gullies form from groundwater-fed springs [Malin and Edgett, 2000; Heldmann et al., 2007], wet debris flows [Costard et al., 2002], dry debris flows [Treiman, 2003], or perhaps by CO2 out-gassing [Musselwhite et al., 2001]. Resolving which mechanisms contribute to gully formation is important in determining whether and how the past Martian climate differed from the current conditions.

On Earth, at least three mechanisms have been observed to form depositional fans: avulsing channelized rivers, sheet flows, and debris flows [Parker et al., 1998; Schumm, 1977; Blair and McPherson, 1994]. Although a depositional fan is usually accumulated by a combination of these processes, fans dominated by debris flows tend to have higher slopes than those dominated by fluvial processes [Harvey, 1984; Williams et al., 2006]. Debris and sheet flows tend to form over-steepened fronts, and, depending on their thickness, result in a convex-up topographic profile [Whipple and Dunne, 1992]. In many cases, deposition on alluvial fans occurs by a combination of debris flows, streamflows and intermediate hyperconcentrated flows [Sohn et al., 1999]. Determining which of these processes dominates often requires detailed fieldwork and facies analysis [Whipple and Dunne, 1992; Sohn et al., 1999, e.g.] not generally possible on Mars.

An Icelandic analog study by Hartmann et al. [2003] found gully morphologies sim-
ilar to Martian gullies with sinuous channels initiated by debris flows, and later scoured by stream flow. Howard et al. [2008] have obtained estimates of the fluvial discharge rates needed to generate sinuous channels observed on distal fans on Mars based on theory from Ikeda et al. [1981]. Their work suggests an average discharge rate of $4.9 \, \text{m}^2 \, \text{s}^{-1}$ per meter channel width with extrema at 0.4 and $13 \, \text{m}^2 \, \text{s}^{-1}$ (Section 2.3.1). However, the theory implemented by Howard et al. [2008] assumes shallow slopes, small viscous forces, and that mean flow is steady and uniform downstream - assumptions that may not be appropriate for Martian gullies. In our study area at $38^\circ \text{S}, 218^\circ \text{E}$ (Figure 2.1), slightly sinuous channel forms are found in all gullies - a morphology consistent with fluvial or wet debris flow processes. Although some gullies show pronounced levees, suggesting that debris flow processes can dominate [Mangold et al., 2003], most gullies, including those in our study, do not.

Gully slopes have been investigated hitherto using MOLA topography [Dickson et al., 2007; Heldmann and Mellon, 2004]. Slope data are important because the rate of sediment transport is a sensitive function of slope (see Section 2.3.1). In this work we obtain higher resolution slope data using stereo pairs of HiRISE images. We combine our slope observations with a simple numerical model of fluvial sediment transport to determine the timescales and water volumes involved in gully formation, assuming that fluvial processes are taking place.

The model results can be compared with proposed sources for water to determine if gully formation via fluvial processes is plausible. Similar numerical work has been done to determine erosion rates due to fluvial processes on Titan [Collins, 2005; Perron et al., 2006], while two recent laboratory studies have examined the hypothesis that Martian gullies are primarily fluvial- [Coleman et al., 2009] and debris flow- [Védie et al., 2008] related processes, respectively. The study by [Coleman et al., 2009] argues that water flows under Earth condi-
Figure 2.1: Section of HiRISE PSP_002514_1420 image located at 38°S, 156°E showing a series of gullies incising into a pole facing crater slope. Lines are locations of some topographic profiles measured using stereo photogrammetry.

Debris flow experiments by the melting of near-surface ground ice in silty materials conducted by [Védie et al., 2008] in a cold room at -10°C closely reproduce a specific type of gully morphology found on Martian sand dunes consisting of a narrow, leveed channel terminating in the absence of a depositional apron [Reiss et al., 2007] suggesting that seasonal variations in a periglacial environment may be responsible for gully formation of this specific type.

In the next section we document our observations of gully morphology in our study area (Figure 2.1), while in Section 2.3 we detail our numerical model and the parameters adopted. Section 2.4 presents the results, and Section 2.5 discusses their implications.
2.2 Slope and topography measurements

Figure 2.1 shows our study area, located at 38°S, 218°E, which consists of a series of about a dozen ~1 km long gullies incised into the pole-facing slope of a crater 20 km in diameter. Our work is focused on this area because all the gullies exhibit a morphology indicative of fluvial activity. By looking at a specific location, our goal is to limit the number of variables influencing gully formation such as differences in climate or regional hydrology. The morphology of these gullies is similar to those found in other mid-latitude areas, but our topography measurements are limited to locations with HiRISE stereo coverage, and thus, we have focused our effort on this one location.

The morphology of the gullies in this location is characterized by a sinuous interior channel (with a wavelength ranging between 30 to 60 m and a sinuosity of < 1.04) incised into the alcove floor and depositional apron. Widths of the interior channel range from 1 to 5 m, and marginal channel levees commonly associated with debris flows [Johnson and Rodine, 1984] are absent in this location. These observations suggest that fluvial processes are at work. The variability in the presence of a channel on the fan, the depth of channel entrenchment on the fan, location of channel on the fan, and fan dimensions (ranging from incipient fan/alcove formation to fans ~ 10^5 m^2 in area) between neighboring gullies suggests a cyclic process of avulsion, entrenchment, and back-filling as is observed on terrestrial alluvial fans [Schumm, 1977; Blair and McPherson, 1994]. The Martian gully fans also lack large (~m-sized) boulders which are present in the exposed alcoves, suggesting preferential downstream transport of smaller grains [Welty et al., 2008], or deposition by dry debris flows [Treiman, 2003].

We use a stereo pair of HiRISE images to measure relative elevation changes between manually-selected points employing a method described by Kreslavsky [2007] (Figure 2.1). In
this method, an observed parallax between pairs of points in a stereo pair of HiRISE images is used to determine a relative change in elevation using the camera viewing orientations. The length of the parallax vector describing the offset between identical points in the two images is used to determine the elevation change. After determining the elevation change, the orientation of the observed parallax vector can be compared with the orientation of an idealized parallax vector using the calculated change in elevation and camera orientations to quantify the error in the measured relief. Typically, we measure slopes between 9 or 10 points along the gully starting above the headwall of the alcove and ending beyond the extent of the depositional apron. Typical errors in slope using this method are between 0.5 to 2° over 100 m length scales. The error in the measured relief increases as the distance between the selected points decreases.

Figure 2.2a shows a close-up of one particular gully with the locations of stereo topographic profiles superimposed. Figure 2.2b shows the longitudinal profile, demonstrating a concave-up profile with slopes declining from $30^\circ$ (58%) at the alcove headwall to $4^\circ$ (7%) at the base of the apron. Figure 2.2c shows transverse profiles across the distal alluvial apron. The low relief of the apron makes the relative errors in topography larger, resulting in larger uncertainties in apron volume.

Based on the topographic relief measurements made using the stereo technique, we estimate a maximum vertical offset of 25 m between the background topography and the floor of the channel. Approximating the alcove as a triangular prism 120 m wide, 400 m long and 25 m deep gives a volume of $6 \times 10^5$ m$^3$. Based on the topography measurements of the distal apron shown in the Figure 2.2c, apron volume is comparable, at $\sim 10^5$ m$^3$ (± an order of magnitude) of eroded material, although this measurement is uncertain due to the large errors in measuring vertical relief over short distances (±5 m in elevation over 50 m length scales).
The stereo measurements show that the background slope on which the gullies are located has a slope of $22^\circ$ (40%). Out of the nine gullies we analyzed, all show a steadily decreasing slope from an average of $30 \pm 4^\circ$ (58 ± 10%) at the alcove headwall to $16 \pm 2^\circ$ (29 ± 4 %) at the head of the apron (Figure 2.2d) [Parsons et al., 2008]. These measurements are in agreement with previous gully slope measurements done at MOLA resolution in a different region [Dickson et al., 2007]. However, these depositional apron slopes on Mars are significantly steeper than terrestrial fluvial alluvial fans, with slopes more closely resembling those of terrestrial clast-rich debris flows [Williams et al., 2006; Blair and McPherson, 1994; Stock and Dietrich, 2006]. On the other hand, the stresses arising from such slopes are lower on Mars because of the lower gravity (Section 2.3.1). To proceed further, we assume that the gullies are formed by fluvial processes, and develop a simple numerical model to investigate the implications of making this assumption.

2.3 Fluvial Sediment Transport

2.3.1 Theory

One of the major unknowns regarding the formation of Martian gullies is the degree to which water is involved in transporting sediment. Proposed gully formation hypotheses span the range from dry flows to wet debris flows to fluvial transport. Head et al. [2008] suggest that the dominant sediment transport process changes with time as a gully evolves. Here, we focus on fluvial sediment transport as one end-member in this spectrum of hypotheses in order to determine whether it is consistent with the expected timescales and water volumes available for gully formation.

There are a number of fluvial sediment transport predictors used to estimate sediment
Figure 2.2: a) Gully within image PSP_002514_1420 shown with a b) longitudinal topographic profile (numbers represent slope in degrees), c) three cross sectional profiles transecting the distal alluvial apron, and d) a histogram of slope measurements made of 9 gullies in the HiRISE stereo pair shown in Figure 2.1.
discharges for a given channel. These predictors utilize empirical data of sediment discharge and relate it to quantifiable physical properties of stream channels such as slope, sediment grain size, and channel depth. However very few predictors have been calibrated to slopes as steep as those of Martian gullies. An additional complication comes from not knowing the mechanism of erosion. For instance, when there is no shortage of sediment, alluvial processes of redistributing sediment dominate, whereas sediment-starved streams will erode via plucking or by particles impacting bedrock [Lamb et al., 2008; Sklar and Dietrich, 2004].

A further complication is determining the mode by which sediment is transported. Within the fluvial transport regime, sediment can be transported as dissolved load, suspended load, or as bed load. Here we implement an empirical sediment transport capacity prediction function developed by Smart [1984]. This predictor is referred to as a sediment transport capacity predictor because, in Smart’s steep-slope experiments, sediment reached up to the flow surface and, at very steep slopes, occasionally left the flow altogether and became airborne for a short time. It is therefore more appropriate to regard the proposed equation as a formula giving the transport capacity, in the absence of fine suspended material, for alluvial materials with mean grain size greater than about 0.4 mm, in channel slopes of up to 11° (20%) in the absence of bed-armoring.

The channel flow depths in Smart’s experiments range between 1 and 10 cm. These shallow flow, steep slope conditions make Smart’s sediment transport predictor particularly relevant to Martian gullies. Due to the lack of laboratory experiments at steeper slopes, we have applied Smart’s transport predictor to the 15 - 20° slopes of Mars. However, because of the reduced gravity on Mars, a 0.1 m deep flow on a 18° slope applies the same stress to the channel bed (described below) as the stress in Smart’s experiments (a 0.06 m deep flow on a
The dimensionless Einstein transport parameter ($\phi$) is related to the discharge rate

$$q_s = \phi \left( \frac{\rho_s - \rho}{\rho} g \right)^{0.5} D_{50}^{1.5}$$  \hspace{1cm} (2.1)

where $\rho_s$, $\rho$, $g$, and $D_{50}$ are the sediment density, fluid density, gravity, and the median sediment diameter, respectively. In Smart’s work, $\phi$ and the non-dimensionalized shear stress, or Shield’s parameter ($\tau^*$), are empirically related

$$\phi = 4.2 S^{0.6} C_s \tau^{*0.5} (\tau^* - \tau_{c^*}^*)$$  \hspace{1cm} (2.2)

$$\tau^* = \tau \left( \frac{\rho_s - \rho}{(\rho_s - \rho)g D_{50}} \right)$$  \hspace{1cm} (2.3)

where $S$ is the local slope, $\tau = \rho gh \sin(S)$ is the bed shear stress, $h$ is the channel depth and $\tau_{c^*}$ is the slope-corrected critical Shield’s stress (the stress at which sediment transport is initiated) [Smart, 1984]. $C_s$ is a factor inversely related to friction ($\approx 4.5$) given by the ratio of the shear velocity ($\sqrt{ghS}$) to the mean flow velocity and is taken to be a constant.

Figure 2.3a shows $\tau_{c^*}$ and contours of $\tau^*$ as a function of grain size and channel depth for a slope of 20°. It also plots $\tau_{s^*}$, the critical stress at which sediment transport begins via the suspension of particles of a given grain size. $\tau_{s^*}$ is the stress at which the shear velocity ($\sqrt{ghS}$) equals the settling velocity for a particle of a given grain size [Dietrich, 1982]. The quantity $\tau^*$ increases with increasing channel depth and decreasing grain size, as expected. Due to the steep slope assumed, for the majority of parameter space plotted in Figure 2.3a $\tau^* < \tau_{s^*}$, indicating that bed load sediment transport is dominant, and the applied stress is much greater than the critical Shields stress ($\tau^* \gg \tau_{c^*}^*$). Smart’s experiments (indicated by the shaded region in Figure 2.3a) resulted in sediment transport as both bed load and suspended load, and Equation 1 represents the total flux from these two modes of transport.
Figure 2.3: a) Non-dimensional shear stress ($\tau^*$) contours, critical shear stress for transport ($\tau_c^*$), and the critical shear stress for suspension ($\tau_s^*$) plotted as a function of grain size and channel depth for a channel on a 20° slope based on flume experiments (conducted in the shaded parameter space) by Smart [1984]. b) Theoretical discharge rates (in m$^2$ s$^{-1}$) for water ($q_w$, equation 2.4) and sediment ($q_s$, equation 2.1) plotted as a function grain size and channel depth for a 20° slope. c) $q_w$ and $q_s$ as a function of slope compared with gully discharge estimates from Howard et al. [2008]. The box represents the average discharge estimate with the error bars representing the minimum and maximum estimates.

The sediment and water discharge rates are contoured in panel b of Figure 2.3 using the same axes as part a. Because $\tau^* \gg \tau_c^*$, Figure 2.3b shows that the water and sediment discharge rates scale as $h^{1.5}$ and are almost independent of grain size, as expected from equations 2.1-2.3. The thick black line corresponds to a channel depth to grain size ratio of 10:1 which is representative of Smart’s experiments. For slopes representative of Martian gullies, as long as $h$ exceeds the grain size $D_{50}$ by a factor of more than about 10:1, the results are independent of grain size. This affords a considerable simplification, since the relevant grain size for Martian gullies is poorly constrained. If this criterion is not satisfied, however, the sediment transport rate will depend on the mean grain size.

The sediment transport rate’s independence of grain size is due to the assumption that the grain friction factor ($C_s$) is constant with grain size. Theoretically, $C_s$ should decrease with increasing grain size (if channel depth remains constant) resulting in slower sediment discharge.
rates at larger grain sizes. The mean measured value of $C_s$ from Smart’s experiments is $6.2 \pm 1.9$, but seems to decrease with increasing grain size (as one would expect due to increased bed roughness) [Postma et al., 2008; Vollmer and Kleinhans, 2007; Shvidchenko and Pender, 2000]. $C_s$ also seems to decrease with increasing slope [Smart, 1984]. We have assumed $C_s = 4.5$ which corresponds to grains 10 times smaller than the channel depth at slopes of $11^\circ$ (the steepest in Smart’s experiments). Our assumption of $C_s = 4.5$ gives relatively small sediment and water discharge rates resulting in conservatively large timescale estimates. Increasing $C_s$ by a factor of 2 would decrease the formation timescale by the same factor. Note that changing $C_s$ will not affect the sediment concentration because both $q_s$ and $q_w$ (see below) are linearly dependent on $C_s$.

Figure 2.3c shows how both the sediment and water discharge rates (per unit channel width) vary with slope for a median grain size of 10 cm and a channel depth of 1 m. The water discharge rate shown in Figure 2.3b,c is calculated assuming a constant friction factor $C_s = 4.5$ and using

$$q_w = C_s h \sqrt{ghS}.$$  \hspace{1cm} (2.4)

where $q_w$ is the water discharge rate per unit width [Smart, 1984]. The box and error bar gives the mean and extrema in water discharge estimates from Howard et al. [2008] for the distal portions of Martian gullies based on slope, width, depth, and channel sinuosity measurements. There is a large uncertainty in the predicted discharge based on the range of possible channel dimensions and the application of terrestrial sinuosity-discharge relationships at low slopes to the steep slopes and lower gravity of Martian gullies. However, these estimates are in general agreement with the discharge rates we calculate using equation 2.4.
Table 2.1: Definitions and measured or theoretical values (or range of values) for parameters used in the numerical simulations

<table>
<thead>
<tr>
<th>Description</th>
<th>Symbol</th>
<th>Value(s)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>sediment density</td>
<td>$\rho_s$</td>
<td>3300 kg m$^{-3}$</td>
<td>[Kleinhans, 2005]</td>
</tr>
<tr>
<td>fluid density</td>
<td>$\rho$</td>
<td>1000 kg m$^{-3}$</td>
<td></td>
</tr>
<tr>
<td>gravity</td>
<td>$g$</td>
<td>3.7 m s$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>mean grain size</td>
<td>$D_{50}$</td>
<td>1, 10 cm</td>
<td>[Kleinhans, 2005]; assumed</td>
</tr>
<tr>
<td>grain friction</td>
<td>$C_s$</td>
<td>4.5</td>
<td>[Smart, 1984]</td>
</tr>
<tr>
<td>sediment packing density</td>
<td>$\phi'$</td>
<td>64%</td>
<td>[Parker et al., 1998]</td>
</tr>
<tr>
<td>apparent valley slope</td>
<td>$\alpha$</td>
<td>22$^\circ$</td>
<td>gives 33$^\circ$ dip of valley wall slope</td>
</tr>
<tr>
<td>initial channel width</td>
<td>$w$</td>
<td>3.0, 8.0$^a$ m</td>
<td>this study</td>
</tr>
<tr>
<td>fan opening angle</td>
<td>$\theta$</td>
<td>60$^\circ$</td>
<td>this study</td>
</tr>
</tbody>
</table>

[a] width increases by a factor of 3.65 downslope in order to conserve water volume.

### 2.3.2 Numerical Model

For a two-dimensional geometry, the evolution of channel elevation $z$ as a result of lateral sediment transport is governed by a continuity equation:

$$\frac{\partial z}{\partial t} = \frac{1}{\phi'} \frac{\partial}{\partial x} q_s$$

where $\phi' = 0.64$ is the sediment packing density [Parker et al., 1998] (1—porosity) and $x$ is the horizontal coordinate. We implement a numerical version of equation 2.5 in which gully and apron geometry in the third dimension is crudely accounted for.

For the gully geometry, we assume that the channel, or alcove, flanks remain at the angle of repose. Thus, as the channel downcuts into the steep background slope, additional material from the channel flanks must be transported downstream (Figure 2.4). The eroded volume is equal to the time and width-integrated sediment flux divided by the packing density. The finite-difference continuity equation in this case is given by

$$\Delta V = w \frac{\Delta q}{\phi'} \Delta t = \Delta x \Delta H_a \left( w + \frac{2H}{\tan(\alpha)} \right) + \frac{\Delta H^2}{\tan(\alpha)}$$

(2.6)
In order to reduce the channel elevation by $\Delta H_a$, the channel must also remove material from the alcove flanks as it infills at the angle of repose (see equation 2.6).

Here $\Delta V$ is the change in alcove volume in a grid-cell of length $\Delta x$ over a time step $\Delta t$, and $\Delta H_a$ is the corresponding change in alcove depth $H$. The channel width is $w$, $\Delta q$ is the difference in discharge rates $q_s$ between the upstream and downstream sides of the cell (cf. equation 2.5), and $\alpha$ is the apparent slope of the valley wall in the direction perpendicular to the channel ($\alpha = 22^\circ$ dipping perpendicular to a $25^\circ$ surface gives a true dip of $33^\circ$) (Figure 2.4). It may be verified that equation 2.6 reduces to a finite-difference representation of equation 2.5 if $\alpha = 90^\circ$, that is, the channel is two-dimensional.

Solving for $\Delta H_a$ using the quadratic equation gives

$$2\Delta H_a = -(w \tan(\alpha) + 2H_a) + \sqrt{(w \tan(\alpha) + 2H)^2 - 4 \frac{w \tan(\alpha) \Delta q \Delta t}{\phi^d \Delta x}}$$ (2.7)

Sediment conservation on the depositional apron requires using the Exner equation to calculate elevation change [Parker, 1991a,b]. Assuming the fan is radially symmetric with an opening angle, $\theta$ of $60^\circ$ (based on observations from our study location), the elevation change on the
depositional fan due to a change in sediment flux is given by

$$\Delta H_f(r) = \frac{-w \Delta q \Delta t}{r \theta \phi / \Delta x}$$

(2.8)

where $r$ is the horizontal radial distance downstream from the fan apex and $\theta$ is expressed in radians.

Equations 2.7 and 2.8 are used to update the channel profile $z(x)$ every time step. The location of the fan apex changes with time, and is the point at which the elevation of the channel bed exceeds the elevation of the original uneroded surface.

The initial model topographic profile consists of a 1.33 km long segment with a slope of $25^\circ$ which shallows to a slope of $2^\circ$ over an additional 0.67 km span. Water is allowed to flow into the model 300 m downslope from the top of the steep slope and occupies an initial channel that is 10 times smaller than the final channel dimensions. The channel depth and width increase linearly (in proportion to each other) to their fixed maximum values over a downslope distance of 20 m. This ”ramping up” of the channel dimensions is done in order to maintain numerical stability. The distance over which the channel is initiated has no physical significance, although we are effectively assuming that all the water involved in forming the gully is released over this 20 m reach of channel. As the slope shallows, the water velocity decreases (equation 4). Because water volume is conserved in our model, the channel cross-sectional area must therefore increase downslope due to the decrease in velocity resulting from a shallowing slope. For simplicity, we assume that the increase in cross-sectional area is accommodated by an increase in channel width [Finnegan et al., 2005]. In our model, the channel width increases by a factor of 3.65 from the steeply sloping portion to the shallow sloping portion. At the downslope boundary (after flowing over the break in slope and depositing most of its sediment), water is allowed to flow out of the simulation - carrying a small portion of sediment with it. During
the simulations mass conservation is checked by calculating and comparing the volume of the alcove to the volume of the depositional apron. Some sediment is lost (due to the flow leaving the model), but the volumes of the alcove and the apron are equal to within about 5%. The simulations are run until a volume of material \(6 \times 10^5 \text{ m}^3\) has been removed from the alcove region, consistent with the observations (Section 2.2).

### 2.3.3 Parameters and Assumptions

In order to determine the range of timescales and water volumes needed to form gullies via fluvial erosion, we must prescribe values for two unknowns: channel depth and sediment grain size. Channel depth is estimated by measuring channel widths in HiRISE images (observed channel widths range from 3 to 25 m) and assuming a width to depth ratio. Howard et al. [2008] use this approach to estimate discharge rates for gullies assuming a channel depth to width ratio of 1:8. Finnegan et al. [2005] give examples of how a river’s depth to width ratio depends on the substrate in which the channel is developed, and can vary from 1:5 for bedrock channels to 1:60 for gravel channels. To explore the range of timescales associated with variations in channel depth, we simulate two cases: one with initial channel dimensions of 1 m deep by 8 m wide, and one with a 0.1 m deep by 3 m wide initial channel.

As shown in Figure 2.3b, changing the grain size does not greatly influence the rate of erosion as long as grains are significantly smaller than the channel depth. However, the grain size is important in determining the regime (bed load vs suspension) that dominates fluvial sediment transport (Figure 2.3a). Previous studies have assumed a grain size of 10 cm for Martian channels [Kleinhans, 2005], although it remains relatively unconstrained with measured mean grain sizes ranging from 1 mm to 30 cm based on lander and rover imagery (ignoring dust)
[Golombek et al., 2003; Herkenhoff et al., 2004b]. The grain size frequency distribution is likely to be bimodal - especially in a fluvial environment and may have peaks in the size frequency at 1 mm and 10 cm [Kleinhans, 2005]. We will assume that the grains are sufficiently small that the results are independent of grain size and that sediment is transported as a mix of suspended and bed load material in a fashion similar Smart’s experiments. If transport via suspension is more efficient than Smart’s predictor (resulting in higher sediment concentrations), then our derived formation timescales will be overestimates.

Our model also assumes that water loss processes (freezing, boiling, infiltration) are negligible over the timescale of gully formation [Heldmann et al., 2005, cf.]. Heldmann’s simulations regarding the stability of liquid water flows on the surface of Mars found that if the flows contain soluble salts at a concentration of 0.02 mole fraction (twice that of terrestrial seawater) flows traveling down an 18° slope could extend for tens of kilometers due to the suppression of the brine solution’s vapor pressure. Based on work by Carr [1983], modest, pure water flows 10 cm deep could persist on the Martian surface under current climate conditions for a few hours before completely freezing. We assume that the concentration of sediment suspended in the flow is sufficient to neglect freezing and evaporation over the 2 km domain of our numerical simulations.

Simulations are performed with a grid size of $\Delta x = 2$ m, with 1000 total points, and a time step of 0.025 and 0.1 s for the 1 m and 0.1 m deep channels, respectively. This time step ensures that the Courant criterion is easily satisfied for typical sediment velocities of $\sim 1 \text{ m s}^{-1}$. 

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2.4 Results

The result from a typical sediment transport simulation are shown in Figure 2.5. This model assumed a channel 1 m deep by 8 m wide with a sediment grain size of 10 cm, and resulted in an alcove volume of \(6 \times 10^5\) m\(^3\) after 14 h of continuous flow. Figure 2.5a shows a perspective rendering of the final model geometry, which resembles the gullies seen in Figures 2.1, 2.2. The insert provides a visual comparison between the model result and the gully shown in Figure 2.2a shown at the same scale. The overall agreement is striking; the principal difference is that the rear of the alcove in the model is steeper than actually observed. This difference arises because material upslope of where the channel is initiated is not allowed to move downslope in the model and results in a steep cliff forming above the point of channel initiation. Figures 2.5b-d show the evolution of channel elevation, slope and downcutting with time. The rate of downcutting decreases with time, because as the channel deepens more material has to be removed from the channel flanks (Figure 2.4). The apex of the fan starts at the base of the steep slope, but propagates upslope with time as the local slope changes, resulting in a depositional fan extent that is in close agreement with observations.

The main result from this particular model is that gully formation can take place extremely rapidly, primarily because the steep slopes - even with relatively small channels - result in rapid sediment transport rates. The roughly ten-hour timescale associated with an 1 by 8 m channel increases to 40 days when the channel is 0.1 by 3 m. In the 1 by 8 m channel case, the maximum water discharge rate is 45 m\(^3\) s\(^{-1}\) and the total volume discharged is \(1.8 \times 10^6\) m\(^3\) for the entire simulation. In the 0.1 by 3 m channel case, the maximum water discharge rate is 0.5 m\(^3\) s\(^{-1}\) and the total volume of water involved in eroding the gully is also \(1.8 \times 10^6\) m\(^3\) based on the water discharge rate from equation 2.4. This total volume implies a sediment:water
Figure 2.5: a) Perspective view of a sediment transport simulation lasting 14 h under constant flow conditions for a channel 1 m deep by 8 m wide with a sediment grain size of 10 cm. $6.0 \times 10^5$ m$^3$ of sediment was transported over the course of this simulation. Topographic, slope, and change in elevation profiles are shown every 5000 s (one tenth of total simulation time) in parts b, c, and d, respectively. A visual comparison between the simulation and an observed gully at the same scale is shown in the inset.

Volume ratio of 33% by volume (equivalent to a sediment concentration of 25% by volume) for both channels. In the next section, we will discuss the significance of this finding, and compare our results with the expected discharge rates from plausible sources for liquid water.

These water, sediment volumes and timescale results depend on our assumptions made in Section 2.3.3. In general, our assumptions have been conservative, giving results that give longer formation timescales. For instance, we have used relatively small channel sizes, but if we were to increase our modeled channel width by a factor of two, then our timescale would decrease by the same factor. Similarly, if we chose a larger value for $C_s$, then the timescale would be shorter than the estimates given above.
2.5 Discussion

The results presented above suggest gully formation timescales, assuming continuous discharge, of days to months, and peak water discharge rates of $\sim 45 \text{ m}^3\text{s}^{-1}$ or $\sim 0.5 \text{ m}^3\text{s}^{-1}$ for channels 1 by 8 m and 0.1 by 3 m, respectively. Quantitative estimates of discharge rates by Heldmann et al. [2005] and Howard et al. [2008] give values of 30 m$^3$ s$^{-1}$ and 28 m$^3$ s$^{-1}$, respectively, similar to our value for the larger channel. Heldmann et al. [2005] also obtained flow durations of $\sim 10^3$ s, somewhat shorter than our estimates (but see below). If gullies are indeed fluvial features, there are two hypothesized sources of water: either groundwater discharge [Mellon and Phillips, 2001, e.g.] or snowpack melting [Christensen, 2003, e.g.]. Below we examine both hypotheses in the light of our numerical model results.

2.5.1 Groundwater Discharge

Groundwater discharge was the first explanation proposed for the formation of Martian gullies [Malin and Edgett, 2000]. We test this hypothesis using the duration and water volume results from our numerical model above, together with theoretical calculations of groundwater discharge from a permeable aquifer. The simplified physical situation is shown in Figure 2.6a: fracture of a surface cap (e.g. an ice plug) perhaps due to increased aquifer pressure [Mellon and Phillips, 2001] permits water to flow out of an aquifer at a rate determined primarily by the permeability. The flow rate will decline with time as drawdown of the aquifer occurs, until flow eventually ceases (e.g. due to freezing).

As described in appendix A, an approximate expression for the resulting discharge rate can be obtained assuming that the channel depth is small compared to the aquifer thickness. In a 2D Cartesian geometry this discharge rate (per unit width) may be approximated by:
\[ q_w \approx T^{3/2} \left( \frac{\kappa \rho g \phi}{\pi \mu t} \right)^{1/2} \] (2.9)

Here \( T \) is the aquifer thickness, \( \kappa \) is the permeability, \( \rho = 1000 \text{ kg/m}^3 \) is the density of water, \( g = 3.71 \text{ m/s}^2 \), porosity (\( \phi \)) is 0.2, \( t \) is time, and \( \mu = 10^{-3} \text{ Pa s} \) is the dynamic viscosity of water. For a given value of \( T \), equation 2.9 actually overestimates the discharge rate by a factor of 4 or more (see appendix A). Integrating this function over time gives the total volume discharged

\[ Q_{tot} \approx w_a T^{3/2} \left( \frac{\kappa \rho g \phi t}{\pi \mu} \right)^{1/2} \] (2.10)

where \( w_a \) is the width of exposed aquifer over which discharge occurs. Figure 2.6b plots the required aquifer thicknesses and permeabilities (solid line) needed to match the model timescales and total discharge using equation 2.10. In this plot, \( w_a \) is assumed to be 100 m - comparable to the alcove width. As the aquifer permeability increases, the aquifer thickness required to deliver the inferred water volume over the inferred duration decreases. Longer flow durations permit lower permeabilities for the same aquifer thickness. Keep in mind that we are underestimating the actual required aquifer thickness because equation 2.9 over predicts the discharge rate.

Because the aquifer thickness cannot exceed the total slope relief of roughly 300 m \( (T_{max}) \), the aquifer permeability must exceed roughly \( 10^{-9} \text{ m}^2 \) for the 1 by 8 m channel case, and \( 1.5 \times 10^{-11} \text{ m}^2 \) for the 0.1 by 3 m channel case. Although the aquifer permeability is not well constrained, work by Manga [2004] suggests a regional permeability of crust containing basaltic lava flows of \( \sim 10^{-9} \text{ m}^2 \). However, other authors suggest a less permeable substrate, averaging \( \sim 10^{-12} \text{ m}^2 \) [Harrison and Grimm, 2009] for the Martian regolith. Our results for a scenario involving a single groundwater discharge event lie within the range of plausible permeabilities (from \( 10^{-12} \) to \( 10^{-9} \text{ m}^2 \)) for the 0.1 by 3 m channel case. However, the 1 by 8 m
channel requires aquifer permeabilities that are probably too high to be reasonable.

As flow proceeds, the aquifer will experience drawdown, where the lateral drawdown distance $\delta$ is roughly $\delta \sim (Q_{tot}/(T\phi))^{1/2}$. Since the gullies are separated by a few hundred meters, drawdown distances in excess of this value will result in one gully cannibalizing the water supply of another. At present it is not clear whether multiple gullies formed simultaneously or not. If they did, then $\delta$ must be less than half the gully separation distance ($\delta_{max} \approx 150$ m).

Figure 2.6b also plots $\delta$ as a function of permeability for the calculated value of $T$. As the permeability increases, so too does the drawdown distance, as expected.

If gullies form simultaneously from a single groundwater discharge event, then there is no permeability that satisfies both the aquifer thickness and drawdown length restrictions (indicated by the shaded gaps in Figure 2.6b). Based on the results shown in Figure 2.6b, either gully water is discharged from a thick (300 m), permeable ($10^{-11}$ m$^2$) aquifer as a single event lasting about 40 days - drawing water from a radial distance of $\sim$ 200 m (beyond the maximum value of 150 m suggested by the observations), or gully formation takes place over many episodes of short-lived ($\sim$ hours to days) groundwater discharge events from a thinner aquifer.

Although it is hard to assess the likely thickness of potential aquifers, clay-bearing layered deposits in Mawrth Vallis appear to be a few tens of meters thick [Wray et al., 2008], while extensive volcanic layers in Valles Marineris are typically 5-50 m thick [McEwen et al., 1999]. We conclude that the gullies in our study region are unlikely to have formed by a single discharge event; more plausibly, they were formed by multiple, short-lived discharges from a relatively thin aquifer. Heldmann et al. [2005] also concluded that short-lived ($\sim 10^3$ s) flows were likely responsible for gully formation.
Figure 2.6: a) Illustration of groundwater discharge rate, $q_w$, from an aquifer of thickness, $T$, and permeability, $\kappa$ where the dotted line represents the water table. b) The combination of $T$ (solid line) and $\kappa$ necessary to discharge $1.8 \times 10^6$ m$^3$ of water in 40 d through a 0.1 by 3 m channel based on equation 2.10 for an aquifer width of 100 m. The dashed line gives the drawdown distance, $\delta$, into a radially symmetric aquifer for the given permeability and calculated aquifer thickness. c) same as b) for a 1 by 8 m channel involving the same total water volume and a timescale of 10 h.

Assuming groundwater sourced from a 30 m thick aquifer is responsible for fluvial activity in gullies, then it would take 10 discharge events, drawing water from a radial distance of 150 m with a porosity of 20% to account for $2 \times 10^6$ m$^3$ of water. If each discharge event lasted 7 days, the permeability required is then $10^{-9}$ m$^2$, consistent with the estimate by Manga [2004]. During each discharge event, this scenario would give discharge rates comparable to that of our 0.1 by 3 m channel simulation. Lower permeabilities or thinner aquifers would give water discharge rates substantially smaller than those determined from our model, and would become too small to transport a significant volume of sediment due to fluvial processes.

Since at least some gullies are only a few My old [Schon et al., 2009, e.g.], this multiple discharge hypothesis requires a recurrence interval of $\sim 100$ kyr, similar to the periodicity of the Martian obliquity cycle. This age also implies a recharge rate of $\sim 1$ m$^3$/yr, or about
$10^{-4}$ m/yr. The obliquity timescale has been suggested to drive the episodic release of groundwater into gullies by Mellon and Phillips [2001] due to pressurization of a saturated, confined aquifer by the propagation of a freezing front. As we will address in the next section, melting of a snowpack due to climate variations may provide the aquifer recharge necessary to make a groundwater-fed gully hypothesis feasible.

### 2.5.2 Snowpack Melting

Direct runoff from melting of an ice-rich mantle is an alternative to a groundwater source for explaining fluvial activity in gullies. However, snowpack melting and subsequent storage in an aquifer is another way in which water may be delivered to gully alcoves. Observations in support of a snowpack melting hypothesis include the global distribution of gullies along mid-latitude bands [Dickson et al., 2007; Heldmann et al., 2007], their occurrence on pole-facing slopes at mid-latitudes in the southern hemisphere [Heldmann and Mellon, 2004], their presence on isolated massifs [Christensen, 2003], and their association with an ice-rich mantling unit [Bridges and Lackner, 2006]. At high obliquity, polar water ice will sublimate during summer and precipitate at lower latitudes [Mellon et al., 1997; Mellon and Jakosky, 1995]. If the seasonal snowpack (~1 m thick) is unprotected by dust, it will sublime and melt over 1 to 2 Martian years [Williams et al., 2008, 2009]. Alternatively, if the ice is protected by dust, it could potentially survive to the present day, resulting in an ice-rich mantling unit located in areas of minimum solar insolation. This mantle could potentially produce meltwater runoff at the present day, suggesting gullies could be currently active [Christensen, 2003]. However, the location of gullies on pole-facing crater slopes at mid-latitudes is not consistent with locations of maximum melting according to Williams et al. [2008].
Based on energy balance models of a snowpack undergoing melting and sublimation, a small amount of meltwater available for runoff can be produced for a few days each Martian year under a wide range of atmospheric and obliquity states. Based on work by Clow [1987] and Williams et al. [2008] 1 - 2 mm of runoff per m$^2$ can be accumulated over a few days during each Martian year. The differences between these models is the location and timing of the predicted melting. Regardless of the model, the maximum discharge rate predicted from snowpack melting is small. A melting rate of 0.25 mm h$^{-1}$ per m$^2$ integrated over a gully alcove $6 \times 10^4$ m$^2$ in area gives a maximum discharge rate of 0.005 m$^3$ s$^{-1}$. Such fluid discharge rates are orders of magnitude smaller than the discharge rates we inferred in Section 2.4, and will not result in significant sediment transport as either bed load or suspended-load.

In Williams et al. [2008], the obliquity must be high enough for atmospheric precipitation of snow to occur (requiring obliquities larger than $\sim 35^\circ$). As the snow seasonally melts, runoff is maximized at locations with the warmest regolith, and the thickest snowpack. The locations at which melting is maximized are equator-facing slopes at 50$^\circ$S, or pole-facing slopes at 20$^\circ$S. In order to generate the total amount of water predicted from our sediment transport modeling, these high obliquity snowpacks would have to seasonally melt for $\sim$15 kyr assuming 2 mm of runoff every Martian year, and an alcove area of $6 \times 10^4$ m$^2$. This water would need to be stored and then released rapidly in order to generate the discharge rates necessary for sediment transport via fluvial processes. Since the obliquity in recent (< 3 Ma) cycles can exceed 35$^\circ$ for up to a cumulative period of 20 kyr [Laskar et al., 2002], such an eventuality cannot be dismissed entirely. Nonetheless, for this mechanism to work, it requires a fairly complicated series of events: near-surface melting followed by downwards fluid infiltration and storage for tens of kyr, followed by sudden, rapid release.
2.5.3 Alternative Sources of Water

Another possible source of aquifer recharge is meltwater produced from geothermal heating of surface or subsurface ice. Geothermal melting of surface ice is an unlikely water source because it would require thick ice sheets (> 1 km) and an elevated heat flux (0.1 W m\(^{-2}\)) [Russell and Head, 2007] to generate liquid water. The estimated thickness of a recently deposited ice-rich mantle is two orders of magnitude smaller (between 1 and 10 m) [Mustard et al., 2001] than the ice thickness required to generate liquid water. Alternatively, melting could occur at the base of a subsurface ice table if it is covered by a thick (100s of meters), very low thermal conductivity 0.045 W m\(^{-1}\) K\(^{-1}\)) material with a moderate heat flux of 0.03 W m\(^{-2}\) [Mellon and Phillips, 2001]. This insulating layer could be composed of dry unconsolidated regolith covering an ice table.

Hydrothermal circulation is another possible mechanism for delivering water to gully alcoves. The heat needed to drive circulation requires an elevated geothermal flux - either from a magmatic intrusion, or by impact cratering. However, gully activity has not been observed to correlate with recent volcanic activity, nor can intrusions reproduce the observed distribution of gullies along mid-latitude bands. Hydrothermal systems generated by newly-formed impact craters would likely be short-lived (< \(10^4\) yrs) for craters 20 km in diameter exposed to freezing conditions at the surface [Barnhart et al., 2010], and are unlikely to direct fluid flow up to the rims of craters, unless guided by permeability structures.

2.5.4 Alternative Sediment Transport Processes

As mentioned in Section 2.3.1, this project focuses on a transport capacity predictor from Smart [1984] that is applicable in the absence of fine suspended sediment and in the
absence of bed-armoring for slopes up to $11^\circ$ in Earth-like gravity. Due to the lack of laboratory experiments at steeper slopes, we have applied Smart’s transport predictor to the 15 - 20° slopes characteristic of Martian gullies. The lower gravity on Mars means that transport on these slopes is treated appropriately by the Smart predictor, at least for shallow channels (Figure 2.3a); nonetheless, further laboratory experiments [Coleman et al., 2009, e.g.] would be helpful. Alternatives to the sediment transport processes involved in Smart’s experiments involve flows with much higher concentrations of sediment such as hyperconcentrated flows, wet debris flows, and dry debris flows (see below).

We have neglected fine, suspended sediment transport in our model. If this mechanism is important, then the gullies could have formed even more rapidly. This in turn would require a thicker aquifer and/or higher permeability (equation 2.10), making a fluvial origin less plausible. Alternatively, if the channel dimensions are smaller than the ones used in our simulations, gully formation could have taken place over the same time periods given above (if suspended sediment transport is accounted for), but potentially involving less water. However, subaerial water flows transporting sediment via suspension could persist for tens of hours under current Martian conditions [Heldmann et al., 2005; Kuznetz and Gan, 2002] resulting in sediment deposition at shallower slopes than sediment transported as bed load, and likely forming a thinner, more extensive depositional apron than what is observed on Mars [Parker, 1999].

Another assumption we have made in the simulations is that the regolith is unconsolidated. This assumption is based on the incoherent appearance of the dusty surface into which gullies tend be incised and the lack of cliff-forming units at our study location. However, the Martian regolith may have some strength due to the presence of bed rock, indurated regolith, or an ice table. If this is the case, then fluvial transport is even less likely because the depth of wa-
ter in the channel must be large enough to overcome the cohesive strength of the regolith before transport can begin. This will mean that more water is required for fluvial transport compared to the unconsolidated case that is assumed in our numerical work. This assumption is made in order to estimate the minimum water volume and timescale required for gully formation, and because the shear strength of regolith in gully environments is not observationally constrained. If the substrate has strength, then more water would be required to carve the gullies than the volumes suggested in Section 2.4, making it harder to provide a water source, suggesting that debris flow mechanisms (or dry mechanisms) are to be preferred. Future studies to determine the strength of the regolith in gully environments are needed to determine the mechanism(s) responsible for gully erosion.

On Earth, the characteristics of the local colluvium can determine whether debris flow or streamflow processes dominate [Blair, 1999]. On Mars, the permeability of the regolith is likely to play a similar role. Furthermore, the deposition of atmospheric dust and/or the potential presence of an ice table in the near surface may influence the permeability and therefore the mechanism by which sediment is transported.

The time-averaged sediment concentration of 25% in our model are equivalent to the maximum concentrations in Smart’s [1984] experiments, and suggest that the transport processes for Martian gullies may fall in the transition region between streamflow and hyperconcentrated flow [Vallance, 2000]. It is possible that fluvial processes are not the primary mode of sediment transport in gullies, but may play a secondary role in modifying the surface [Head et al., 2008] with wet or dry debris flows being responsible for the majority of sediment transport, as is the case in steep upland channels on Earth [Lancaster and Casebeer, 2000; Hartmann et al., 2003]. As noted in Section 2.1, it is very hard to distinguish streamflow- and debris flow-
dominated fans purely by remote sensing, although the absence of pronounced levees arguesagainst the latter process [Costard et al., 2002, c.f.]. On the other hand, it has been argued that
the slope-area relationships observed in the few Martian gullies for which we have high resolu-
tion (∼ 1 × 1 m grid spacing) topography suggest debris flow processes are at work [Lanza
et al., 2009]. From a purely theoretical viewpoint, debris flows are attractive because they re-
quire much less water to transport the same sediment volume. Theoretical modeling of debris
flows has received less attention than that of streamflows, but some work has been done [Stock
and Dietrich, 2006, e.g.] and should certainly be applied to Mars in the future.

2.6 Conclusions

In this chapter we have focused on the end-member hypothesis that fluvial processes
are responsible for gully formation. Fluvial sediment transport on steep gully slopes is rapid.
Channels 1 by 8 m and 0.1 by 3 m in dimension transport a sediment volume of 6 × 10^5 m^3 in
10 h and 40 d, respectively, under constant flow conditions. Both the 1 by 8 m channel and the
0.1 by 3 m channel require 1.8 × 10^6 m^3 of water, resulting in a sediment:water volume ratio of
33% (equivalent to a sediment concentration of 25% by volume).

Except in the unlikely event that aquifer thicknesses exceed 300 m, a single event
of groundwater release is not plausibly responsible for gully formation. About ten discharge
episodes could generate the correct gully morphology assuming a ∼ 30 m thick aquifer and per-
meabilities that are consistent with other estimates [Manga, 2004]. In the case where multiple
episodes of discharge occurs, the maximum interval between events is similar to the periodicity
of obliquity cycles which were shown by Mellon and Phillips [2001] to result in pressuriza-
tion of a hypothetical confined aquifer. The required aquifer recharge rates can be met by
high obliquity seasonal melting of snow packs (perhaps aided by the presence of salts), or, less likely, by melting due to geothermal heat assuming a thick ($\sim$ 100 m), low thermal conductivity ($0.045 \text{ W m}^{-1} \text{ K}^{-1}$) surface material [Mellon and Phillips, 2001]. Although there are no observational constraints on the gully formation timescale, there are examples of episodicity of fan deposition [Schon et al., 2009], consistent with the multiple-discharge hypothesis.

Snowpack melting cannot produce the water discharge rates necessary for fluvial sediment transport, unless long-term (tens of kyr for a single discharge event) storage of the resulting meltwater occurs [Christensen, 2003, cf.]. Also, basal melting of an ice sheet due to geothermal heat would require ice thicknesses on order of 1 km [Russell and Head, 2007], and, thus, is unlikely to be a source for gully water in the recent past.

In reality, gully formation may include a mixture of streamflow and debris-flow processes. If the substrate has strength, then more water would be required to carve the gullies than the volumes suggested in Section 2.4, giving support to debris flow mechanisms (or dry mechanisms). Differentiating between these end-member processes by remote sensing is difficult, illustrating the need for future Mars rovers. High resolution topography and geological and hyperspectral mapping may help to test some aspects of the fluvial end-member hypothesis, for instance the estimated number ($\sim$10) and recurrence interval ($\sim$100 kyr) of discharge events [Schon et al., 2009, c.f.], the thickness of candidate aquifers, and the strength of the regolith. From a theoretical standpoint, applying terrestrial models of debris flows to Mars [Stock and Dietrich, 2006, e.g.] will permit this end-member hypothesis to be tested against current or future observations.
Chapter 3

Lobate debris aprons

3.1 Background

Lobate Debris Aprons (LDAs) surrounding plateaus, massifs, and valley walls at mid-latitudes are a present-day reservoir of ice in the Martian near-surface [Holt et al., 2008; Plaut et al., 2009]. Since they were identified in Viking images, LDAs have been interpreted as ice-related features produced from the viscous flow of a mixture of debris and ice similar to rock glaciers on Earth [Squyres, 1979; Carr and Schaber, 1977; Squyres and Carr, 1986; Carr, 1996]. However, the recent findings from the Shallow Radar (ShaRAD) instrument on board Mars Reconnaissance Orbiter suggest LDAs contain relatively pure ice with thicknesses ranging from 300 to 700 m [Holt et al., 2008; Plaut et al., 2009]. These mid-latitude ice reservoirs are found in geographic association with other young, water- and ice-related features such as a mid-latitude mantling unit [Mustard et al., 2001], gullies [Malin and Edgett, 2000], and viscous flow features [Milliken et al., 2001]. Examining the past climate conditions responsible for LDA formation may provide insight into the formation of these other mid-latitude features [Head et al., 2010].
These late Amazonian (∼ 100 Ma) features record a significant episode of widespread hydrologic activity on Mars. Episodic climate variations resulting from changes in the tilt of Mars’ rotation axis can potentially redistribute water ice between the poles and lower latitudes over $10^5$ - $10^6$ yr timescales [Laskar et al., 2002; Fanale et al., 1986; Mellon and Jakosky, 1995]. The lack of present-day ice accumulation in regions where LDAs are found suggests that past episodes of climate change are responsible for forming these massive ice deposits [Madeleine et al., 2009].

Before data from ShaRAD was available, quantitative study of LDAs was limited to the analysis of topography and visual data to determine the rheology of LDA ice based on profile shape [Li et al., 2005; Mangold and Allemand, 2001; Pierce and Crown, 2003], to determine whether sub-surface ice was present at all [Ostrach et al., 2008; Kress and Head, 2008] and to constrain the age of LDAs based on crater counts [Mangold, 2003; Baker et al., 2010]. Below we will develop a numerical model of LDA ice flow using ShaRAD data constraining both the thickness and dust concentrations of LDAs [Holt et al., 2008; Plaut et al., 2009], combined with experimental results from deformation of ice/ice-rock mixtures [Goldsby and Kohlstedt, 2001; Durham et al., 1992; Mangold et al., 2002] and theory of ice grain growth in the presence of solid particulates [Durand et al., 2006; Barr and Milkovich, 2008]. In particular we make use of ShaRAD’s recent observations and LDA age constraints from crater density measurements to better constrain the rheological parameters that produce LDA lengths and thicknesses that are in agreement with observations.

The rest of the chapter is organized as follows: first, we will address topographic, visual, and radar observations pertaining to LDAs and how these observations place constraints on the rheology, age, and evolution of LDAs. Next, in section 3.3, we discuss the rheology
of ice and ice-dust mixtures constrained by recent experimental, theoretical, and observational work, and how the rheology used in our model differs from previous work. A description of our model is given in sections 3.4 and 3.5, and results from our simulations are given in section 3.6. Finally, a discussion of our results and conclusions from this study are given in sections 3.7 and 3.8.

3.2 Observations

Debris apron complexes consisting of a single apron or, more commonly, a laterally extensive mass of multiple, coalesced aprons derived from a common source are found in several distinct regions on Mars [Pierce and Crown, 2003]. In the southern hemisphere, a relatively small population of LDAs are found in the Argyre basin, and a more abundant population of at least 90 debris apron complexes are found to the east of Hellas impact basin [Squyres and Carr, 1986; Squyres et al., 1992; Pierce and Crown, 2003]. The region with the highest abundance of LDAs on Mars is the fretted terrain in the northern mid-latitudes (∼ 40°N) where at least 191 LDAs can be found [Kochel and Peak, 1984; Crown et al., 2006]. In total, these deposits may contain ∼ 10^5 km^3 of ice, or about 10% of the volume of the northern polar cap [Crown et al., 2006].

The fretted terrain is a region along the highland-lowland boundary containing angular mesas separated by flat floored valleys [Sharp, 1973]. Debris aprons up to 800 m thick and 30 km long emanate from the 1-2 km high scarps surrounding these mesas [Lucchita, 1984]. Studies using Mars Global Surveyor (MGS) data sets have described the geometry and distribution of aprons as well as providing crater-age dates and constraints on the rheological parameters of the ice [Mangold and Allemand, 2001; Pierce and Crown, 2003; Li et al., 2005]. Mars Or-
biter Laser Altimeter (MOLA) profiles of LDA surfaces show that they are gently sloping at \(\sim 1^\circ-4^\circ\) with distal margins that steepen to up to \(7^\circ\) [Pierce and Crown, 2003].

The presence of a buried reflector in published radargrams [Holt et al., 2008; Plaut et al., 2009] suggests deposits of massive ice \(\sim 400 \text{ m}\) thick have flowed outward from adjacent massifs. In addition to the ice thickness, the low \((< 13 \text{ dB}/\mu\text{s})\) one-way attenuation rate suggests the ice is contaminated by no more than a few tens of percent dust by volume [Plaut et al., 2009; Heggy et al., 2007].

Neutron spectroscopy observations by Feldman et al. [2004a], as well as thermal emission observations by Bandfield [2007] indicate near-surface ice is present poleward of \(\pm 50^\circ\) latitude. These methods sample the uppermost meter of regolith and suggests the surface layer of LDAs is ice-poor. A lack of near-surface ice on LDAs, together with a lack of an observed near-surface reflection from radar limits the thickness of an overriding ice-poor regolith deposit to between \(\sim 1\) to \(30 \text{ m}\) thick [Boynton et al., 2002; Feldman et al., 2004a; Plaut et al., 2009]. Such a layer is likely sufficient to halt the loss of ice by sublimation [Jakosky et al., 2005; Chevrier et al., 2007; Bryson et al., 2008]. An additional constraint on the surficial ice-depleted layer is provided by the change in crater morphology on LDAs in Deuteronilus Mensae between crater diameters of \(\sim 400\) and \(500 \text{ m}\) [Ostrach et al., 2008] implicating the influence of ice in the cratering process when impactors penetrate to a depth of a few tens of meters.

A regional crater age dating study covering many LDA surfaces in Deuteronilus Mensae by Mangold [2003] gave an overall age of several 100 Mys based on the largest (few 100 m) diameter craters. Other studies suggest younger ages for LDA surfaces, but fall within the late Amazonian between 50 and 300 Ma [Mangold, 2003; Li et al., 2005; Hartmann, 2005; Morgan et al., 2009; Baker et al., 2010]. Mangold also observed the preferential removal of craters less
than 175 m in diameter over ~ 10 My timescales, perhaps due to a combination of sublimation pitting, small scale mass wasting, and dust removal/deposition. A lack of deformed craters on LDA surfaces may suggest that deformation is slow, or has stopped at the present time Carr [2001].

If LDA ice was deposited during a sustained period of high obliquity, then scenarios for the Martian obliquity history determined by Laskar et al. [2004] suggest LDAs are most likely older than 50 My, and probably older than 100 My based on the likelihood of Mars having an obliquity of 50° or more (Figure 1.2). There is some evidence for very old (> 1 Ga) LDAs at lower latitudes (25-30°N) that have since been removed leaving shallow depressions around the massifs and plateau walls they used to occupy [Hauber et al., 2008]. These features may represent an earlier generation of LDA formation, or may have formed concurrently with the LDAs we observe today. To place tighter constraints on their age and evolution, we must turn to numerical models (see below).

In a study analyzing the topographic profiles of 36 LDAs in Mareotis Mensae, Deuteronilus Mensae and Protonilus Mensae (all at northern mid-latitudes), Li et al. [2005] found differences in the shape of LDA profiles and classified LDAs into types I, II and III. In the absence of erosion or processes that remove ice from within LDAs, the shape of LDA topographic profiles is determined by the rheology of the ice or ice-dust mixture that make up these deposits. In the classification presented by Li et al. [2005], which will be referred to later, type I most closely matches the convex topographic profile expected from a Newtonian rheology (n = 1, see next section) and is most consistent with the presence of ice; type II has a convex shape, but is less pronounced than type I and type III LDAs have profiles that are only slightly convex. Li et al. [2005] suggest that LDAs of type II and III may have different ice concentrations or sublima-
tion rates than type I LDAs which results in the shape differences. A further discussion of new observations in regard to this study is given in Section 3.7.

3.3 Ice Rheology

3.3.1 Previous Work

Previous efforts to investigate the evolution of ice-rich deposits on Mars generally fall within two categories: observation-based models that attempt to constrain the rheology based on the shape of topographic profiles of the ice-rich deposit, and experiment-based models that assume a rheology based on deformation experiments and/or terrestrial observations of ice flow and use a time-dependent flow model to constrain other variables such as the age, temperature, precipitation rate, or dust fraction.

Using the shape of an ice-rich deposit to determine the rheology assumes that erosion and ice removal processes have not significantly influenced the shape of the ice deposit, and that the topography data used to constrain the rheology does not sample convergent or divergent flow. Although Winebrenner et al. [2008] fit topography from the northern polar layered deposits to a model that accounts for the aggradation and ablation of ice as well as flow divergence, prior models applied to LDAs do not account for these effects [Mangold and Allemand, 2001; Li et al., 2005; Bourgeois et al., 2008]. Comparing model-derived profiles to observations requires a careful selection of observational data. This is especially true near the toe of the ice-rich feature where topography is most sensitive to differences in rheology (see Winebrenner et al. [2008] Figures 6-8), but, unfortunately, the toe is where slopes are steepest and where modification by mass wasting, sublimation, and erosion is most likely.

The alternative method, using a time-dependent flow model to predict the form of the
observed ice-rich deposit requires that an applicable, experiment-based ice rheology is used, and that reasonable assumptions are made regarding the initial and boundary conditions. Also, determining when the simulation is completed requires a justifiable criterion. The first attempt to model LDAs in this fashion was made by Colaprete and Jakosky [1998] using an ice rheology based on Patterson [1994] and Glen’s flow law. Glen [1955] used laboratory experiments on (warm) ice to derive a flow law in which the strain rate of ice is proportional to the stress raised to the third power ($n = 3$). The model developed by Colaprete and Jakosky [1998] accounted for the viscous effect of incorporated dust and also considered the accumulation of ice from precipitation. They assumed a flow distance of 5 km represented the distance LDA and lineated valley fill deposits had advanced, but the model was only run for a maximum of 2 My - assuming that these were very young features. This assumption is not supported by more recent crater counts (see Section 3.2). Colaprete and Jakosky [1998] suggested that these ice deposits were likely formed by clean ice, but at the time of this study there were no independent constraints on the dust content nor on the basal slope (which they assumed was $0.2^\circ$). In general, endeavors to model the time-dependent flow of LDAs have attempted to place constraints on the ice temperature, precipitation rate, timescale, ice content, and/or ice rheology needed for LDA formation. However, in order to constrain one of these parameters these studies had to make assumptions about the remaining, unconstrained parameters in addition to assuming a particular subsurface geometry. Recent observations (such as from ShaRAD [Plaut et al., 2009; Holt et al., 2008]) and discoveries regarding the climate history of Mars (such as the recent obliquity history [Laskar et al., 2002]) have led to better constraints on many of these parameters.

These findings prompted Fastook et al. [2008] to go a step further and use climate model-predicted precipitation rates from different obliquity histories together with a glacial
flow model including observed Martian topography to constrain the climate history that best reproduced the relict glacial features on the Tharsis Montes. However, this model (and many others) uses Glen’s flow law to model ice flow rather than using a rheology consistent with more recent ice and ice-dust mixture deformation experiments [Goldsby and Kohlstedt, 2001; Durham et al., 1992; Mangold et al., 2002]. Although implementing this more complex ice rheology in a numerical model introduces more variables, such as the grain size of the ice, simplifications regarding how grain size is controlled by deforming ice in the presence of dust can be made [Barr and Milkovich, 2008; Kieffer, 1990] to make numerical models of Martian ice deposits more realistic without introducing more uncertainty. Models implementing a more complex rheology for ice and ice-dust mixtures have already been developed and applied to the Martian polar layered deposits [Pathare et al., 2005] and the shells/mantles of icy satellites [Barr et al., 2004; Barr and McKinnon, 2007]; here we extend these types of models to LDAs.

3.3.2 Our approach

In order to model the flow of ice at mid-latitudes on Mars, we must specify a rheology that is appropriate for the ice grain size(s) within, and the range of differential stress experienced by, LDAs. First, the differential stress ($\tau$) and strain rate ($\dot{\epsilon}$) experienced at the base of an ice sheet of thickness $h$ are:

$$\tau(z) = \rho g (h - z) \left( \frac{\partial h}{\partial x} + \sin \theta \right)$$

(3.1)

$$2\dot{\epsilon} = \frac{\partial v}{\partial z} = 2 A \tau^n d^{-p}$$

(3.2)

where $v$ is velocity in the $x$ direction, $n$ is the stress exponent, $d$ is the ice grain size, $\rho$ is the ice (or ice-dust mixture) density, $g$ is gravity, and $\theta$ is the basal slope. The quantities $z$ and $x$ are the slope-perpendicular and downslope coordinate directions, respectively, and $z = 0$ at the base of
the ice. The dependence of strain rate on the applied stress and grain size through the values of $n$ and $p$ is a function of the creep regime. $A$ is a temperature- and dust fraction-dependent flow parameter defined below.

As discussed in Goldsby and Kohlstedt [2001], polycrystalline ice can deform via diffusion, grain size sensitive (GSS), or dislocation creep depending on the applied stress and the ice grain size. The total strain rate is calculated by combining the strain rates from these three contributing mechanisms (Goldsby and Kohlstedt [2001], equation 3). Because a single deformation mechanism is generally dominant at any particular stress, we explore different ice rheologies by using a value of either $n = 2$ or $n = 4$ depending on the process that controls the ice grain size (grain boundary pinning or dynamic recrystallization, respectively, see Sections 3.3.3, 3.3.4 below). The creep rate in the GSS regime ($n = 2$) depends on ice grain size ($d$) with an exponential of $p = 1.4$, whereas deformation via dislocation creep ($n = 4$) is grain size independent ($p = 0$) [Goldsby and Kohlstedt, 2001]. The equation for the rheological parameter, $A$, is

$$A_i = A'_i e^{ \phi_b}$$

(3.3)

where $i = \text{disl}, \text{gss}$ denotes the coefficient for dislocation creep and grain size sensitive (GSS) creep, respectively, and $A'$ is the temperature independent flow parameter. $A'$ is equal to $4 \times 10^{-19} \text{ Pa}^{-4} \text{s}^{-1}$ [Goldsby and Kohlstedt, 2001] and $5 \times 10^{-15} \text{ Pa}^{-2} \text{s}^{-1}$ (fitted to Goldsby and Kohlstedt [2001], see Figure 3.1) for dislocation and GSS creep, respectively. $Q$ is the activation energy, $T$ is the ice temperature, $\phi$ is the dust volume fraction, and $b = 2.9$ is a constant [Goughnour and Andersland, 1968; Durham et al., 1992; Mangold et al., 2002] at $\phi < 55\%$ where the rheology of the dust-ice mixture is not dominated by interactions between dust particles [Durham et al., 1992]. The values or range of values for these parameters are listed in
3.3.3 Constraining the ice grain size

Ice deformation via GSS creep is grain size dependent, and we must therefore attempt to constrain the ice grain size in order to determine the ice rheology. Deformation via dislocation creep is grain size independent, but only dominates ice deformation under high differential stress (> 1 MPa) at temperatures relevant to Mars. To better constrain the ice rheology, we invoke two processes that limit the ice grain size: deformation-driven dynamic recrystallization and grain boundary pinning by incorporated dust grains. Dynamic recrystallization provides an upper limit on the ice grain size within LDAs, whereas grain boundary pinning may better represent ice grain sizes in dusty ice deposits [Durand et al., 2006].

3.3.4 Dynamic recrystallization and GSS creep

A deforming, extremely clean ice deposit reaches a steady-state grain size when the rate of ice grain growth from grain boundary diffusion is equal to the grain-disruption rate resulting from dislocation creep deformation [DeBresser et al., 1998; Barr and Milkovich, 2008]. Therefore, the equilibrium ice grain size for dynamic recrystallization is the grain size at which the strain rate from grain boundary diffusion creep is equal to the strain rate from dislocation creep. This dynamic recrystallization grain size is inversely proportional to the applied differential stress, resulting in a range of ice grain sizes that vary with the differential stress within an ice deposit. In the case where the ice is extremely clean, the total strain rate $\dot{\varepsilon} = \dot{\varepsilon}_{\text{disl}} + \dot{\varepsilon}_{\text{diff}} = 2\dot{\varepsilon}_{\text{disl}}$ and the grain size will range from a value of 50 mm at $\tau = 10$ kPa to 1.3 mm at $\tau = 100$ kPa.

The various ice rheologies used in our model are illustrated in Figure 3.1 where the
differential stress ($\tau$) is plotted against effective viscosity ($\eta = \tau/\dot{\varepsilon}$) using equations 3.2-3.3. The thick solid, dashed, and dotted lines give the ice rheology for the indicated ice temperature and ice grain size based on the composite flow law of Goldsby and Kohlstedt [2001]. These experiments characterize ice deformation via a composite flow law consisting of contributions from diffusional flow, GSS creep, and dislocation creep. The slope of these lines changes at high stress due to a change in the dominant deformation mechanism from GSS creep to dislocation creep. The unlabeled thin lines represent our model approximation to these rheologies at stresses relevant to $\sim 500 \text{ m}$ thick ice sheets (shaded region) of between 10 -100 kPa using equation 3.2 with $n = 2$. Except for the 0.25 mm, 205 K case, these thin lines give a good approximation to the composite flow law of Goldsby and Kohlstedt [2001] because GSS creep is the dominant creep mechanism at stresses relevant to LDAs on Mars. The dislocation creep rheology ($n = 4$) sustained by dynamic recrystallization at low ($< 100 \text{ kPa}$) stress requires extremely clean ice, and follows the line labelled “dynamic recrystallization.” Note that the slope of these lines in log-space is given by $1 - n$ because we are plotting $\eta$ instead of $\dot{\varepsilon}$ versus stress.

Within the GSS creep deformation regime, we use three different rheologies based on ice grain sizes of 5, 1, and 0.25 mm (Figure 3.1). A discussion of the grain boundary pinning effect of incorporated dust grains and the concentration/sizes of dust grains necessary to produce ice grains of this size will be addressed in Section 3.7.1.

### 3.4 Model

We developed a glacial flow model simulating changes in ice thickness, $h$, as ice flows outward over a flat or sloping surface. Our approach is based on Nimmo and Stevenson [2001] who describe the viscous flow of a non-Newtonian fluid driven by pressure gradients associ-
Figure 3.1: Ice viscosity $\eta = \tau / \dot{\varepsilon}$ versus differential stress $\tau$ using the composite flow law from Goldsby and Kohlstedt [2001] (thick curving dotted, dashed, and solid lines) and our model approximation (thin dotted, dashed, and solid lines) for the indicated temperatures and grain sizes. In our pure ice simulations we use a rheology indicated by the line labelled “dynamic recrystallization” [Barr and Milkovich, 2008] in which grain size ranges from 50 mm at $\tau = 10$ kPa to 1.3 mm at $\tau = 100$ kPa. The shaded region corresponds to the range of differential stress experienced at the base of an ice sheet 500 m thick in our simulations.
ated with thickness variations. We modify their work by making some simplifying assumptions described below. The derivation that follows is similar to previous glacier models by, for example, Colaprete and Jakosky [1998], Mahaney et al. [2007], Pathare et al. [2005], and Patterson [1994].

Combining equations 3.1 and 3.2 and integrating in the $z$ direction gives the depth-dependent horizontal velocity of ice as it flows outward:

$$v(z) = 2A \left( g \rho \left( \frac{\partial h}{\partial x} + \sin \theta \right) \right)^n \int_0^h (h - z)^n \, dz$$

(3.4)

where the constant of integration is determined assuming that there is no basal slip ($v = 0$ at $z = 0$). To ensure mass conservation, we utilize the 2D, Cartesian continuity equation:

$$\frac{\partial h}{\partial t} = -\frac{\partial}{\partial x} \int_0^h v \, dz$$

(3.5)

Finally, combining equations 3.2, 3.3, 3.4, and 3.5 gives the rate of change in glacier thickness for both GSS creep ($n = 2$) and dynamic recrystallization-sustained dislocation creep ($n = 4$) deformation regimes:

$$\frac{\partial h}{\partial t}_{n=2} = \frac{1}{2} A'_{gss} d^{-p} \rho^2 g^2 \frac{\partial}{\partial x} \left[ h^4 \left( \frac{\partial h}{\partial x} + \sin \theta \right) \right] e^{-\frac{Q}{RT} - b \phi}$$

(3.6a)

$$\frac{\partial h}{\partial t}_{n=4} = \frac{2}{3} A'_{disl} \rho^4 g^4 \frac{\partial}{\partial x} \left[ h^6 \left( \frac{\partial h}{\partial x} + \sin \theta \right) \right] e^{-\frac{Q}{RT} - b \phi}$$

(3.6b)

In our model we assume that $T$ and $\phi$ (and, therefore $\rho$) are spatially homogeneous and are taken out of the derivative with respect to $x$ whereas $h$ and the local slope are kept inside. The simplifying assumption of a laterally and vertically constant temperature is appropriate because we are most concerned with the basal portion of the LDA deposit where most of the deformation takes place, and a thermal conduction temperature profile will result in little
Table 3.1: Definitions and measured or theoretical values (or range of values) for parameters used in the numerical simulations

<table>
<thead>
<tr>
<th>Description</th>
<th>Symbol</th>
<th>Disl. Value(s)</th>
<th>GSS creep Value(s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>activation energy(^{a})</td>
<td>(Q)</td>
<td>60 kJ mol(^{-1})</td>
<td>49 kJ mol(^{-1})</td>
</tr>
<tr>
<td>Flow Coefficient(^{b})</td>
<td>(A')</td>
<td>(5 \times 10^{-15}) Pa(^{-2}) s(^{-1})</td>
<td>(4 \times 10^{-19}) Pa(^{-4}) s(^{-1})</td>
</tr>
<tr>
<td>stress exponent(^{b})</td>
<td>(n)</td>
<td>4</td>
<td>2</td>
</tr>
<tr>
<td>grain size</td>
<td>(d)</td>
<td>1.3-50 mm</td>
<td>0.25-5 mm</td>
</tr>
<tr>
<td>grain size exp.(^{b})</td>
<td>(p)</td>
<td>0</td>
<td>1.4</td>
</tr>
<tr>
<td>dust fraction</td>
<td>(\phi)</td>
<td>(\ll 0.03%)</td>
<td>0.03%–4%(^{f})</td>
</tr>
<tr>
<td>dust frac. coeff(^{d}).</td>
<td>(b)</td>
<td></td>
<td>2.9</td>
</tr>
<tr>
<td>density of creep layer</td>
<td>(\rho)</td>
<td>0.90 g cm(^{-3})</td>
<td>0.90-0.94 g cm(^{-3})</td>
</tr>
<tr>
<td>density of solid particles(^{e})</td>
<td>(\rho_s)</td>
<td>3.0 g cm(^{-3})</td>
<td></td>
</tr>
<tr>
<td>dust particle size(^{e})</td>
<td>(r_d)</td>
<td>100 (\mu)m (1 (\mu)m for (d = 0.25) mm)</td>
<td></td>
</tr>
<tr>
<td>slope</td>
<td>(\theta)</td>
<td>0, 1(^{\circ})</td>
<td></td>
</tr>
<tr>
<td>gravity</td>
<td>(g)</td>
<td>3.7 m s(^{-2})</td>
<td></td>
</tr>
</tbody>
</table>

\(^{a}\) Goldsby and Kohlstedt [1997]  
\(^{b}\) Goldsby and Kohlstedt [2001]  
\(^{c}\) Plaut et al. [2009]  
\(^{d}\) e.g. Mangold et al. [2002]  
\(^{e}\) Barr and Milkovich [2008]  
\(^{f}\) this study; assuming \(r_d = 1 \mu\)m for \(d = 0.25\) mm and \(r_d = 100 \mu\)m for other grain sizes

Temperature variation over this basal layer. For the purposes of this work, our model assumes \(T\) is also constant over time in order to constrain the average temperature over the lifetime of LDAs (∼ 300 Ma) that results in LDA profile dimensions consistent with the observations. Table 3.1 lists the values (or range of values) we use in our simulations for the parameters above.

### 3.5 Numerical scheme

Our one-dimensional model simulates changes in ice thickness \((h)\) of an assumed initial ice deposit using equations 3.6a and 3.6b as ice flows outward over a flat or sloping surface assuming no ice accumulation or loss. The model domain consists of 350 discrete
elements, each 100 m in length. The thick portion of the initial ice deposit starts at the right-
most end of the profile, and the horizontal velocity is set to zero at both the right and left
boundaries to prevent flow in or out of the domain. As previously mentioned, the horizontal
velocity at the base of the ice is also set to zero (no basal sliding). We incorporate a variable
time-step calculated using the Courant criterion to insure that fast flow (thick ice) is associated
with a short time step, but the time step is allowed to increase as the ice deposit thins and
flow velocities decrease. The shape of the initial profile is given by a monomial of the form
\[ h = ax'^{1/4} \]
where \( x' \) is the distance from the toe of the initial ice sheet, and \( a = \frac{1000}{l^{1/4}} \) where
\( l \) is the length of the initial LDA in meters. This initial shape is more realistic for glacial flow
than some other arbitrary geometry (i.e. rectangle or triangle), and it reduces computation
time by minimizing the high flow rates that can result from the relaxation of large gradients or
thicknesses. Because the shape evolves towards a similarity solution, initial profile shapes will
become more similar over time as flow takes place [Huppert, 1982], so our initial profile choice
will not influence our results (see Section 3.6).

Our simulations are based on five assumed initial ice deposits with different total ice
volumes, although the initial maximum thickness is set equal to 1 km for all cases based on the
estimated ice sheet thickness during regional glaciation in the Amazonian [Dickson et al., 2008,
2010]. The horizontal extent of the initial ice sheet was set equal to 1, 2.5, 5, 7.5, or 10 km
for the five cases. These initial lengths correspond to a total cross sectional area of: 0.83, 2.0,
4.0, 6.0, and 8.0 \( \times 10^6 \) m\(^2\) when these deposits overlie a flat surface. For simulations over a
1° sloping surface, rather than change the profile shape of the initial deposits, we decrease the
total volume of ice by removing the 1° wedge occupied by the sloping ground surface from
the bottom portion of the initial deposit resulting in total areas of: 0.82, 1.95, 3.8, 5.5, and 7.1
\( \times 10^6 \) m\(^2\) for the five initial lengths given above.

For each of the five initial ice thickness profiles, we ran 24 simulations. The ice grain size was assigned values of 0.25, 1, or 5 mm for the GSS creep simulations, whereas the grain size in the dislocation creep simulations took on a range of values between < 1.3 to 50 mm determined by dynamic recrystallization. Temperatures of 195, 205, and 215 K were used based on the current, 205 K, mean annual surface temperature at 40° latitude [Mellon et al., 2004], and the modeled variations from 205 and 195 K over the past 5 My [Schorghofer, 2008]. The 215 K temperature case assumes a sustained period of elevated temperatures prior to the 5 Ma period modeled by Schorghofer [2008]. The final variable was the basal slope, which was either 0° or 1°. By running simulations with different combinations of ice temperatures, dust volume fractions and slope we can determine the values for these parameters that produce LDAs that most closely match the observed length, thickness and age. Due to the non-linear dependence of \( \eta \) on \( T \), the warmest periods are likely to have controlled the flow timescale of LDAs over the past \( \sim 100 \) My. Therefore, the temperature used in the simulations that best reproduce the observed LDA length, thickness, and age will reflect the warmest periods in recent Martian history.

### 3.6 Results

Figure 3.2 illustrates the effect of different model parameters on the ice flow rate in the GSS creep regime. Figure 3.2a shows a 1 Gy simulation of ice flow for an initial ice deposit 5 km long using an ice temperature of 205 K and \( d = 5 \) mm. The initial profile is shown by the thick, solid line and the thin, solid lines indicate the topographic profiles at 200 My time intervals. A MOLA profile of an LDA at 38.6°N, 24.3°E (dashed line) is shown for comparison. ShaRAD
Figure 3.2: a) An example of a 1 Gy simulation of ice flow for an initial ice deposit length of 5 km (thick line) using the indicated ice temperature, basal slope, and grain size for a GSS creep rheology. The thin, solid lines indicate the topographic profiles at 200 My time intervals. Upper and lower dashed lines are MOLA profiles for LDA #15 and 24.5 at 42.0°N, 18.4°E and 38.6°N, 24.2°E, respectively. b) is the same as a) plotting only the final profiles for simulations using the indicated temperatures. c) same as b) for $d = 1$ mm over a 100 My period. d) same as a) with a basal slope of 1° (here the upper MOLA profile has been shifted to the right to align the LDA front with the model).
data at this location suggests that this particular LDA lies on a flat surface [Plaut et al., 2009]. The upper dashed line represents MOLA data from 42.0°N, 18.4°E - a location where ShaRAD data indicates the presence of a basal slope below the LDA of about 0.5° (see Appendix B).

In our simulations, flow is initially fast due to the large ice thickness, but significantly slows as the ice sheet thins, due to the increase in viscosity at lower differential stress (see Figure 3.2). Panel b of Figure 3.2 is identical to part a, except we have only plotted the end-state of the simulation, and have included the final profiles for simulations with ice temperatures of 195 and 215 K to illustrate the effect of temperature on the ice run-out distance. As expected, lower temperatures result in higher viscosities and a reduction in the flow rate whereas higher temperatures result in an accelerated flow rate. The effect of decreasing the grain size to 1 mm is shown in part c. Note that the simulation time is now 100 My - indicating that flow occurs ∼10 times as fast as the 5 mm grain size simulation shown in part a (see equation 3.2). Lastly, model profiles from a set of 1 Gy simulations using $d = 5$ mm and a slope of 1° is shown in part d. The ice deposit has flowed a greater horizontal distance, and has a greater apparent thickness (due to the presence of the basal slope) than the simulations over a flat surface, although the thickness of the ice deposit itself is reduced in this simulation. The presence of a basal slope in this simulation produces model profile shapes more similar to the shape of the upper-most MOLA profile than simulations over a flat surface. Note that the upper-most MOLA profile has been shifted to the right in order to produce a better fit in Figure 3.2d, but we could have produced a good fit without shifting the MOLA profile if a simulation with a greater initial ice volume were used. In contrast, even if a larger initial ice volume were used, the constant-slope proximal surface seen in the upper-most MOLA profile would not be reproduced in a simulation with a flat basal slope.
We tested whether the shape of our assumed initial profile had any influence on the results by running simulations with same ice volume, but with different initial profile shapes. We tested a rectangular initial profile with the same initial length as our monomial profile, but with a thickness that gave the appropriate ice volume (800 m for the 10 km long initial profile). We also tested a triangular initial profile with an initial length 50% longer than the monomial case, and a height that gave the appropriate ice volume (1070 m high and 15 km long for the 10 km long monomial profile). Simulations of the shape of the ice deposit using the rectangular and triangular initial profiles matched that of the monomial case after 1 My and 5 My, respectively, for flow at 205 K and $d = 1$ mm. Compared to the timescales relevant for this study ($\sim 100$ My), these short response times indicate that our results will not be influenced by our assumption of the initial profile shape (as expected from the similarity solution argument).

LDA thickness and length observations from Li et al. [2005] are shown in Figure 3.3 by the circles and stars. The filled circles represent LDAs with the most convex topographic profiles (type I in Li et al. [2005]). The lower-most MOLA profile shown in Figure 3.2 (indicated by the hexagram in Figure 3.3) has a slightly convex, topographic profile and falls under the type II classification as defined by Li et al. [2005]. The upper-most MOLA profile shown in Figure 3.2 falls under the same, type II, classification and corresponds to the right-most star in Figure 3.3. The open circles in Figure 3.3a-d are LDAs of type II or III (even more linear in shape). These less convex LDAs were interpreted by Li et al. [2005] as evidence for the loss of interstitial ice to sublimation, although the existence of a sloping substrate may also explain the shape of some LDAs on Mars (see Figure 3.2d and Figure 3.3). The starred points are LDAs with ShaRAD data to constrain the subsurface slope (see Appendix B), with the hexagram being
the shorter LDA shown in Figure 3.2.

The 36 thickness and length measurements from LDAs in the Li et al. [2005] study used MOLA tracks that were not always parallel to the flow direction, or were locations where LDA flow is either convergent or divergent. Although these measurements were sufficient for their study, the locations of some of these observations may not be consistent with the assumed Cartesian flow geometry used in our simulations. Nonetheless, we make use the data set from Li et al. [2005] to make comparisons with our model because the thickness measurements are accurate and the apparent LDA length measurements made along MOLA tracks in Li et al. [2005] are < 30% longer than the true length.

Lines showing how the ice sheet thickness and length evolves with time for the five different initial ice deposits are overlain on the observations in Figure 3.3a-d. Four of the 24 sets of simulations are shown in Figure 3.3a-d, illustrating the variation in the ice flow timescale with regard to ice grain size, temperature, and slope. Our reference simulation using $T = 205$ K, $d = 5$ mm on a flat slope is shown in Figure 3.3a. Results of simulations using clean ice are not shown due to the very high grain size and viscosity (Figure 3.1) resulting in an unrealistically long timescale ($> 4$ Gy) for LDA formation. All the simulations shown in Figure 3.3 have a large initial ice flow velocity resulting in a rapid decrease in ice thickness as indicated by the steeply sloping solid lines at the top of each plot. Eventually the velocity slows as the ice deposits thin and lengthen resulting in the slightly curved trajectories given by the solid lines. Time contours are plotted as dashed lines. In Figure 3.3a-c, these contours roughly trend with the length and thickness observations made by Li et al. [2005] indicating that these LDAs share a similar time of formation.

In the simulations where flow occurs over a surface with a 1° slope (Figure 3.3d),
the vertical drop along the length of the LDA decreases less rapidly over time, and actually begins to increase again for larger initial ice deposits. Although the ice deposit is still thinning as it flows downslope, the change in elevation acquired from progressing further down the sloping surface compensates (and begins to overcome) the thinning effect for the larger initial ice deposits. This effect results in a more tightly curving trajectory shown by the solid lines in Figure 3.3d. The time contours in Figure 3.3d give predicted ages for the thickest LDAs that are more consistent with the 90 to \( \sim 300 \) My crater age-dating estimates by Baker et al. [2010]; Mangold [2003]. If we assume these crater age-dates are accurate, then our model predicts the presence of a basal slope below the thickest LDAs in the survey by [Li et al., 2005] - a prediction confirmed by ShaRAD for the two starred points in Figure 3.3 based on published radargrams (see Appendix B).

Using plots like Figure 3.3a-d, we can determine the model-predicted age of the aprons assuming a particular ice temperature, grain size, and slope. Plots of these predicted ages for 37 LDAs using the dynamic recrystallization model, and the GSS creep model for \( d = 5, 1, \) and \( d = 0.25 \) mm for ice at different temperatures and basal slopes are shown in Figure 3.4a, b, c, and d, respectively. These northern, mid-latitude LDAs are divided into regional groups by the solid vertical lines (covering the longitude range indicated at the top of the figure) and the LDA # corresponds to the designation given by Li et al. [2005]. We have included our additional measurement at LDA number 24.5 based on the location of adjacent measurements in Li et al. [2005]. The lower-most MOLA profile shown in Figure 3.3 corresponds to this additional measurement. The LDAs labeled with an arrow (# 15, 16, and 24.5) are locations where the ShaRAD radar instrument has constrained the basal slope as indicated in the parenthesis (see Section 3.7). The color-coded symbols located at LDA #0 give the median ages for each
Figure 3.3: a) Time-varying ice sheet thicknesses (vertical drop) and lengths for five different initial ice deposits (solid lines) using the indicated model parameters overlain on LDA observations from Li et al. [2005] of convex (filled circles) and more linear (open circles) topographic profiles. Time contours are plotted on top of the simulations (dashed lines). The starred points indicate locations where ShaRAD data has constrained the basal slope with the hexagram representing a measurement made in this study. b) is same as a) using $d = 0.25$ mm. c) is same as a) using $T = 215$ K, and d) is same as a) except that flow occurs over a $1^\circ$ sloping surface.
set of simulations with the ‘x’ indicating the median age for the 205 K, 1° simulations.

The predicted age for a given LDA changes by a factor of about 4 for the clean ice simulations on a flat surface when the temperature changes from 215 to 205 K, and again from 205 to 195 K. Also, the model-predicted ages decrease by roughly an order of magnitude when the ice grain size changes from the dynamic recrystallization simulations to simulations with $d = 5$ mm, from 5 mm to 1 mm, and again from 1 mm to 0.25 mm. This hastening of ice flow is expected from equation 3.2.

Of the different rheologies we implement, GSS creep with a grain size of either 5 mm or 1 mm (Figure 3.4b, c) produce LDAs with ages most consistent with the crater density measurements, whereas dislocation creep and GSS creep with $d = 0.25$ mm rheologies tend to over- and under-predict, respectively, the crater age dates (we discuss this more below). For a given ice temperature and slope, the model-predicted age varies by 4 orders of magnitude for the clean ice (dynamic recrystallization) simulations, and by 2 orders of magnitude for the GSS creep simulations for the different LDAs included in this study. This large discrepancy in age between different LDAs is not in agreement with crater age date estimates that suggest LDAs are between 90 and few hundred My old [Baker et al., 2010; Mangold, 2003] (shaded region in Figure 3.4). Therefore, we suggest that no single basal slope-ice temperature combination can give an appropriate age for all LDAs, and that regional climate, local topography, and/or LDA rheology (e.g. ice grain size) must vary between different LDAs in order to give ages that are more in agreement with one another. Based on the results shown in Figure 3.3 and 3, the presence of a basal slope may provide the simplest explanation for both the age and shape discrepancies among LDAs. Alternatively, different LDAs could be of different ages - we discuss these issues further below.
Figure 3.4: a) Model-derived ages of individual LDAs (numbering scheme is from Li et al. [2005]) using the indicated ice temperature and assuming the LDA rheology is governed by dynamic recrystallization of ice grains. The color-coded points at LDA #0 give the median ages for each set of simulation parameters (‘x’ corresponds to the 1° slope simulations). LDA observations by Li et al. [2005] were made between 35°N and 50°N latitude and over the range in longitudes indicated at the top of the figure. The shaded region gives the range in LDA age determined from crater counts [Baker et al., 2010; Mangold, 2003]. The arrows highlight locations where the basal slope has been determined by ShaRAD observations [Plaut et al., 2009]. Panels b), c), and d) are the same as a) for simulations using the indicated values for d using a GSS creep rheology for ice.
3.7 Discussion

As mentioned previously, our numerical model of ice flow relies on two major assumptions: 1) the ice grain size and temperature in the basal portion of the ice sheet (where the majority of deformation occurs) is spatially and temporally homogeneous, and 2) the size of the ice grains within LDAs is controlled by either dynamic recrystallization for clean ice or by grain boundary pinning due to the presence of dust. We assume a constant basal ice temperature because, although obliquity variations are likely to drive temperature changes at the equatorial and polar regions, obliquity-induced, mid-latitude temperature variations are subtle because the annually averaged insolation at mid-latitudes is nearly independent of obliquity [Schorghofer, 2008; Mellon and Jakosky, 1995]

In equation 3.2, the two assumptions mentioned above give values of $n = 2$ and $n = 4$ for GSS creep in the presence of dust and dislocation creep of clean ice, respectively. However, the value of $n$ for ice in mid-latitude deposits on Mars is a matter of some debate. Comparison of theoretical topographic profiles of ice flow with MOLA topographic profiles of LDAs in the Deuteronilus Mensae region by Mangold and Allemand [2001]; Li et al. [2005] suggest that LDAs deform according to simple plastic rheology ($n = 1$ with no basal sliding and a yield stress of 0.6 to 1.3 bar [Squyres, 1978]). These topographic observations would suggest buried ice within LDAs must have a grain size of 10 $\mu$m or less - otherwise they would contradict the power law creep behavior of ice predicted from experiments by, e.g., Russell-Head and Budd [1979]; Durham et al. [1992]; Goldsby and Kohlstedt [2001]. However, even if the ice consisted of 10 $\mu$m grains, if the dust volume concentration is $< 30\%$ as suggested by Plaut et al. [2009], then the viscosity would be reduced to $\sim 10^{14}$ Pa s (based on Goldsby and Kohlstedt [2001]) and would flow very rapidly over 100 kyr timescales. In our view, it
is likely that the $\sim 20^\circ$ ice flow fronts seen in our simulations would be modified by mass wasting and/or sublimation during or subsequent to ice flow to produce shapes more similar to the observed LDA profiles. Therefore, instead of relying on the (potentially modified) shape of LDAs to constrain the rheology of these deposits, we assume an ice rheology more consistent with experimental results.

In addition, the observed lack of deformed craters on LDA surfaces [Carr, 2001] is consistent with our model of ice flow. Because most of the deformation is concentrated in the beginning of the simulation when the ice thickness is large (equations 3.6a, 3.6b), there is only a short window of time for a crater to become deformed following the emplacement of the LDA deposit. As the ice spreads and thins, ice flow stalls and deformation of the surface slows dramatically. For instance, over the course of the 1 Gy simulation depicted in Figure 3.2a, the surface of the thick portion of the ice deposit experiences a horizontal extensional strain of 2.4, but a strain of 1.6 had already been reached after the first 200 My of flow. Crater modification by surface strain, viscous crater relaxation and sublimation pitting have likely erased the oldest craters on LDAs. Large, deformed craters would be easier to find, however, on an LDA with a simple plastic rheology because the decrease in flow rate would be more gradual. This more slowly evolving LDA would result in a longer period of surface deformation - making deformed craters easier to find.

For a given combination of basal slope and ice temperature, our simulations suggest that mid-latitude LDAs in the northern hemisphere of Mars potentially span a large range in age - from 10 My to more than 1 Gy for ice with $d = 5$ mm at 205 K flowing over a flat surface (Figure 3.4b). Focusing on the median ages, $d = 5$ mm grain size, 205 K ice flowing on a flat surface gives the most reasonable results. However, rather than assume that a particular basal
slope and ice temperature is appropriate for all LDAs, we should consider the influence of local
topography and regional climate in determining the age of a given ice deposit. Some evidence
for variations in basal slope are evident in ShaRAD data corresponding to LDA numbers 16 and
24.5 [Plaut et al., 2009]. LDA # 24.5 has a flat basal surface if a dielectric constant appropriate
for clean ice (3.3) is used to migrate the subsurface signal. However, if the same dielectric
constant is applied to the reflection from LDA #16, the migrated signal has a regional slope of
about 0.5°. Note that the predicted ages for LDA #16 at 205 K on a 1° slope (dashed line) and
LDA #24.5 at 205 K on a flat surface (line marked with +) provide consistent ages of ~ 10^5 yrs,
but assuming the slope and temperatures are the same at both locations would give predicted
ages that differ by about a factor of five.

Alternatively, differences in LDA rheology, such as differing ice grain sizes, could
give LDA ages that are more consistent with one another. Incorporating differing amounts
of dust in LDA ice could result in different ice grain sizes because the pinning of ice grain
boundaries by dust depends on the dust concentration. We discuss this issue further below.

3.7.1 Grain Boundary Pinning

Under Martian polar conditions, polycrystalline ice grains will grow over time in or-
der to lower the free energy at the grain boundaries in a process known as sintering - reaching
at least 200 μm in the upper meter of the northern polar ice cap based on modeling by Kieff-
fer [1990]. However, if solid particulates are dispersed within the polycrystalline ice, further
migration of a grain boundary is hindered when it intersects the location of a dust grain [Alley
et al., 1986]. Although the presence of clathrates or bubbles may also impede ice grain growth
[Durand et al., 2006], we assume that dust grains are primarily responsible for pinning the grain
boundaries of polycrystalline ice. This grain boundary pinning effect of incorporated dust limits the ice grain size \(d\) to a value of

\[
d = 40 \mu m \left( \frac{r_d}{1 \mu m} \right) \left( \frac{1\%}{\phi} \right)^{\frac{1}{2}}
\]  

(3.7)

where \(r_d\) is the dust particle radius and \(\phi\) is the dust volume fraction (Poirier [1985] and modified from Barr and Milkovich [2008]). The grain sizes predicted by equation 3.7 are very similar to ice grain sizes in the dustiest portions of ice cores from Greenland and Antarctica [Thorsten-son et al., 1997; Delmonte et al., 2004], but equation 3.7 is only true for a specified (unimodal) dust grain size. In these cores, an observed dust grain radius of between \(r_d = 0.5 \mu m\) [Durand et al., 2006] and \(r_d = 1.5 \mu m\) [Ruth et al., 2003] at concentrations of \(\sim 5\) mg kg\(^{-1}\) by mass (0.0002\% by volume) [Durand et al., 2006; Ruth et al., 2003; Steffensen, 1997] are associated with ice grains 3 mm in diameter [Durand et al., 2006]. Equation 3.7 predicts ice grain sizes of 1.4 and 4.2 mm for \(r_d = 0.5, 1.5 \mu m\), respectively. Note that the ice grain size limited by grain boundary pinning is proportional to the size of the incorporated dust grains (equation 3.7) because, for a given volume fraction of dust, the concentration of particles decreases with increasing particle size.

In order to determine the appropriate dust grain size for Mars we rely on infrared spectroscopy and \textit{in situ} measurements from lander/rover missions. LDAs in the northern hemisphere of Mars lie within a region of high thermal inertia (90-320 J m\(^{-2}\) K\(^{-1}\) s\(^{-1/2}\)) and moderate albedo (0.2-0.26) suggesting the surface is comprised of a duricrust with some sand, rocks and bedrock [Mellon et al., 2008]. Although LDAs are found in high thermal inertia regions, the surfaces themselves have an anomalously low thermal inertia compared to their surroundings. A thermophysical unit similar to LDA surfaces was studied by the Phoenix Lander where grain size measurements using microscopic imaging of soils found the dust grain size distribution has
two peaks: one at $\sim 100 \, \mu m$, and another at $< 10 \, \mu m$ [Goetz et al., 2010]. Thermal emission spectroscopy gives low thermal inertia measurements in continental-scale (e.g. Tharsis Montes) dust deposits measured by Mars Global Surveyor as well as in small-scale aeolian deposits as seen by the Mars Exploration Rover Mission (MER). These measurements indicate deposition of fine-grained aeolian material with particle sizes smaller than $40 \, \mu m$ [Herkenhoff et al., 2004a; Golombek et al., 2008].

So far we have only mentioned the ice grain size when referring to our simulations, since this is most directly related to viscosity. In order to convert ice grain size to dust volume fraction requires the dust grain size to be specified. Atmospheric observations by Mars lander missions [Pollack et al., 1995; Tomasko et al., 1999; Lemmon et al., 2004] suggest dust particles entrained in the Martian atmosphere have a radius of 1 $\mu m$, although sand-sized particles ($\sim 200 \, \mu m$) deposited by winds have been observed on the decks of the MER rovers [Goetz et al., 2010]. We assume a large dust particle diameter of 100 $\mu m$ in order to place an upper bound on the dust volume fraction (equation 3.7). Using this dust particle size and ice grain sizes of 5 and 1 mm gives dust volume fractions of 0.16% and 4%, respectively, using equation 3.7. Because a dust particle radius of 50 $\mu m$ gives a dust fraction beyond the limits placed by ShaRAD for the 0.25 mm ice grain size case, we assume a dust grain radius of 1 $\mu m$ (giving a dust fraction of 0.03%) in the $d = 0.25 \, mm$ simulations. These dust fraction constraints are similar to the $\phi \leq 5\%$ constraints placed on the north polar layered deposits by ShaRAD [Grima et al., 2009] and the $\phi \sim 15\%$ estimate for the south polar layered deposits based on gravity measurements [Zuber et al., 2007].

Aside from the grain boundary pinning effect, dust fractions of less than 5% have little influence on the flow timescale either through viscosity enhancement from particle-particle
interaction in the ice-dust mixture (equation 3.3) or through flow enhancement due to the increased density of the ice-dust mixture. At dust fractions between roughly 20 and 55%, however, prior simulations by Pathare et al. [2005] and our own simulations show that the increased density of the ice-dust mixture compared to clean ice dominates over the viscosity effect resulting in increased flow rates.

### 3.7.2 Past Climate Implications

Based on our simulation results, an LDA rheology associated with an ice grain size of 5 mm best reproduces the LDA length and thickness measurements assuming an ice temperature of 205 K and an LDA age of $\sim 100$ My. If grain boundary pinning is responsible for limiting the ice grain size to 5 mm, a dust fraction of $\leq 0.16\%$ would be required assuming a dust grain diameter of $\leq 100 \mu m$ (equation 3.7). Assuming these ice deposits were formed from the precipitation of ice, then we can constrain the precipitation rate if we know the dust deposition rate.

Using a dust deposition rate of $20 - 45 \mu m$ per Earth year based on Pathfinder data [Johnson et al., 2003] as a proxy for past dust deposition rates, we can constrain the precipitation rate required to generate an ice sheet with a given dust concentration by simply dividing the dust deposition rate by $\phi$. Using $\phi = 0.16\%$ as a maximum dust fraction in this calculation gives a lower bound of $1.2$ cm yr$^{-1}$ for the precipitation rate.

Climate models have suggested that increased wind speeds would have made the Martian atmosphere dustier at high obliquity [Haberle et al., 2003; Newman et al., 2005], and would have enhanced meridional transport of air masses carrying water vapor from sublimating tropical mountain glaciers during the winter season resulting in northern mid-latitude precipitation
[Madeleine et al., 2009]. In this climate scenario, the annual accumulation rate of precipitation in Deuteronilus Mensae is $\sim 1 \text{ cm yr}^{-1}$. This precipitation rate is consistent with the above estimate based on $\phi = 0.16\%$ and $r_d = 50 \mu\text{m}$. A smaller precipitation rate would result if we assumed a slightly large dust particle radius, or a slightly smaller ice grain size.

At a precipitation rate of $1 \text{ cm yr}^{-1}$, it would take between $3 \times 10^4$ to $10^5$ yr to accumulate a deposit of comparable thickness to LDAs assuming no ablation, short compared to their actual age. However, given the presence of surface textures indicative of ice sublimation, this formation timescale estimate is likely a lower bound.

### 3.8 Conclusions

Using laboratory experiments, observations from terrestrial ice sheets, radar and topography data of Martian LDAs, and theory regarding the rheology and grain interactions of ice-dust mixtures, we have attempted to better constrain the rheology of LDAs. Our simulations assume no precipitation or ablation of ice and that the basal portion of LDAs have a spatially and temporally homogeneous dust fraction and temperature. By varying the temperature, ice grain size, and initial ice volume used in these simulations we find that LDA length and thickness measurements are best reproduced over a $\sim 100$ My period by an ice rheology corresponding to an ice grain size $d = 5 \text{ mm}$ and an ice temperature $T = 205 \text{ K}$ (Figure 3.4). If incorporated dust grains with a diameter of $100 \mu\text{m}$ limit the ice grain size due to grain boundary pinning, then an ice grain size of $5 \text{ mm}$ corresponds to a dust volume fraction of $\phi = 0.16\%$ (equation 3.7), consistent with ShaRAD observations.

However, the rheology of $d = 5 \text{ mm}$, $T = 205 \text{ K}$ ice does not provide reasonable age estimates for all the LDAs in this study, and we must invoke regional heterogeneity in some
or all of the variables in our model to explain this variability. One likely source of regional variability is the basal slope. Simulations of ice deposits flowing down a sloping surface provide better fits to the observed topography over $\sim 100$ My timescales (see Figure 3.2d and 3.3d) for certain LDAs than simulations over a flat, horizontal surface. In order to give LDA ages that are consistent with crater age estimates, and in order to generate topographic profiles similar to the observations, our simulations require the presence of a basal slope $\sim 1^\circ$ below about one fourth of the 37 LDAs analyzed in this study (Figure 3.4b). Basal slopes of $0.5^\circ$ below LDA numbers 15 and 16, and $0^\circ$ below LDA 24.5 measured using radar (Appendix B) correspond to the slopes we would expect at those locations by comparing an assumed LDA age of $\sim 100$ Ma (based on crater counts) with the ages predicted from our model for different ice temperatures and basal slope (Figure 3.4).

If, however, differences in LDA rheology rather than the presence of a basal slope are responsible for the variability in the model-predicted ages, then our results suggest that the majority of LDA deposits are composed of coarse grained ice ($d > 1$ mm) assuming LDAs deform via GSS creep (equation 3.7 and assuming an LDA age of $\sim 100$ My and a temperature of 205 K). The dust fraction required to limit ice grain grown to this size depends on the dust particle size. Assuming $d > 1$ mm and a dust particle radius less than 50 $\mu$m, limits the dust fraction in LDAs to less than 4% (equation 3.7), again consistent with ShaRAD's constraints.

Another, perhaps less likely, explanation is that temperature differences among different LDAs due to either insolation or geothermal variations result in LDAs with different predicted ages. We deem this hypothesis less likely because we do not find a trend in the model-predicted ages with either latitude or facing direction.

If precipitation of snow during high obliquity is responsible for the formation of LDA...
deposits at mid-latitudes, then the range of likely dust fractions (\(\leq 0.16\%\)) places a lower limit of 1.2 cm yr\(^{-1}\) on the precipitation rate assuming a dust particle radius of \(r_d = 50 \mu m\) (Section 3.7.2). Assuming a precipitation rate of 1 cm yr\(^{-1}\) a minimum period of \(3 \times 10^4\) to \(10^5\) yr would be required to form LDAs.
Chapter 4

Crater Softening

4.1 Background

The Martian subsurface is arguably the largest reservoir for H$_2$O on the planet. Based on thermophysical models, Clifford [1993] suggested that the Martian cryosphere could easily extend to 5 km depth globally. Given a surface porosity of 20%, a 5 km deep cryosphere amounts to a volume of water that could cover Mars in a 540 m deep global ocean. Constraining the thickness of the ice-rich regolith on Mars through geologic observations and remote sensing is therefore important in determining the global water budget.

Although direct evidence of water ice in the upper meter of the Martian subsurface has only recently been provided by neutron and gamma-ray spectrometer data and thermal emission observations [Boynton et al., 2002; Feldman et al., 2004b; Bandfield, 2007], its presence had been strongly suspected since the Viking missions on both theoretical and observational grounds. Remote sensing observations [Mustard et al., 2001] and numerical climate models [Fanale et al., 1986; Paige, 1992; Mellon et al., 1997] suggest that equatorial regions are generally depleted in near-surface ice, while an ice-rich mantle extends poleward of 30°.
To probe for ice at greater depths, we rely on remote sensing observations from radar reflectometry and on geological observations such as topographic roughness, crater softening, viscous flow features, and rampart ejecta. One of the few Mars radar results published to date is by Plaut et al. [2007] in which the thickness of the ice-rich south polar layered deposits are measured to be 3 km. Holt et al. [2008] use radar observations at lower latitude to measure ice-rich material 800 m in thickness around isolated massifs. Future radar observations will likely constrain the thickness and distribution of ice-rich regolith over broader scales as coverage increases (see Chapter 3).

Global surface roughness measurements were made by Kraslavsky and Head [2003]. Their work suggests that preferential sublimation of ice on equator-facing slopes has occurred at short lengthscales (~300 m) in a mid-latitude band at 45° in the recent past (<10 Ma). At longer lengthscales (~30 km) and longer timescales (~100 My), an intact ice-rich regolith can result in viscous relaxation of crater topography. Relaxation of a deformable, presumably ice-rich, regolith layer of approximately 1 km in thickness may explain softened craters at mid to high latitude according to Jankowski and Squyres [1992]. The presence of lobate debris aprons and viscous flow features provide additional evidence for a large reservoir of ice-rich material in the near-surface of the Martian regolith [Colaprete and Jakosky, 1998; Mangold et al., 2002; Milliken et al., 2001, see Chapter 3]. Lobate ejecta deposits surrounding craters on Mars provide evidence for water ice in the near subsurface [Strom et al., 1992]. Kuzmin et al. [1988, 1989] found a strong latitudinal dependence regarding the onset diameter of craters with rampart ejecta. They suggest that this latitudinal dependence reflects an enrichment of ice in the Martian regolith with increasing latitude. An earlier study by Mouginis-Mark [1979] found that craters with a pancake ejecta morphology occurred almost exclusively poleward of 40° in
both hemispheres.

Here we investigate crater relaxation as a probe of the thickness of the inferred sub-surface ice-rich layer. Building on work by Kreslavsky and Head [2003] and Jankowski and Squyres [1992] we explore crater slope asymmetry as a means of determining the thickness of ice-rich regolith on Mars. First, we develop a theoretical model for crater slope asymmetry generation via viscous relaxation of crater topography. Next, we describe the method by which we analyzed real craters using Mars orbiter laser altimeter (MOLA) data [Smith et al., 2001]. In section 4.4 we compare crater relaxation simulation results to the observations from various latitude regions on Mars in order to constrain the model parameters. Finally, we discuss the model and observational results regarding the thickness of ice-rich material in the Martian regolith and present our conclusions.

4.2 Theory

4.2.1 Downslope Creep of Ice-Rich Regolith

As described in Chapter 3, the downslope transport of ice-rich Martian regolith may be modeled as a diffusion creep process which smoothes out topographic variations over time [Colaprete and Jakosky, 1998; Pathare et al., 2005]. Downslope creep is driven by a shear stress

$$\tau = \rho gz \frac{\partial h}{\partial r}$$

(4.1)

where $\rho$ is the regolith bulk density, $g$ is the gravitational acceleration, $z$ is depth, and $\frac{\partial h}{\partial r}$ is the local slope. The bulk density is a function of the density of the solid particles (2500 kg m$^{-3}$ in our model), the density of ice (900 kg m$^{-3}$), and the dust fraction. The response of ice to applied stresses is typically nonlinear (non-Newtonian); here we assume that it deforms according to
Glen’s Flow law (which may not be appropriate as described in Chapter 3):

\[ \dot{\varepsilon} = \frac{\partial v}{\partial z} = A \tau^3 \]  

(4.2)

where \( \dot{\varepsilon} \) is strain rate, \( v \) is the down-dip velocity, and \( A \) is inversely related to viscosity [Colaprete and Jakosky, 1998]. Goldsby and Kohlstedt [2001] suggest a stress exponent of 1.8 for ice at Mars conditions, and include a grain size dependent strain rate. The stress exponent increases to 4 at a stress of 1 MPa.

Crater softening due to ice-driven creep may vary spatially within a single crater due to temperature variations induced by slope-related insolation variations. Because creep processes are more rapid at higher temperature, one would expect equator-facing crater slopes to creep more and become shallower than pole-facing slopes at low to moderate obliquities. The relaxation of topography (\( h \)) is governed by the thickness of the ice-rich layer, the viscosity at the surface of the creeping layer (\( A^{-1}_{\text{sf}} \)), and the local slope. In order to minimize computation time, we make the simplifying assumption that the porosity of the ice-rich Martian regolith decreases exponentially with depth. This porosity increase can be expressed in terms of \( A \) as follows:

\[ A(z) = A_{\text{sf}} \left( e^{-z/\delta} \right) \]  

(4.3)

where \( z \) is depth, increasing downward and the \( e \)-folding depth (\( \delta \)) of porosity will hereafter be referred to as the thickness of the creeping layer. Here we assume that a decrease in the ice volume fraction with depth influences the regolith rheology. We note, however, that the regolith viscosity actually decreases with depth [Mangold et al., 2002] because the strain rate increases more quickly with depth than \( A \). Our assumed porosity structure is a conservative assumption because it tends to slow the rate of crater relaxation relative to a uniform porosity structure. The
value of $A$ at the surface of the creeping layer is given by

$$A_{sfc} = A_o \left[ e^{\frac{Q(T - T_m)}{RT_m}} - c\phi \right]$$  \hspace{1cm} (4.4)

(modified from Pathare et al. [2005]) where $Q$ is the activation energy, $R$ is the ideal gas constant, $T$ is the local temperature, $T_m$ is 273 K, $A_o$ is related to the viscosity of polycrystalline ice at 273 K, $\phi$ is the dust fraction, and $c$ is a constant [Durham et al., 1992]. This dust fraction coefficient, $c$ is different than that used in Chapter 3 because $c$ is dust fraction-dependent. Here, the larger dust fractions associated with an ice-rich regolith result in a viscosity that increases more rapidly with dust fraction [Durham et al., 1992; Mangold et al., 2002] - giving a larger value of $c$ when compared to $b$ in Chapter 3. Assuming the viscosity structure given above, radial symmetry, the non-Newtonian rheology given in equation 4.2, and a constant ice-rich layer thickness, the following equation for the change in topography with time may be derived:

$$\frac{\partial h}{\partial t} = -24\delta^5 (\rho g)^{\frac{1}{3}} \frac{\partial}{\partial r} \left( r A_{sfc} \left( \frac{\partial h}{\partial r} \right)^3 \right)$$  \hspace{1cm} (4.5)

where $r$ is the radial (horizontal) coordinate. The derivation of equation 4.5 is given in Appendix C and the values of the parameters used in equations 4.4 and 4.5 are listed in Table 4.1.

A viscosity of $10^{14}$ Pa s is appropriate for ice of 1.0 mm grain size at 273 K (extrapolated from Goldsby and Kohlstedt [2001]). The value of $A_o$ we use ($10^{-25}$ Pa$^{-3}$ s$^{-1}$) corresponds to a viscosity of $10^{15}$ Pa s under an applied stress of $10^{-5}$Pa. Based on the slightly high viscosity used in our model, our estimate of the creeping layer thickness will be conservative. In our numerical simulations we consider $\phi$ values that range from 0.2 (dust-depleted ice) to the critical fraction at which the viscosity of a rock-ice mixture will be rock-dominated ($\phi = 0.72$) [Mangold et al., 2002].
Table 4.1: Definitions and measured or theoretical values (or range of values) for parameters used in the numerical simulations

<table>
<thead>
<tr>
<th>Description</th>
<th>Symbol</th>
<th>Value(s)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>activation energy</td>
<td>$Q$</td>
<td>50 kJ mol$^{-1}$</td>
<td>[Goldsby and Kohlstedt, 2001]</td>
</tr>
<tr>
<td>reference viscosity</td>
<td>$\eta_o$</td>
<td>$10^{15}$ Pa s</td>
<td>[Goldsby and Kohlstedt, 2001]</td>
</tr>
<tr>
<td>dust fraction</td>
<td>$\phi$</td>
<td>0.2–0.72 (upper limit)</td>
<td>[Mangold et al., 2002]</td>
</tr>
<tr>
<td>dust frac. coeff.</td>
<td>$c$</td>
<td>8</td>
<td>[Pathare et al., 2005]</td>
</tr>
<tr>
<td>density of creep layer</td>
<td>$\rho$</td>
<td>1.07–2.01 g cm$^{-3}$</td>
<td>based on $\phi$, above</td>
</tr>
<tr>
<td>gravity</td>
<td>$g$</td>
<td>3.7 m s$^{-2}$</td>
<td></td>
</tr>
</tbody>
</table>

4.2.2 Numerical Model

We modeled the relaxation of topography using a finite-difference discretization of equation 4.5 in which the only initial topographic variation is due to a single peak-to-peak sinusoidal curve. The peak-to-peak distance is the crater diameter ($D$), and the initial depth is set to $D/15$. This depth to diameter ratio was based on the maximum ratio observed in our analysis of crater topography and gives craters that are $\sim$40% deeper than the depth to diameter relationship given by Garvin and Frawley [1998] for craters 30 km in diameter. Parabolic crater profiles were also used as an initial condition, but only sinusoidal results are presented here. We use a relatively large depth : diameter ratio of 1/15 because it is the largest ratio we observed in our crater analysis, whereas the study by Garvin and Frawley [1998] is likely to include craters which have relaxed via viscous creep. The motivation for using a sinusoidal initial profile is to ensure that our constraint on the creeping layer thickness is as conservative as possible. These assumptions are discussed in detail in Sections 4.5.1 and 4.5.2.

The numerical simulation is radially symmetric, and requires that the north and south faces of the simulated crater be modeled independently. The topographic profile, with the initial sinusoidal crater center at the left edge, is divided into 200 elements, each 300 m in length. The
time step used in the simulation is 1000 yrs. Because the two halves of the crater are modeled separately, the elevation of the center of the crater will differ between the two simulations. Although this is not realistic, it is more computationally efficient than a three-dimensional simulation. At the crater center, no material is allowed to flow out, so the height of the crater center is set to that of the adjacent node.

### 4.2.3 Influence of Obliquity and Slope on Temperature

The latitude, local slope, and obliquity all influence the local temperature. We determine the temperature of a horizontal surface on Mars at a given obliquity using theoretical insolation calculations by Ward [1974]. The annual average equatorial temperature on Mars is currently 220 K [Mellon and Jakosky, 1995]. This temperature corresponds to an insolation of 180 W m$^{-2}$ based on calculations by Ward [1974]. We can relate insolation values at other latitudes to temperature using the following relation:

$$T_{\text{horiz}} = \left( \frac{I}{I_o} \right)^{1/4} T_o$$

where $T_{\text{horiz}}$ is the temperature of a horizontal surface and $I$ is the local insolation in W m$^{-2}$ given by Ward [1974]. $I_o$ is 180 W m$^{-2}$ and $T_o$ is 220 K. Figure 4.1 shows how temperature varies both as a function of latitude and obliquity in our simulations.

In simulations lasting less than or equal to 10 My, we allow obliquity to vary with time based on the recent obliquity history from Laskar et al. [2002]. Longer-term simulations require a statistical approach to determine the probability of different obliquity states because the Martian axial tilt history is chaotic over long timescales, and it is thus impossible to determine the exact obliquity at a given time in the past beyond about 30 Ma [Laskar et al., 2004]. The obliquity during a particular 100,000 yr time interval within the 100 My and 250 My simulations is
Figure 4.1: Local temperature of a horizontal surface on Mars as a function of latitude for the indicated obliquity state. Calculated using the method described in Ward [1974] and equation 4.6. Randomly selected based on a Gaussian distribution of obliquities given for the appropriate time into the Martian past (either 100 My or 250 My) determined by Laskar et al. [2004]. We assume the effect of slope is to change the local latitude by the angle of the slope. While not exactly correct, this assumption is a good approximation of more complicated approaches (e.g. Aharonson and Schorghofer [2006]). For example, an equator-facing slope $\theta$ at a latitude $\phi$ is assigned a temperature of a latitude ($\phi - \theta$). Before discussing the results from these simulations, we will first describe how we quantified slope asymmetry of craters in both the simulations and in the MOLA topography.
4.2.4 Slope Asymmetry

Based on the theory outlined above, we expect regolith creep to result in craters demonstrating a slope asymmetry with shallower equator-facing slopes than pole-facing slopes. We quantify the slope asymmetry in a similar fashion to Kreslavsky and Head [2003] by using the parameter

\[ A = \frac{S_n - S_s}{S_{ave}} \]  

(4.7)

where \( S_n \) and \( S_s \) are the maximum slopes on the north and south faces of the crater, respectively. In the simulations \( S_{ave} \) is the mean maximum slope for the two faces. A negative value for \( A \) indicates that the south face of the crater is steeper in slope than the north face. In the next section we describe the method used to measure crater slope asymmetry on Mars using gridded topography from MOLA.

4.3 Method of Crater Analysis

We used four criteria to select craters on Mars for slope asymmetry analysis. First, craters were chosen from seven latitude bands, centered on: 60°N, 47°N, 1.5°S, 20°S, 30°S, 40°S, and 60°S. The locations of these regions are listed in Table 4.2. Second, we selected craters based on size, ranging from 16 to 40 km in diameter. Our analysis requires craters that are large enough to be resolved in gridded MOLA topography, and small enough to respond to the diffusion of short wavelength topography. Also, we avoided overprinted craters, or craters with heavily modified rims. Finally, craters on steep regional slopes were not included in our survey to avoid potential bias in the calculation of \( A \).

The southernmost region analyzed in this study is located in Promethei Terra. 4.2a is Viking image of part of this region. Looking closer, 4.2b gives the context for one of the
craters we analyzed. We measured crater slope asymmetry using 18 radial topographic profiles taken every 20 degrees of azimuth overlain on gridded MOLA topography at a pixel resolution of 460 m (Figure 4.2c). Each profile extends from the center of the crater to the crater rim, or just beyond the rim. The topography was interpolated at a resolution of 375 m for all profiles. Four profiles transect the north and south crater faces and 5 profiles transect the east and west faces. For each crater face we averaged the appropriate profiles together to get a stacked profile (Figure 4.2d).

We then compared the maximum slopes from the north and south crater faces using
the stacked profiles to determine slope asymmetry. The crater slope asymmetry is quantified by the parameter \( A \), where \( A = (S_n - S_s)/S_{ew} \) where \( S_{ew} \) is the average of the maximum slopes from the stacked east and west profiles. The quantification of slope asymmetry in crater topography, above, is slightly different than equation 4.7 because the topography is 3D and the simulations are done in 2D.

### 4.4 Results

We use a 100 My simulation time as an assumed minimum age of craters 16 to 40 km diameter on Mars. The crater chronology of Amazonis Planitia presented by Hartmann and Neukum [2001] suggest craters 16 to 40 km diameter give a surface age of \( \sim 1 \) Gyr. Assuming a constant cratering rate, the average crater age would be 500 My. Even though we are analyzing craters that tend to be the least degraded (i.e. craters that are not overprinted), we can assume most craters of this size are older than 100 My. Results from 100 My simulations will give conservatively high estimates of the ice-rich creeping layer thickness if we are underestimating the true crater age. Running \( > 1 \) Gyr simulation times would be inappropriate given the current uncertainties associated with long-term climate evolution on Mars. The majority of our simulations model 100 and 250 My of crater relaxation.

#### 4.4.1 An Example Simulation

We calculate the theoretical change in topography of an initial crater profile over time using equations 4.1 through 4.6 with specified values for latitude, \( \delta \), and \( \phi \). The background surface temperature varies with time in the simulation due to the changing obliquity, as discussed in Section 4.2.3. An example of a modeled topographic profile is shown in Figure 4.3a.
Figure 4.3: a) Simulation of 100 My of topographic relaxation of a 30 km diameter crater at 40°N latitude assuming a deformable layer 125 m in thickness with a 80% dust fraction. The final crater has a slope asymmetry of $|A| = 0.240$. The discontinuity at 0 km is due to calculating the evolution of the two sides of the crater separately (see text). b) The temperature along the initial and final profiles varies because of the effect of slope on the angle of incident sunlight. The shift in the regional temperature between the two profiles is due to an obliquity of 37° initially compared to 26° obliquity at the end [Laskar et al., 2002].

for a 30 km diameter crater at 40° latitude with $\delta = 125$ m and $\phi = 0.8$ over a 100 My period. North is indicated by the arrow. Notice the asymmetry in slope that has developed in the final profile due to the hastening of creep on the equator-facing slope (right hand side of crater). Because the simulations are radially symmetric, the north and south crater faces must be modeled independently. Therefore, they give slightly different results for the crater depth resulting in a discontinuity in topography at the bottom of the crater. However, this discontinuity does not affect our calculation of the slope asymmetry. The value of $|A|$ in Figure 4.3a is 0.24. Figure 4.3b shows how the temperature along the initial and final profiles varies due to the effect of slope on the angle of incident sunlight. The shift in the regional temperature between the two profiles is due to a change in obliquity from 37° at the beginning of the simulation to 26° at the end.
Figure 4.4: Numerical model results of the crater asymmetry parameter ($|A|$) as a function of latitude and $\delta$. These results assume a 30 km diameter crater, a 40% dust fraction, a 100 My relaxation period, and the indicated deformable layer thickness $\delta$.

4.4.2 100 My Simulation Results

Figure 4.4 summarizes the result of 100 My simulations in which latitude and $\delta$ are varied.

The simulation results show that at high latitudes surface temperatures are too cold and flow is so slow that no significant asymmetry develops. At mid latitudes the regional temperature is higher relative to the poles. Also, the temperature difference between the two slopes is significant due to the differences in insolation between the north and south crater faces. The
Table 4.2: Crater statistics for each region based on topographic observations $^a$

<table>
<thead>
<tr>
<th>Latitude</th>
<th>Longitude</th>
<th># of Craters</th>
<th>$A$</th>
<th>Mean Diam (km)</th>
<th>Mean Depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>58–63° N</td>
<td>0–360° E</td>
<td>27</td>
<td>0.02±0.19</td>
<td>26.1±6.4</td>
<td>1.01±0.41</td>
</tr>
<tr>
<td>47–54° N</td>
<td>63–170° E</td>
<td>11</td>
<td>−0.034±0.32</td>
<td>23.7±5.3</td>
<td>0.62±0.47</td>
</tr>
<tr>
<td>0.25–2.75° S</td>
<td>30–70° E</td>
<td>11</td>
<td>−0.17±0.39</td>
<td>28.0±8.7</td>
<td>0.67±0.29</td>
</tr>
<tr>
<td>18–22° S</td>
<td>330–360° E</td>
<td>13</td>
<td>−0.155±0.28</td>
<td>29.8±8.1</td>
<td>1.15±0.34</td>
</tr>
<tr>
<td>28–35° S</td>
<td>130–150° E</td>
<td>23</td>
<td>−0.134±0.29</td>
<td>29.6±8.9</td>
<td>1.16±0.56</td>
</tr>
<tr>
<td>40–44° S</td>
<td>0–38° E</td>
<td>12</td>
<td>0.163±0.11</td>
<td>25.7±7.3</td>
<td>0.76±0.28</td>
</tr>
<tr>
<td>58–63° S</td>
<td>130–170° E</td>
<td>23</td>
<td>−0.07±0.25</td>
<td>29.9±7.1</td>
<td>0.91±0.36</td>
</tr>
</tbody>
</table>

$^a$Errors are ±1 standard deviation.

result is a local maximum in asymmetry at between 30° and 45° latitude, depending on the value for $\delta$. Finally, near the equator, temperature is approximately the same for the north and south crater faces so no asymmetry can develop. As expected, the rate of flow and thus the development of asymmetry is a strong function of $\delta$ (equation 4.5). One important result from these simulations is the decrease in $|A|$ at low to mid-latitudes when the ice-rich layer thickness is increased from 100 m to 150 m. This decrease in $|A|$ is a result of craters relaxing to the point that the opposing crater faces become more similar in slope with time. Thus, highly relaxed craters will have a lower value of $|A|$ than moderately relaxed craters. This effect will be discussed further in Section 4.4.5.

4.4.3 Observational Results

The observed crater slope asymmetry in the six regions we analyzed is summarized in Table 4.2. In no case was a mean value of $|A|$ obtained that differed significantly from zero, indicating that the predicted systematic slope asymmetries (Figure 4.4) are not observed. Figure 4.5 compares the observed crater slope asymmetry to results from topographic relaxation simulations lasting 100 My. The colored dots are measured slope asymmetries for individual
craters. The black squares and vertical lines indicate the mean and $\pm 1$ standard deviation in $A$, respectively, for craters in a given latitude region. As shown in Figure 4.5, there is no statistically significant slope asymmetry at any of the latitude regions we analyzed. This observation provides an upper limit on the extent to which creep has occurred. Since the creep rate depends mainly on $\delta$ and $\phi$, the characteristics of any creeping ice-rich layer present can therefore be constrained. The black curves in Figure 4.5 show the expected asymmetry signal for a 30 km diameter crater evolving over a 100 My period for various creeping layer thicknesses with $\phi = 0.4$. The fine line shows results from simulations with a higher dust fraction ($\phi = 0.7$) and a 100 m thick deformable layer.

These results show that the mid-latitude slope asymmetry observations provide the strongest constraint on the creeping layer characteristics. The $\delta = 50$ m, $\phi = 0.4$ curve and the curve with $\delta = 100$ m and $\phi = 0.7$ marginally fit within one standard deviation of the observations. These combinations of $\phi$ and $\delta$ provide an upper bound on the creeping layer characteristics. As expected, there is a trade-off between $\delta$ and $\phi$ in generating crater slope asymmetry. As will be discussed in Section 4.4.5, longer timescale simulations will result in tighter constraints on $\delta$ and $\phi$ as one would expect.

### 4.4.4 Trade-offs Between $\delta$, $\phi$, and Simulation Time in Generating Slope Asymmetry

To further explore the trade-off between $\delta$, $\phi$, and creep duration we carried out a suite of runs lasting 10, 100, and 250 My. These simulations used values for $\delta$ and $\phi$ that ranged from 25 m to 175 m and from 0.2 to 0.8, respectively. Based on the observed slope asymmetries (Figure 4.5), simulations that produced slope asymmetry values for mid-latitudes
Figure 4.5: Observed crater asymmetry measured at various latitudes (open and filled dots) compared to theoretical results (thick solid and dashed curves) for varying creeping layer thicknesses ($\delta = 50, 100$ and $150$ m) from Figure 4.4 with $\phi=0.4$ over a 100 My period. The black squares and vertical lines indicate the mean asymmetry values and $\pm 1$ standard deviation, respectively, for a particular latitude region. The dashed line shows results from simulations with a high dust fraction ($\phi = 0.7$) and a 100 m thick deformable layer.
Figure 4.6: Asymmetry contours of $|A| = 0.2$ given values for $\delta$, $\phi$, and simulation duration. A crater diameter of 30 km, and a latitude of $40^\circ$ were used in these simulations. The domain of $|A| \leq 0.2$ for each simulation duration is indicated by the tick marks.

$(40^\circ)$ with $|A| < 0.2$ were deemed consistent with the observations. The tick-marked lines in Figure 4.6 show the domain of $\delta$, $\phi$, and creep duration that give a slope asymmetry value of 0.2 or less. As expected, a longer simulation duration results in tighter constraints on $\delta$ and $\phi$.

For a given value of $|A|$, higher values of $\phi$ permit larger thicknesses $\delta$, and vice versa. For 100 My simulations with a dust fraction near the viscous transition to a solid particle rheology $\phi \approx 0.7$ [Mangold et al., 2002] give a maximum layer thickness of $\sim 100$ m. A smaller dust fraction ($\phi = 0.2$) results in a correspondingly smaller maximum layer thickness ($\sim 60$ m). For the 250 My simulations these maximum layer thicknesses are reduced to $\sim 90$ m and $\sim 45$ m,
respectively.

4.4.5 250 My Simulation Results

Our longest crater relaxation simulations lasted 250 My. We chose reasonable values for $\delta$ and $\phi$ (150 m and 0.4, respectively) based on the shorter simulations. Initial and final topography and slope of this simulation are shown in Figure 4.7a and b, respectively. The evolution of crater's north-south slope asymmetry (black line) and depth to diameter ratios for the north and south crater faces (red lines) during the simulation is shown in Figure 4.7c.

Once the crater has infilled beyond about one quarter of its original depth the slope asymmetry begins to decrease with time as shown in Figure 4.7c. This decrease in $|A|$ for highly softened craters is due to a reduction in the creep rate at low slopes (equation 4.5). After about one quarter of the crater has been infilled, the shallower of the two crater slopes experiences a slowed creep rate due to its low slope, allowing the creep process on the other crater face to catch up. As the crater topography continues to relax, the two slopes become more similar, resulting in a decrease in $|A|$. Because highly relaxed craters have a low $|A|$ value, 250 My simulations with a large $\delta$ and small $\phi$ result in craters that satisfy the $|A| < 0.2$ agreement with the observations. However, as will be discussed later in this section, observed depth : diameter ratios for the craters we analyzed suggest highly relaxed craters are not found in the regions and size range used in this study.

The simulations were carried out using a time-dependent obliquity calculated using the statistical approach described in Section 4.2 (Figure 4.7d). In this figure we plotted obliquity every million years although it varied every 100 time steps (100,000 yrs) in the simulation to make the plot readable. At high obliquity ($>55^\circ$) the temperature of a horizontal surface gets
Figure 4.7: a) Simulation of 250 My of topographic relaxation of a 30 km diameter crater at 40° latitude assuming a deformable layer 150 m in thickness with a 40% dust fraction. The final crater has a slope asymmetry of $|A| \approx 0.18$. b) The slope along the initial and final topographic profiles. c) The change in slope asymmetry ($A$) over time is shown by the solid black line, and the dashed lines indicate the change in the depth : diameter ratio ($d/D$) for the north and south crater faces. d) Obliquity variation over time using the statistical approach described in Section 4.2.3. We plotted obliquity every $10^6$ years although it varied in the simulation every 100 time steps (100,000 yrs).
warmer toward the poles [Ward, 1974] resulting in a reversal of the normal variation in creep rates. On average, however, the obliquity is about 34° over 100 to 250 My timescales [Laskar et al., 2004]. Thus, during most of the numerical simulation, the temperature is decreasing toward the poles resulting in craters with a negative value for $A$ in the northern hemisphere and a positive value in the southern hemisphere.

As the crater infills during our simulations, the diameter increases as the creep flow softens the crater rim. The combination of infilling and crater widening cause the crater’s depth : diameter ratio to decrease over time depending on the creep rate. Comparing depth : diameter ratios from the numerical model and the observations places an additional constraint on the creep rate. Figure 4.8a shows depth : diameter ratio plotted against $A$ for both simulation results and observations. The depth : diameter ratio for the equator and pole-facing slopes are given by the solid and dotted lines, respectively. Although the simulation is for a crater at 40°S latitude, observations from all the latitude regions we analyzed are plotted to illustrate the range in observed depth : diameter ratio. The colored boxes give mean values for $A$ and depth : diameter ratio for a particular latitude region with $\pm 1$ standard deviation given by the error bars. The simulation starts with a depth : diameter ratio of 0.067 (or 1/15) which decreases with time as the crater relaxes. The cross-hairs are separated by 25 My of simulation time; $\delta = 150$ m and $\phi = 0.4$ in this 250 My simulation. The end-point of the simulation shown in Figure 4.8a has $|A| \approx 0.18$ and a depth : diameter ratio near the lower extreme of the observed craters. The lack of low depth : diameter ratio craters on Mars suggests few 16-40 km diameter craters have relaxed to the degree shown in Figure 4.7 or 4.8a. Increasing $\phi$ or decreasing $\delta$ results in depth : diameter ratios more comparable to those observed; Figure 4.8b shows that an increase in $\phi$ to 0.6 results in a better match to the observations. A similar effect could be
obtained by decreasing $\delta$.

A depth : diameter ratio versus $A$ plot for a crater relaxation simulation at $60^\circ$S latitude is compared to observations from $60^\circ$S and $60^\circ$N in Figure 4.8d. The negative of the mean $A$ value for $60^\circ$N is plotted here in order to compare it with the southern latitude simulation. This simulation uses $\delta = 175$ m and $\phi = 0.6$. Figure 4.8c gives a similar plot for a simulation at $25^\circ$S with $\delta = 125$ m and $\phi = 0.6$. Observations from 30 and $20^\circ$S are shown for comparison. In both cases the end-point of the simulation (250 My) produces values of $A$ and depth : diameter ratio consistent with the observations.

4.5 Discussion

4.5.1 Summary of Results

The 100 My simulation results shown in Figure 4.6 suggest the ice-rich layer at $40^\circ$ latitude could be either 60 m in thickness for clean ice or, alternatively, 100 m thick with a dust fraction of 0.7. Given the range of asymmetries in the observations, values for $\delta$ and $\phi$ are likely to be both spatially and temporally heterogeneous. At latitudes higher than $40^\circ$, crater relaxation is slowed by colder temperatures and a thicker creeping layer will result in a north-south slope asymmetry that still lies within the range of the observations.

One result from the 250 My simulations is that the age of the crater is important in determining the slope asymmetry that we observe today. Figure 4.7 shows that the maximum slope asymmetry occurs after about 100 My of topographic relaxation. Subsequently, the slope asymmetry slowly decreases as more material infills the crater. The time at which the maximum asymmetry is reached depends on the size of the crater, the thickness of the creeping layer, the dust fraction, and latitude. The faster the creep of regolith, the sooner the maximum slope asym-
Figure 4.8: a) Simulation results and observations of depth : diameter (d/D) ratio plotted against A. The equator- and pole-facing slopes are given by dashed and solid lines, respectively. This simulation lasts 250 My for a crater at 40°S latitude. The data from Figure 4.7c are used in this plot. The cross-hairs are separated by 25 My of simulation time; δ=150 m and φ=0.4 in this simulation. The colored boxes give observed mean values for A and d/D ratio for a particular latitude region with ± one standard deviation given by the error bars (Table 4.2). b) Same as a) except φ is increased to 0.6 and the simulation results are compared to mid-latitude crater observations only. c) Similar plot for a simulation at 25° with δ = 125 m and φ=0.6. Observations from 30 and 20°S are shown for comparison. d) d/D ratio versus A plot for a crater relaxation simulation at 60°S using a δ of 175 m and φ of 0.6. Observations from 60°S and the negative of the mean A value for 60°N are plotted here in order to compare it with the southern latitude simulation.
metry will be reached. Based on the observed depth : diameter ratios, the craters we analyzed on Mars have not relaxed so much that the slope asymmetry is decreasing with time (Figure 4.8a). Rather, the observed slope asymmetry is occurring in the early to mid stage of crater softening. Figure 4.8d suggests, for a dust fraction of 0.6, that a creeping layer thickness of 175 m gives values of $A$ and depth : diameter ratio that are in close agreement with the observations at 60° latitude. At lower latitudes, simulations with a progressively thinner creeping layer (150 m at 40° latitude and 125 m at 25° latitude) with $\phi = 0.6$ match the observed depth : diameter ratios and $A$ values (Figure 4.8b and c). A thickening creeping layer toward the poles is supported by cryosphere models put forth by Clifford [1993]; Fanale et al. [1986], and Kuzmin et al. [1989].

4.5.2 Model Assumptions

Our numerical model is less sophisticated than that used by Pathare et al. [2005] (although the basic physics is the same); the advantage of our simpler model is the ability to explore parameter space rapidly. We assume an exponentially increasing dust fraction with depth attributed to the incorporation of more rocky material (equation 4.3). We also assume a non-linear relationship between shear stress and shear strain rate based on Glen’s flow law. Ice is assumed to be present at/within the surface during the simulations, extending to a constant depth scaled according to the creeping layer thickness. As described in Mellon and Jakosky [1995], surface ice is currently unstable at latitudes less than $\sim 60^\circ$, resulting in a desiccated layer extending a few meters into the Martian regolith. However, these authors also show that surface ice is stable at all locations on Mars at obliquities $> 32^\circ$. The thickness of the desiccated layer is small compared to the creeping layer thicknesses ($\sim 100$ m) used in our simulations, and will have little effect on its rheology. The presence of ice deep within the Martian subsurface
has been suggested by Clifford [1993], extending to perhaps 10 km. We have not accounted for where ice is, or was, stable in the Martian subsurface, rather we have simply modeled how craters would respond to ice-rich regolith creep assuming that ice is present. One should keep in mind, however, that the instability of ice at the surface does not necessarily correlate with a lack of ice at depth.

Many of the assumptions used our numerical model were made to ensure that our estimate of the ice-rich layer thickness was an upper limit. For instance, we use a sinusoidal initial topographic profile instead of a parabolic profile because the slopes of the parabolic profile are more steep and result in a larger slope asymmetry as the parabolic profile begins to relax. In general, a larger creeping layer thickness generates more asymmetric crater slopes as long as the crater doesn’t become significantly infilled (Figures 4.4 and 4.6). Using a sinusoidal initial profile forces the creeping layer to be thicker than it would be for a parabolic profile in order to produce the same slope asymmetry.

Another conservative assumption used in our model is the viscosity at the surface of the ice-rich creeping layer \( \eta_o \) and the vertical porosity structure given by equation (4.4). The \( 10^{15} \) Pas surface viscosity is for rather coarse-grained ice (2.0 mm) [Goldsby and Kohlstedt, 2001] at 253 K and \( 10^5 \) Pa giving a higher creeping layer viscosity than would be present for finer grained ice. Assuming a relatively high viscosity slows the rate of crater relaxation, making our estimate for the creeping layer thickness an upper limit.

Finally, assuming a relatively large depth : diameter ratio (compared to Garvin and Frawley [1998]) ensures that our creeping layer thickness is an upper limit. Although a larger depth : diameter ratio results in steeper slopes, the creeping layer thickness must still be greater in order to produce a softened crater with a depth : diameter ratio that lies within the range of...
observations (Figure 4.8).

Although we have accounted for obliquity variations in our simulations, we did not address the effect of a time dependent eccentricity, nor variations in the longitude of periapse. Both of these effects will have a smaller effect on the temperature distribution than obliquity variations [Laskar et al., 2002].

4.5.3 Additional Remarks

Although the craters we analyzed are complex based on their size [Melosh, 1989], few central peaks were observed. This finding suggests infilling has occurred, and that, if this infill is due to viscous creep of an ice-rich regolith, then a slope asymmetry ought to also be observed. However, there remain many questions regarding the mechanism by which craters are softened on Mars. For instance: why are no systematic crater slope asymmetries observed in any of the regions we studied on Mars, especially in the high latitude regions?

One possibility is that the ice-rich layer thickness varies spatially within a crater. At low to moderate obliquities the pole-facing crater slope is colder than the equator-facing slope. However, creep hastening on the warm slope might be subdued if the ice-rich layer is thin. If the cold slope has a thicker creeping layer, the creep rates for the north and south slopes could be approximately equal - resulting in no slope asymmetry even though topographic relaxation proceeds. The preferential surface deposition of ice-rich mantling material on pole-facing slopes [Aharonson and Schorghofer, 2006] might also affect the topography and reduce the effective north-south topographic asymmetry. However, the thickness of mantling material required to change the slope on a 16 km diameter crater by 5° is about 700 m. This thickness seems unlikely given the 10 m mantling thickness estimates provided by Mustard et al. [2001].
Aeolian deposition has also been invoked as a means of infilling craters [Zimbelman et al., 1998]. However, this hypothesis cannot explain the latitudinal distribution of crater softening. The amount of material needed (700 m thickness) would take $25 \times 10^6$ yrs at current deposition rates measured at the Mars Pathfinder landing site [Johnson et al., 2003].

Based on our numerical results, one mechanism that does not appear able to explain the observations is temperature variations induced by changes in obliquity. Although the angle of incident sunlight and year-average temperature varies with obliquity, our simulations (e.g. Figure 4.7) suggest that obliquity variations are not sufficient to override the development of shallower gradients on equator-facing slopes in craters 16 to 40 km in diameter.

Kreslavsky and Head [2003] suggest that preferential erosion of equator-facing slopes at mid-latitudes due to melting ground ice at high obliquity is responsible for the north-south slope asymmetry that they observe in global surface roughness measurements at a 300 m length-scale. Equator-facing slopes are systematically steeper than pole-facing slopes between 40 and $50^\circ$ latitude in both hemispheres. This effect could be due to melting ground ice [Kreslavsky and Head, 2003] or shallow active permafrost layer processes [Kreslavsky et al., 2008], both occurring at periods of high obliquity. Our measurements are made on a specific class of features (craters) at a much longer length scale ($\sim 30$ km), and do not show a systematic slope asymmetry at the latitudes we analyzed. There are at least two possible reasons for the discrepancy. First, the process proposed by Kreslavsky and Head [2003] may not be effective at changing slopes over the much longer baselines used in our study. Second, craters may not undergo the same modification processes as other regions because of the manner in which they formed. For instance, the initial impact will likely have driven off local volatiles [Senft and Stewart, 2008]; subsequent diffusion of ice into the empty pore space will certainly occur, but on a timescale
that depends on the poorly-known permeability structure of the regolith [Clifford and Hillel, 1983; Fanale et al., 1986]. One potential way of distinguishing between these two hypotheses would be to look for slope asymmetries in craters smaller than those investigated here.

4.6 Conclusions

Topographic observations of craters on Mars indicate that there is no statistically significant dependence of crater slope asymmetry on latitude (Section 4.4.3, Table 4.2). Generally, slope asymmetry ranges between $A = -0.2$ and $A = 0.2$ in all of the seven regions we analyzed. Numerical models of crater relaxation using reasonable creeping layer thicknesses and dust fractions suggest that the formation of significant slope asymmetry should occur over $\sim 100$ My timescales. Comparison between topographic observations of craters on Mars and our numerical models suggest that a clean ice-rich layer on Mars does not extend deeper than about 125 m at mid-latitudes. If the creeping layer responsible for crater softening on Mars contains 50 - 70% dust by volume, then the creeping layer thickness could extend to 150 m. At low latitude the creeping layer is likely thinner, perhaps 100 m or less, based on depth : diameter ratio comparisons between the simulations and the observations (Figure 4.8). These thickness estimates are conservative and would be lower if, for example, longer simulation times were used.

It is unclear why the mean crater slope asymmetry in a particular region is so close to zero. Most puzzling is the number of craters with values of $A$ that are opposite in sign to what one would expect theoretically (Figure 4.5). A range of crater asymmetry values (of a given sign) within a particular region on Mars is expected because crater slope asymmetry evolves over time (Figure 4.7c). We expect there to be some scatter in $A$ due to the range of
crater ages present within a particular region, but not a variation in sign as a result of creep-related processes. One possibility is simply that regolith creep is sufficiently slow (e.g. the creeping layer is thin) that this process is overwhelmed by other geological processes such as gully formation, tectonic deformation, or aeolian deposition. Also, glacial flow of surficial ice and permafrost-driven (e.g. solifluction) creep [Perron et al., 2003] may dominate crater modification.

The strong dependence of crater softening on wavelength [Jankowski and Squyres, 1992; Pathare et al., 2005] is another possible explanation of why no systematic crater asymmetries are observed. Because short wavelength features will relax more quickly than longer wavelengths, it is possible that smaller craters than those we have examined will show systematic crater slope asymmetry. Alternatively, an increase in the creeping layer thickness on cold slopes coupled with a reduction of ice-rich material on warm slopes would significantly reduce the expected crater slope asymmetry.

Assuming a 125 m thick global near-surface, ice-rich layer incorporating a dust fraction of 0.2 accounts for $\sim 1.5 \times 10^7$ km$^3$ of water. This upper-limit volume estimate is a factor of 12 smaller than the Noachian ocean volume proposed by Clifford and Parker [2001], and is a factor of 4 short of the end-state of the hydrosphere evolution model that these authors propose. The extent of the northern ocean proposed by Baker et al. [2000] (reaching Olympus Mons) exceeds our upper limit water volume estimate by a factor of 20. More modest-sized oceans [Parker et al., 1989, 1993; Carr and Head, 2003], extending to the outer Vastitas Borealis formation, are comparable to our estimated volume.
Chapter 5

Summary

5.1 Findings and interpretations

Young, water-related geologic features on Mars have led to a shift in the perception of the Red Planet from a place where all the hydrologic activity was buried deep in the past, to a place where water and ice still play an active role in modifying the surface. The recent data acquired from spacecraft missions has sparked a renaissance in Mars science in which an understanding of its climate history is beginning to take shape. Reproducing observations through numerical modeling of physical processes is an important technique by which hypotheses relating to the Martian climate can be tested and new discoveries can be made. Using this approach we have been able to place constraints on how some features relating to the past climate formed. Based on our findings we can make some interpretations regarding the role of water and ice in shaping the Martian surface in the recent past.
5.1.1 Gullies

By analyzing the formation of gullies via fluvial erosion we found that sediment transport rates would be very rapid resulting in gully formation in a matter of days to months under constant stream-flow conditions in a channel initially 1 m deep by 8 m wide, and 0.1 m deep by 3 m wide, respectively. However, we found no potential reservoir of water that could supply the volume of water required to form the gullies in our study location over such a short time span. Therefore, we can conclude that fluvial erosion is not the only process by which sediment is carved out of gully alcoves, and/or that ∼ 10 episodes of groundwater release separated by enough time to recharge the supplying aquifer (∼100 kyr) is required. We deem the melting of a seasonal snow pack and hydrothermal springs as insufficient and unlikely, respectively, sources of water to drive fluvial erosion in gully alcoves, but they may be important processes in recharging an aquifer which supplies water to gully alcoves.

The observation of gully fans with significant differences in crater densities across their surface [Schon et al., 2009, e.g] as well as the existence of gullies at different stages of maturity (Figure 2.1) supports the episodic activity hypothesis. Also, after analyzing the potential sources of water to drive fluvial sediment transport in gullies, we found that only episodic release and recharge of water from an aquifer with a thickness of at least 30 m was sufficient in supplying the necessary volume of water. I interpret these results as evidence against the role of fluvial transport, and in favor of debris flows transporting the majority of the sediment involved in gully formation. In sum, these findings suggest that the influence of liquid water in forming Martian gullies may have been exaggerated, and, instead, comparatively small amounts of liquid water episodically destabilized the steep gully slopes and occasionally flowed onto the surface to form these features.
5.1.2 LDAs

Our work simulating the flow of ice to form LDAs has led to a few important findings regarding the nature of these ice deposits. First, if LDAs are nearly pure ice, as the ShaRAD data suggest, then an ice grain size of 5 mm at a temperature of 205 K gives an appropriate viscosity for these deposits to flow the observed distance over $\sim 100$ My timescales. Based on the ice grain size-limiting effect of incorporated dust, this grain sizes suggest a dust volume fraction of 0.16% for dust grains 50 $\mu$m in diameter. Surprisingly (at least to me), we can use this data to constrain the precipitation rate assuming deposition of snow was responsible for forming LDAs. Assuming this is the case, then the range of likely dust fractions ($\leq 0.16\%$) places a lower limit of 1.2 cm yr$^{-1}$ on the precipitation rate assuming a dust particle radius of $r_d = 50$ $\mu$m (Section 3.7.2). Although this estimate is consistent with independent predictions from climate models, it is very sensitive to our choice of dust grain size. Assuming a precipitation rate of 1 cm yr$^{-1}$ a minimum period of $3 \times 10^4$ to $10^5$ yr would be required to form LDAs. This result is an example of how quantitative geomorphic models of features observed at the surface can constrain past climate conditions.

Equally significant is the fact that, using a single combination of ice grain size and temperature, our model does not provide reasonable age estimates for all the LDAs simultaneously. Therefore, we must invoke regional heterogeneity in some or all of the variables in our model (such as changing the basal slope) to explain these large apparent age differences. This implies that we may be able to use our model to predict where a basal slope is present at a location where there is no ShaRAD data.
5.1.3 Crater softening

Although we initially expected to observe a slope asymmetry in Martian craters due to differential insolation-driven heating of an ice-rich regolith, topographic observations did not support such a phenomenon. However we were able to use the lack of observed slope asymmetry together with depth to diameter ratio measurements to constrain the maximum amount of ice present in the regolith. We found that regolith at mid-latitude and equatorial regions must be relatively immobile beyond a depth of 125 m and 100 m, respectively. The total amount of water associated with these creeping layer thicknesses is significantly smaller than that contained in a proposed northern hemisphere ocean [Baker et al., 2000; Parker et al., 1989, 1993; Carr and Head, 2003] and is smaller than the amount invoked in a saturated regolith model proposed by Clifford and Parker [2001]. These results would suggest that either much of the water that was present early in Mars history was lost to space subsequent to the formation of a northern ocean, or that there are other processes dominating crater modification than that used in our model.

Our model assumed abundant ice in the regolith - with surface concentrations of $\sim 60\%$ by volume. However, ice may not be present in these concentrations at the depths we considered, in which case the regolith would be relatively immobile. If this is the case, we must explain crater softening by some other mechanism or combination of processes which may include: dust deposition, lava flows, deposition of snow packs, and gully formation. Because softened craters are generally located at high latitudes, a climate-based process is likely responsible for their formation. Characterization of the surface morphology of crater floors occupied by concentric fill, generally found at mid to high latitudes, is nearly identical to that found on the surface of LDAs suggesting that some of these craters may currently be filled with massive ice deposits below a protective regolith layer [Levy et al., 2010]. If future radar observations
support this hypothesis, crater softening may be a symptom of episodic ice ages like those that formed LDAs rather than a long-term regolith creep process like the one we investigated here.

However, not all craters at high latitudes contain concentric fill. An analogy to the relationship between LDAs and viscous flow features [Milliken et al., 2001], and “ghost” LDAs [Hauber et al., 2008] may provide a possible explanation. Viscous flow features and “ghost” LDAs provide evidence for past episodes of LDA formation where the ice is no longer present. Similar, prior episodes of massive ice deposition on craters at mid to high latitudes may have smoothed the crater topography as ice flowed into the crater interior. After forming a crater with subdued topography, these ice deposits may have then been removed by a change in obliquity conditions. This hypothetical scenario would not form a slope asymmetry because, instead of a viscously deforming regolith modifying the crater, erosion by flowing ice would cause the crater topography to become subdued. Testing this hypothesis is a project for future work.

5.2 Future work

As is usually the case, with new scientific findings come yet more questions to be answered. Given the data currently available, this work is limited to only placing constraints on the viable mechanisms which could have formed the geomorphology we observe. We have not yet been able to conclusively determine how the climate has modified the Martian surface in recent times - we have merely made steps toward such a finding. Much work remains to be done.

Specifically, further work regarding LDA formation should be conducted to verify our model’s ability to predict the presence of basal slopes and to reproduce LDA profile shapes. This will require careful selection of MOLA profile tracks that are parallel to the LDA flow.
direction, and do not sample convergent or divergent flow. This data would also allow a better assessment of the influence of variables such as flow direction and latitude on the predicted age of LDAs. Alternatively, a more sophisticated, three dimensional model could be developed in order to constrain the temperature, dust content, and basal slope variables by comparing the observed LDA topography directly with the model. More observations using visual data should be made to identify deformed craters superposed on LDA surfaces that could be used as strain markers that could better constrain the extent of the initial ice deposit.

This work could be broadened to study craters with concentric fill. In some instances, the concentric crater fill terrain actually appears to flow outside of the confines of the crater [Levy et al., 2010; Dickson et al., 2010] allowing the flow distance to be precisely determined, and a lower-bound on the flow thickness to be made. In craters containing concentric fill in its classic form, work similar to the LDA project could be carried out. Namely, deformed craters on the surface of concentric fill deposits provide information about the minimum flow distance. Also, more preliminary work on the distribution and topography of softened craters and craters with concentric crater fill could provide insights into the regional extent of recent glaciation events. Assuming glacial flow is responsible for smoothing crater topography, this data would help determine the amount of sediment carried by the ice deposit as it flowed over the surface. Finally, applying a radially symmetric version of the numerical model used in this work could constrain the volumes of ice and flow timescale(s) involved in forming concentric crater fill.

Further investigations of the relationship between gullies, LDAs, concentric fill, and lineated fill found within mid-latitude valleys may yield a better understanding of the timing, geographic extent, and quantities of ice involved in recent glaciation events. Subjecting these features to future observations and modeling efforts will lead to further constraints on past
climate conditions.

5.3 Looking ahead

The effects of climate change on Mars have made their mark on the surface geomorphology and recent missions are now collecting the data necessary to significantly advance our understanding of how and when these changes took place. Although we are far from understanding the history of Mars to the degree that is required to make detailed comparisons with Earth’s climate history, we are taking the first steps toward that goal. Eventually, if we are to pursue Mars exploration to the degree necessary to make comparisons with Earth, we will need to deploy missions that collect the same data we use to study this planet. Namely, seismometers to study Martian impacts, volcanism and the Martian interior, mass spectrometers to age date samples using radioisotopes, and ice-coring devices to read the recent climate change record within the polar caps. Although these types of missions may seem far-fetched, the first generation of these missions will be launched within the next decade with NASA’s Mars Science Laboratory rover “Curiosity” studying the geochemistry and atmospheric isotopic composition and the MAVEN orbiter mission studying the rate of atmospheric gas loss from the exosphere.

However, before these “far-fetched” missions take flight, there is still much to be done using the current data. Reproducing observations through numerical modeling of physical processes is an important technique by which hypotheses can be tested and new discoveries can be made using the existing data.
Appendix A

Derivation of groundwater discharge rate equation

Turcotte and Schubert [2002] give the following linearized expression for the evolution of the phreatic surface \( h \) in an unconfined aquifer:

\[
\frac{\partial h}{\partial t} = \frac{\kappa \rho g h_0}{\mu \phi} \frac{\partial^2 h}{\partial x^2} \quad (A.1)
\]

where \( h_0 \) is the height of the phreatic surface prior to uncapping of the aquifer, \( x \) is the distance from the aquifer cap and other variables are defined in Section 2.5.1. If the height of water in the channel is \( h_1 \) then the solution to equation A.1 is

\[
h(t, x) = (h_0 - h_1) \text{erf} z + h_1 \quad (A.2)
\]

where the similarity variable \( z = (\mu \phi / \kappa \rho g h_0 t)^{1/2} x / 2 \).

The Darcy velocity \( u \) is given by

\[
u = -\frac{\kappa \rho g}{\mu} \frac{\partial h}{\partial x} \quad (A.3)
\]
Together with the definition of \( z \), equations A.2 and A.3 may be used to derive the Darcy velocity at the aquifer cap \((z=0)\). Setting \( h_1 = fh_0 \), the discharge rate per unit width \( q_w \) is given by

\[
q_w = h_1 u = \left( \frac{\kappa \rho g \phi}{\mu \pi h_0 t} \right)^{1/2} h_0^2 f (1 - f) \quad (A.4)
\]

Taking the factor \( f (1 - f) \) to be of order unity then results in equation 2.9 where we have identified \( h_0 \) with the aquifer thickness \( T \). In practice, \( f (1 - f) \leq 0.25 \), which means our discharge rates overestimate real values by at least a factor of 4, resulting in an under prediction of aquifer thicknesses. Note, however, that the original assumption of linearity (equation A.1) is violated if \( f \ll 1 \).
Appendix B

Radar constraints on basal slopes below LDAs

Observations from ShaRAD in the Deuteronilus Mensae region provide evidence for both flat and sloping surfaces beneath LDA deposits [Plaut et al., 2009]. Although only horizontal reflectors are mentioned in their work, using the data published by Plaut et al. [2009], we found evidence for a sloping substrate beneath some LDAs. In order to determine the orientation of a buried reflector, one must convert the time it takes for the reflected radar signal to return to the spacecraft into a distance (known as migrating the signal). This distance \( d \) is given by

\[
d = \frac{c}{2\epsilon} \Delta t
\]  

(B.1)

where \( c \) is the speed of light in a vacuum, \( \epsilon \) is the dielectric constant of the material through which the incident and reflected radar wave traveled, and \( \Delta t \) is two-way travel time (the time it takes for the radar to travel from the surface, to the reflector and back). When the dielectric constant for ice (\( \epsilon = 3.3 \)) is used to migrate the reflected radar signal below LDAs, many of the reflected signals change from a curving line when plotted in the time domain to a flat line that is continuous with the surrounding topography when plotted in the depth domain (Figure B.1a,b).
Figure B.1: ShaRAD radar data published in Plaut et al. [2009] for LDAs at 38.6°N, 24.2°E. and 42.1°N, 18.5°E (corresponding to LDA #15/16 and #24.5) in panels a) and c), respectively. These radargrams plot the strength of the returned signal as a function of time (the same scale bar applies to both a and c). b) After applying a dielectric constant of $\epsilon = 3.3$ assuming the presence of ice within the LDAs the reflected signal (indicated by the black arrows) becomes flat and coplanar with the surrounding topography. d) Applying $\epsilon = 3.3$ to the radargram in part c produces a subsurface slope indicated by the dotted line. The thin, solid lines serve as visual guides with slope values of 0° and 0.5°.

This continuity between the surrounding topography and the reflected signal is evidence supporting the presence of ice within LDAs. The example shown in Figure B.1a,b corresponds to LDA #24.5 shown in Figure 3.2.

However, one radargram that is not migrated in Plaut et al. [2009] (Figure 2 in their paper and (Figure B.1c here) gives a reflector that is sloping when a dielectric constant of 3.3 is used to convert time to depth (Figure B.1d, dotted line). This LDA, located at 42.1°N, 18.5°E (corresponding to LDA #15 and 16 in Chapter 3) has a basal topography (dotted line) that can be approximated by a 0.5° sloping surface. The significant slope below this particular LDA which may also exist below other LDA deposits (see Section 3.6).
Appendix C

Derivation of the crater relaxation equation

Here we derive equation 4.5, which describes the relaxation of topography resulting from the flow of ice-rich regolith downslope. We assume an exponentially increasing viscosity with depth (equation 4.3), a non-Newtonian rheology (equation 4.2), a constant e-folding viscosity depth $\delta$, and conservation of mass. Combining equations 4.1 and 4.2 gives:

$$\frac{\partial v}{\partial z} = A_{sfc} e^{z/\delta} \left( \rho g z \frac{\partial h}{\partial r} \right)^3 \quad (C.1)$$

Integrating equation C.1 and applying the $v = 0$ at $z = \infty$ boundary condition gives the vertically resolved downslope velocity field for ice-rich regolith creep

$$v(z) = A_{sfc} \left( \rho g \frac{\partial h}{\partial r} \right)^3 \left( -\delta e^{-z/\delta} z^3 - 3\delta^2 e^{-z/\delta} z^2 - 6\delta^3 e^{-z/\delta} z - 6\delta^4 e^{-z/\delta} \right). \quad (C.2)$$

To translate this velocity field into a change in topography, we use the following, radially symmetric, conservation of mass relationship:

$$\frac{\partial h}{\partial t} = -\frac{1}{r} \frac{\partial}{\partial r} \left( r \int_0^\infty v \partial z \right) \quad (C.3)$$

where $r$ is the radial (horizontal) coordinate. Finally, we combine equations (C.2) and (C.3) to obtain the relaxation equation given in (4.5).
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