GEOLOGY, GEOMORPHOLOGY, AND HYPSOMETRY OF SOUTHEAST MARGARITIFER TERRA, MARS, AND THE LOWER ESCALANTE RIVER, UTAH

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ABSTRACT

GEOLOGY, GEOMORPHOLOGY, AND HYPsometry OF SOUTHEAST MARGARITIFER TERRA, MARS, AND THE LOWER ESCALANTE RIVER, UTAH

Corey Matthew Fortezzo

The valley networks and channels on Mars were first discerned as drainage features from Mariner 6 and 7 images. Since this discovery, the formational processes, timing, and duration have been topics of debate in the Mars community. Although not stable at the current Martian atmospheric pressure and temperature, valley networks in the southern highlands suggest that water was stable on the surface in the past. Valleys have been attributed to both fluvial and groundwater sapping processes. This study examines the geology, geomorphology, and basin morphometry of a portion of the southern highlands (17.5° - 27.5°S, 0° -10°W) to determine when these features formed and which process formed these valleys.

The geology of the study area includes a megaregolith, basin, and crater groups. The megaregolith is a combination of impact, volcanic, and eolian materials broadly emplaced during the Middle Noachian to Early Hesperian periods (~3.9 - ~3.4 Ga). Two megaregolith units, upper and lower, are preserved in the study area. The lower megaregolith (Nm₁) is a heavily dissected unit and preserves the majority of the valley networks in the study area. The upper megaregolith (HNm₂) contains ridges, interpreted as the surface expression of thrust faulting due to contraction, and overlies the lower
megaregolith, where present. The basin group includes an ancient fractured basin floor unit (Nb$_1$), preserved, post-impact, basin infill (Nb$_2$), more recent basin infill associated with the water transport (HNb$_3$), and the materials lying on the floors (ANb$_4$) of the majority of the craters (c$_1$, c$_2$, and c$_3$) in study area. Valley geomorphologies and deposits of materials from the mouths of the valley networks are preserved in units within the Late Noachian to Early Hesperian. Thus, valleys likely formed during the Late Noachian with activity waning into the Early Hesperian.

Basin morphometry and hypsometry were used to determine the process forming the valleys. Hypsometry examines the shape of a basin to determine which process(es) likely formed basins. This type of study has successfully predicted the processes forming basins on the Earth, and, to a limited extent, has been tested on Mars. For this study, I used the tributaries to the distal portions of the Escalante River, southeastern Utah as a proxy for the valleys forming on Mars. Using hypsometry, this study demonstrates that terrestrial basins with valleys similar to those on Mars fit the quantitative criteria of valleys formed by both groundwater-sapping and precipitation run-off processes. Although the hypsometric analysis predicts the dominant process that formed the valleys in the basins, it does not account for rock-type, bedding orientation, rheology, hydrologic properties, or the differences in physical characteristics of the body. Further research is suggested for constraining further the variables that produce both types of valleys.
ACKNOWLEDGEMENTS

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I want to acknowledge the Thursday afternoon science club members for their mental support while I finished this thesis. Thanks to Steve Rice and Doug Portis for constantly asking me to explain why I was studying Mars, and then making fun of me, no matter my answer. To my fellow graduate students, thanks for your commiseration over the years.

Finally, I would like to thank my family, although you have not always understood what I am doing with my life, you have always been supportive. I would not have been able to get to where I am without you. A special thanks to Ma and Pa Fortezzo for buying me my first telescope when I was just a little boy. I can trace the beginnings of my science career back to sitting in your bedroom, looking at the night sky through that telescope, and waiting patiently for stars to fall. Thank you.
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CHAPTER 1

INTRODUCTION

The study of the surface of Mars has advanced significantly since the Mariner and Viking missions of the 1960’s and 1970’s. With its launch in 1996, the Mars Global Surveyor (MGS) acquired the first new data since the Viking missions. After MGS, three additional satellites were successfully inserted into orbit around Mars: Mars Odyssey (ODY, launched in 2001), Mars Express (MEX, launch in 2003), and Mars Reconnaissance Orbiter (MRO, launched in 2005). These orbiting spacecraft provide high-resolution images, multi-spectral datasets, and topographic information, which allow for new investigations and provide the basis for reexamining results based on Viking data. These more detailed data increase the number of resolvable features on the surface, and thus increase the likelihood of discerning landforms that are analogous to those studied on Earth. Applying terrestrial analogs to Martian surface features provides an additional investigative tool with which planetary scientists can determine the process(es) that shape Martian landforms. This study uses these recent data and terrestrial analogs to resolve the geologic history, and particularly the role of water in producing an area of the dissected landscape of the southern highlands of Mars.

Importance of this study

The recent and ongoing accumulation of surface data for Mars allows for localized, detailed studies of the geologic processes that have altered the Martian surface. Of particular interest to planetary scientists is the role of water on the surface of Mars. What was the temporal extent of liquid water on Mars? What is the main source of water, melting ground-ice or precipitation? Is groundwater sapping or precipitation
runoff the key process forming the channels and valleys in the Martian highlands? Was the Martian fluvial activity long-lived or episodic?

This study adds to the knowledge about the role, temporal extent, formational (and possible “extinctional”) controls of fluvial processes in an area of Mars where the record of early impacts and fluvial processes are well preserved. While this study encompasses only a small portion of the Martian southern highlands (~350,000 km², ~0.2% of the total Martian surface area, slightly larger than the state of Arizona), it contains a diversity of features representative of those found throughout the Martian southern hemisphere (i.e., ancient, large, eroded impact basins, channels, valley networks, gullies, fan morphologies, linear surface structures, gravity-driven process morphologies, and aeolian features). Therefore, this study area provides not only a window into the regional fluvial history, but is a representative sample of the geologic processes affecting a large portion of the southern hemisphere throughout Martian history.

Area of Study

The southern hemisphere of Mars is an intensely cratered, fluvially dissected, high-standing terrain, which starkly contrasts with the low-lying, nearly horizontal northern plains (Figure 1.1). The highland surface rocks consist largely of fragmented crustal materials comminuted and ballistically emplaced through impact processes. The density of this “megaregolith” increases with increasing depth (Melosh, 1989), until it grades into fractured, probably igneous, bedrock beneath. “Regolith” is the surficial layer of soil, whereas “megaregolith” represents materials exposed to impact processes for long periods (Melosh, 1989). Volcanic, eolian, and ejecta blankets mantle the
megaregolith in places (Scott and Tanaka, 1986; Guest and Greeley, 1987). Valleys and channels of various sizes often incise these materials. The rugged southern hemisphere terrae (Figure 1.1, see Appendix A for definition of Latin descriptor terms) are the oldest exposures of surface materials, having the highest density of preserved craters (Guest and Greeley, 1987; Scott and Tanaka, 1986). Margaritifer Terra (Figure 1.1) also contains large depressions that are partly filled with mosaics of irregularly-shaped, scarp-bounded mesas. These features, termed “chaos” or “chaotic terrain”, are thought to have formed because of large-scale discharge of subsurface volatiles, likely water, and perhaps carbon dioxide, causing the surface to collapse. Fluvial processes in these regions led to a complex history of erosion, transport, and deposition of material from the uplands in the south to the northern lowland basin.

The study area (17.5ºS – 27.5ºS, 350ºE - 360ºE, Figure 1.2) includes portions of the Noachis and Arabia Terrae, and small portion of Margaritifer Terra on the western edge. Noachis basin, a ~350 km diameter, degraded, ancient, flat-floored impact structure dominates the map area. The study area slopes to the north-northwest following the regional trend created by the rim of Hellas basin. To the north of Noachis basin, a ~300 km wide, increased slope, diffusely marks the boundary between the upper- and lower-highlands and separates Noachis Terra and Arabia Terra (Figure 1.2). This boundary delineates a topographic change from the upper highlands of Noachis Terra to the lower highlands of Arabia Terra. The density of preserved fluvial and impact structures decreases within this sloping boundary. This boundary may represent the margins of the ejecta from the impact that formed Hellas basin. Intensely dissected, highland plains rise to form the plateau of Noachis Terra to the south of Noachis basin.
Figure 1.1: Color shaded-relief image of the Martian surface. The red line (dashed where buried by Tharsis) indicates the approximate location of the highland-lowland boundary (HLB) scarp that separates the southern highlands from the northern plains. The dashed blue line represents the diffuse boundary between the upper and lower highlands. The black box indicates the location of the study area, which includes portions of Noachis, Arabia, and Margaritifer Terrae. See Appendix A for description of the names. At the equator, $30^\circ$ is equal to $\approx 1,800$ km. Simple Cylindrical projection of Mars Orbiter Laser Altimeter (MOLA) data (3.7 km/pixel).
Figure 1.2: Color shaded-relief image of the study area. Black dashed line represents the approximate location of the diffuse boundary between the upper highlands of Noachis Terra and the lower highlands of Arabia Terra. Blue dashed line indicates the boundary between to the direction of Margaritifer Terra. Image is ~570 km across at 17.5ºS and ~530 km across at 27.5ºS, nearly the size of the state of Arizona. Simple-cylindrical projection of MOLA data (463 m/pixel).

Materials in Noachis Terra are representative of the oldest exposures on the surface of Mars. Large valleys are present on the outer flanks of Newcomb crater, located east of
Noachis basin, and small valleys and gullies incise the interior crater wall. To the west of Noachis basin, a north-south trending drainage divide separates Noachis Terra from a network of radial valleys at the eastern margin of Margaritifer Terra (Figure 1.2). These valley networks, named Paraná Valles, dissect the rugged highland terrain of Margaritifer Terra. The summary of previous work, below, provides a general geologic history.

**Objectives of study**

This study is a multi-task project intended to: (1) define geologic context and temporal extent of Martian fluvial processes by mapping discrete geologic/geomorphic units based on textural components, erosional processes, morphologies, remotely-sensed mineralogy and relative chronology using orbiting satellite imagery and spectrometer data; (2) calculate linear, areal, and relief measurements from Martian drainage basins within the study area; (3) calculate linear, areal, and relief parameters from analogous terrestrial valleys of the southwestern Colorado Plateau of the United States; and (4) discuss the results of the comparative study and the implications for Martian fluvial history.

**Objective 1: Define the temporal extent and geologic context of fluvial activity**

Defining geologic units, assigning relative ages, and determining which units preserve, mantle, or are unaffected by fluvial processes, creates the foundation for understanding a temporal extent and geologic context of fluvial activity in this area of Mars. Using a geographic information system (GIS) to delineate discrete geologic and stratigraphic units and perform spatial analyses allows the user to determine relative relationships both qualitatively and quantitatively.
A GIS simultaneously gathers graphic and tabular data into an analyzable spatial database. Vector mapping in a GIS allows characterization of geologic units and linear structures and stores dimensional attributes. The spatial database makes it easier to carry out detailed analyses of the mapped features and calculated attributes (e.g., crater dating, drainage basin analyses, dimensional analyses, etc.). In addition, the project GIS easily overlays multiple raster datasets with varying spatial resolution allowing for simultaneous viewing of image, topographic, and mineralogic information. This objective consumed a significant amount of time for image processing, which creates the image bases for mapping and interpretation.

Objective 2: Morphometric analysis of Martian and terrestrial drainage basins

This objective involved analyzing the relief and area of Martian basins within the study area to move beyond first-order qualitative comparisons with terrestrial analogs. The relationship between area and relief define the three-dimensional, spatial characteristics of the basins, and are used to delineate the dominate process in drainage basin formation on Earth. Calculations and ratios derived from these first-order measurements were used to build a numerical dataset for comparison with the terrestrial analog dataset from Objective 3.

Objective 3: Terrestrial study of the analogous valleys

A similar remote-based morphometry study of the distal tributaries to the Escalante River and the southern Escalante arm of Lake Powell in southern Utah provide the terrestrial analog data for comparison with the findings from Objective 2. This is a classic terrestrial sapping valley location with similar morphologies to those of the valleys in the study area (e.g., Laity and Malin, 1985). These valleys are presently active
and provide useful insights into how these morphologies could have developed on Mars. To understand more thoroughly the controls on the morphology of these tributaries, two field investigations to the Escalante River area studied the local geology, erosional processes, and fluvial dynamics. Additional measurements from tributaries of the Escalante River, where surface runoff is the dominate process were collected in order to further differentiate the formational processes.

**Objective 4: Synthesis of mapping and morphometric studies**

This objective combines the results from the previous three objectives in order to compare and contrast the physical parameters, geologic context, morphology, measured quantities, and calculated ratios. The focus of the discussion is to: (1) quantitatively examine the relationship between similar terrestrial and Martian basins and valleys and the formational processes; and (2) investigate quantitative outliers, those areas with dissimilar basin and valley morphometry, to determine reasons why these disparities exist. This synthesis also compares these results with the previous studies from similar areas of Mars and the Earth.

**Previous Work**

**Geologic Maps**

Previous studies used Mariner and Viking mission data to produce geologic maps of the study area at global and regional scales. Saunders (1979) produced a geologic map of a portion of the southern highlands (0° - 30°S, 0° - 45°W) at 1:5,000,000 scale using Mariner 9 data (1 -2 km/pix), which included the study area. The Saunders (1979) geologic map interpreted the study area as smooth materials and mountainous areas surrounded by ancient, cratered highlands.
Scott and Tanaka (1986) mapped the geology at 1:15,000,000 scale using global Viking image datasets (100 – 300 m/pix). They interpreted the area as the dissected, subdued crater, and smooth units of the highland plateau sequence. The dissected unit was interpreted as a fluvially dissected mélange of ejected material, pyroclastic material, and lavas. The subdued crater unit was interpreted as a thin mantle of eolian material and thin lavas where the underlying cratered topography remained distinguishable. The plains-forming, smooth member was interpreted to consist of a thicker mantle of eolian and volcanic material, which inundated regional surfaces and sequestered signatures of the underlying topography.

A Grant et al. (2009) 1:1,000,000 scale geologic map abuts the western edge of the study area. This larger-scale map indicated a heavily dissected area with various impact and eolian processes active on the ancient surface throughout its geologic history. This mapping study used a 100 m/pix image base and, where available, used sub-20 m/pixel images to further describe the textures of the units. Grant et al. (2009) suggest that the regional highland surface underwent severe impact and fluvial processes early in its history. This area, as well as the surrounding region, endured intense fluvial activity and were areas of intense erosion, transport, and deposition of material. As fluvial activity waned in the region, two widespread resurfacing events embayed, subdued, or inundated the area through volcanic and eolian processes. Since the end of fluvial activity, small impacts and eolian processes are the dominant resurfacing processes.

*Fluvial Studies*

The valley networks and channels on Mars were first discerned as drainage features from Mariner 6 and 7 images (e.g., Schultz and Ingerson, 1973). Better
resolution and global coverage during the Mariner 9 (McCauely et al., 1972; Masursky, 1973) and Viking Orbiter missions (Carr et al., 1976) revealed that valley networks occurred (and are preserved mainly within) the ancient, cratered, southern highlands (Pieri, 1979; Carr and Clow, 1981; Mars Channel Working Group, 1983). The Mars Channel Working Group (1983) defined valleys as “elongate troughs or systems of troughs that lack basal channel-related flow characteristics”. The designation as a “channel” is reserved for those valleys showing flow characteristics (i.e., incised troughs, bars, lee deposits around obstructions). The overall morphology of a valley system offers insight into the role and persistence of water on the early Martian surface. As such, previous studies sought to quantitatively delineate their extent, characterize their morphometry, and constrain their origins (e.g., Carr, 1979; Pieri, 1980; Carr and Clow, 1981; Baker and Partridge, 1986; Grant 1987, 2000; Williams and Phillips, 2001; Grant and Parker, 2002).

Recent morphometric and hypsometric (the ratio of basin-relief to basin-area, suggestive of the formational process and maturity) studies of basins on Mars (Luo, 2002; Grant and Parker, 2002; Grant and Fortezzo, 2003; Fortezzo and Grant, 2004) suggested a combination of fluvial and sapping processes likely contributed to Martian valley formation. Grant and Fortezzo (2003) and Fortezzo and Grant (2004) highlighted the importance of impact cratering in shaping basin hypsometry. On Earth, fluvial processes modify topography derived mostly from tectonic processes. In contrast, impact cratering creates initial basin topographic relief on Mars (Baker, 1988), which almost immediately yield basin morphometries similar to those of highly-modified, tectonically-initiated, terrestrial basins (Luo, 2000 and 2002; Grant and Fortezzo, 2006). The impact structures
may yield surface area/elevation ratios that are “pre-conditioned” to give a hypsometric signature similar to those for terrestrial fluvial environments (Grant and Fortezzo, 2003). Grant and Fortezzo (2006) provided further evidence for this “pre-conditioning” of the Martian landscape by demonstrating that the outer flanks of Martian impact structures have longitudinal profiles very similar to those of equilibrated terrestrial river basins.

Although Martian valleys and channels have been studied for more than 30 years (e.g., Sharp and Malin, 1975; Pieri, 1980; Carr 1981, 1996; Carr and Clow, 1981; Baker, 1982; Malin and Carr, 1999; Grant, 2000; Craddock and Howard, 2002; Grant and Parker, 2002; etc.), the formational mechanisms and water source(s) remain ambiguous. For example, is the source of water atmospheric or subsurface? If the source is precipitation, is surface-runoff or groundwater sapping forming these morphologies? If ground-ice is the source, what mechanisms are causing the melt of this groundmass? The Margaritifer Terra and Noachis Terra regions (Figure 1.1 & 1.2) are of particular interest for valley formation studies, because these areas include some of the highest drainage densities (valley length per basin area) still preserved on Mars (Saunders, 1979; Carr and Clow, 1981; Grant, 2000; Grant and Parker, 2002).
CHAPTER 2

MARS OVERVIEW

This section summarizes over three decades of Martian studies through a presentation of (1) post-Viking mission orbital instrument results used in this study, (2) Mars’ physical parameters and terrestrial comparison, (3) Mars’ global physiographic characteristics, (4) the Martian stratigraphic scheme, and (5) the summary geologic history of Mars. Although this chapter only provides very cursory presentations of the geologic and evolutionary character of Mars, extensive reference materials are provided.

Post-Viking Era Mars Exploration

Following the end of the Viking missions in 1980, there was a sixteen-year hiatus before the next successful mission to Mars. The Mars Global Surveyor (MGS) launched in November 1996 and carried the Mars Orbiter Camera (MOC) imaging instrument, the Mars Orbiter Laser Altimeter (MOLA) ranging instrument, and the Thermal Emission Spectrometer (TES). The MOC instrument contained two cameras: (1) a wide-angle (WA) camera for relatively low-resolution (2 – 0.250 km/pix), regional perspectives, and (2) a narrow angle (NA) camera for relatively high-resolution (~1.4 – 11 m/pix), local perspectives (Albee, et al., 2001). The MOLA instrument surveyed the surface and provided detailed topography with a 160 m diameter laser shot every 300 m on the surface (Albee, et al., 2001). The TES instrument mapped the coarse mineralogy of the entire Martian surface using bands between 6 and 50 μm (Albee, et al., 2001). The resulting mineral maps have a resolution of 3x2 km/pix. The MGS mission ended when communications were lost in 2007. Results included a global topography, coarse crustal

In 2001, NASA launched the Mars Odyssey (ODY) mission and it (as of this writing) continues to collect data. ODY carries the Thermal Imaging System (THEMIS) instrument, which collects thermal infrared (IR) images during the daytime and nighttime related to the surface temperature, as well as higher-resolution visible-range (VIS) spectra during the daytime (Christensen, et al., 2004). Comparisons of daytime and nighttime IR images provide important information about the thermophysical characteristics of the Martian surface and near subsurface (<1 m), including thermal inertia (resistance to temperature changes) and reflectivity (normal albedo) (Christensen, et al., 2004).

NASA launched the Mars Reconnaissance Orbiter (MRO) mission in 2005, and (as of this writing) continues to collect data. MRO carries the High-Resolution Imaging Science Experiment (HiRISE) camera, the Context Camera (CTX), and the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) multi-spectral scanner. The HiRISE instrument is collecting the highest resolution (25 cm/pix) visible range images ever from the Martian surface (McEwen, et al., 2007). The CTX instrument is a large-area, high-resolution imager that provides context for the HiRISE and MOC images (Malin, et al., 2007). Data from the CRISM instrument provide detailed information about the mineralogy of the exposed crust and mantling deposits (Murchie, et al., 2007). The CRISM instrument collects data with varying resolutions, over a number of spectra to provide both regional and local mineralogy.

The European Space Agency launched the Mars Express (MEx) mission in 2004 (and as of this writing) it is still collecting data. MEx carries the German High-
resolution stereo camera (hsrC) imaging instrument, the French observatoire pour la
mineralogie, l’eau, les Glaces et l’Activite (OMEGA) visible and infrared mineralogical
mapping spectrometer, and the Italian Mars advanced radar for subsurface and
ionosphere sounding (MARSIS) multi-frequency synthetic radar altimeter with ground
penetration abilities. Currently there are no HRSC images or MARSIS radar data
available in the study area. There is OMEGA data within the map area but was not used
because of convoluted processing, and the availability of other mineralogic datasets.

These recent missions flew instruments helpful in the task of unraveling the
history of Mars and, more specifically, pursuing the NASA directive to “follow the
in an effort to determine the origins for life on Earth and if life could exist, or did exist,
on other planets. The data from these instruments provide a plethora of new information
that is increasing and refining our understanding about the physical characteristics, global
physiography, geologic time, and the geologic history of Mars.

Physical Characteristics

Mars and Earth are generally markedly different in their orbital, atmospheric, and
physical characteristics. To a large degree, these contrasts dictate how geologic
processes affect the planets. The following section summarizes these characteristics with
specific focus on the differences (Table 2.1).

Mars travels at an average distance ~1.5 times farther from the sun than the Earth.
Although believed to have changed through time (Costard, et al., 2002), Mars currently
tilts on its axis at ~25°, slightly greater than that of Earth, giving Mars similar cyclic
seasons. The orbital period for Mars is 687 days, so Martian seasons last almost twice as
long as terrestrial seasons. The greater ellipticity of the Martian orbit around the sun affects the length of hemispheric seasons by decreasing the speed of the orbit at aphelion and increasing the orbital speed near perihelion (Pasachoff, 1998).

Table 2.1: Physical Characteristics of Mars and Earth.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mars</th>
<th>Earth</th>
<th>Ratio (Mars/Earth)</th>
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<tbody>
<tr>
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<td>Avg. Distance from the Sun (AU)</td>
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<td>1.52</td>
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<td>Orbital Period (days)</td>
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<td>365.25</td>
<td>1.88</td>
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<tr>
<td>Length of day (seconds)</td>
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<td>86400</td>
<td>1.03</td>
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<tr>
<td>Obliquity (degrees)</td>
<td>25.19</td>
<td>23.45</td>
<td>1.07</td>
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<tr>
<td><strong>Atmospheric</strong></td>
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<tr>
<td>Avg. Pressure (Mean Radius, mb)</td>
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<td>0.01</td>
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<tr>
<td>Avg. Density (Surface, kg/m3)</td>
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<td>1.217</td>
<td>0.02</td>
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<tr>
<td><strong>Surficial</strong></td>
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<td></td>
</tr>
<tr>
<td>Mass ($10^{24}$ kg)</td>
<td>0.64185</td>
<td>5.9736</td>
<td>0.11</td>
</tr>
<tr>
<td>Radius (km)</td>
<td>3397</td>
<td>6371</td>
<td>0.53</td>
</tr>
<tr>
<td>Volume ($m^3$)</td>
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<td>10.8321</td>
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</tr>
<tr>
<td>Density ($g/cm^3$)</td>
<td>3.9335</td>
<td>5.515</td>
<td>0.71</td>
</tr>
<tr>
<td>Total Surface Area ($10^{14} m^2$)</td>
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<td>1.4796</td>
<td>0.98</td>
</tr>
<tr>
<td>Mean Temperature (Daytime, °K)</td>
<td>242</td>
<td>293</td>
<td>0.83</td>
</tr>
<tr>
<td>Mean Temperature (Nighttime °K)</td>
<td>184</td>
<td>283</td>
<td>0.65</td>
</tr>
</tbody>
</table>


Mars has ~10% and ~15% of Earth’s mass and volume, respectively and has a mean radius about half that of the Earth. The density of Mars is 3.934 g/cm$^3$ compared to 5.515 g/cm$^3$ of Earth. This smaller size gives Mars a lower acceleration of gravity, ~3.7 m/s$^2$, compared to the Earth’s 9.8 m/s$^2$. The Martian surface area is roughly equivalent to the area of exposed continental crust of Earth (Kieffer, et al, 1992). The average daily temperature fluctuates ~58ºK diurnally, although several authors (e.g., Carr, 1996; Craddock and Howard, 1999) hypothesize that the minimum and maximum temperatures
have varied through time due to changes in solar radiation, atmospheric density, and obliquity.

The atmosphere of Mars is ~95% carbon dioxide, with nitrogen, argon, oxygen, carbon monoxide, water, neon, krypton, xenon, and ozone, decreasing from ~3% to 0.04 ppm, respectively (Kieffer, et al., 1992). The average surface pressure is 6.36 mbar but scales inversely with elevation; lower elevations have higher densities and higher elevations have lower densities. Martian surface pressure is markedely different from the 1014 mbar of Earth and it varies much more on Mars. The atmospheric density is ~1.6% of Earth’s. This thinner atmosphere at the surface produces wind speeds with ~1/9th of the density of those on Earth. As such, a 160 km/h (100 mph) wind on Mars would feel like a ~18 km/h (~11 mph) breeze on Earth.

Although not currently stable, liquid water was likely stable in the Martian past based on both physical characteristics (e.g., Carr, 1996; Craddock and Howard, 2002) and geomorphologic indicators (e.g., Masson, et al., 2001). At average temperatures and pressures (Figure 2.1), water cannot currently exist in a liquid phase (Carr, 1996). However, under current atmospheric pressures and surface temperatures, water can exist in the subsurface, or on the surface in the polar regions, as a solid, or as a gas in the atmosphere (Carr, 1996). The geomorphic evidence of erosion by liquid water on the surface suggests that Mars had a sufficiently warmer and denser atmosphere in the past to support liquid water (e.g., Carr, 1996; Craddock and Howard, 2002).

Global Physiography

Before discussing the topography of the planet, it is important to note the derivation of elevation values. The mean radius of Mars is 3397 km, from the inferred
spherical center of the planet to the surface. This value is also the zero elevation point.

The total relief on Mars is >29,000 m, with a low of -8,177 m in Hellas.

![Figure 2.1: Water and carbon dioxide phase diagram showing that under typical Martian surface conditions (black circle) water is stable in solid form. Red box indicates the current range of temperatures and pressures on the Martian surface. Note that the majority of the range is outside of the liquid phase. Areas suitable for liquid water would be lower-elevation, high atmospheric pressure, high temperature areas, likely in the low latitudes during the summer. Additionally, temperature is the driving force driving phase change. Liquid CO\textsubscript{2} would be stable on the surface at 5 bars with the current Martian temperature range (Hoffman, 2000). Adapted from Carr (1996).](image)

using the MOLA dataset. This allows for negative and positive values above and below the Planitia and a high of 21,171 m atop Olympus Mons.

Mars contains two major physiographic provinces: the northern lowlands and the southern highlands (Figure 1.1). The highland-lowland boundary (HLB), or dichotomy boundary, separates the two regions. The HLB is an east-west trending scarp that
encircles the globe between ~15°S and ~50°N except where partially buried by the Tharsis rise (Figure 1.1). The northern lowlands cover most of the northern hemisphere. Elysium rise is the only large volcanic province surrounded by lowlands. The Borealis basin surrounds Planum Boreum, the north polar plateau of Mars. Planum Boreum is a scarp-bounded plateau that rises above the surrounding plains, and has spiral troughs, large reentrants, gentle undulations, steep scarps, and marginal convex slopes (Thomas, et al., 1992). On the opposite side of the planet, Planum Australe, the Martian south polar region, has a similar type of topography (Thomas, et al., 1992).

South of the HLB, the terrain can be divided into four major types: highland, basin, volcanic, and canyon. The highlands include the topographically rugged and densely cratered Noachis Terra, Arabia Terra, Terra Cimmeria, and Terra Sirenum. These terrains are the highest, non-volcano elevations on the planet. Argyre and Hellas Planitiae are two of the largest and oldest, well-preserved impact basins the region and include some of the lowest elevations on the Martian surface. The floor of Hellas Planitia is the lowest surface elevation on Mars. In contrast, the Tharsis rise volcanic province includes the highest elevation (the summit of Olympus Mons). The Tharsis rise is the largest volcanic construct on the planet. To the south of Tharsis, the Valles Marineris canyon system trends west-northwest to east-southeast (Lucchitta, et al., 1992). This rift system opens on its eastern end into large, block-filled depressions (termed “chaos” or “chaotic terrain”) with floors dissected by braided and meandering, steep-walled, wide valleys.

The thickness of the Martian crust varies over the entire planet, ranging from 3 km within Hellas Planitia to a threshold of ~90 km thick at the Tharsis rise with an
average value of ~50 km (Zuber, 2000 and 2001). The crustal thickness models for Mars use both topography and gravity data, but lack seismic velocity data. The Earth’s and the Moon’s crustal thickness models are tied to seismic data giving them a subsurface robustness that the Mars models lack (Zuber, 2001). The models dependence on topography causes it to vary directly with changes in elevation. The crustal thickness models indicate that there is a true global dichotomy between the northern plains and southern highlands, not just an erosional surface.

**Geologic Time Scale**

As in terrestrial geology, there is a need to describe Martian geology relative to time. On Earth, the geologic time scale combines superposition, fossil records, detailed stratigraphic correlations, and radiometric techniques. For the Moon, a combination of geologic relationships, crater densities, and radiogenic ages of returned samples were combined to produce its time scale.

The principle of superposition (Monroe and Wicander, 1998), where a younger unit cuts, overlies, or overlaps older materials is the fundamental basis for relating geologic units. On the Moon, unit delineations based on superposition indicated that older units always display a higher density of craters than that of younger units (Stöffler and Ryder, 2001). Thus, in the absence of superposition relationships or where units are not in contact with one another, a study of the crater size-frequency distributions (SFD) could determine relative ages (Stöffler and Ryder, 2001).

To apply SFD’s to determine surface ages, a scale of impact rates with respect to crater diameter and geologic time was developed. Since the end of the period of heavy bombardment (discussed below), the frequency of impacts and the diameter of the
structures have both decreased. The crater production function estimates the original SFD (Ivanov, 2001). Essentially, a predictable number of craters should exist on a surface of a certain age, as long as their size is sufficient for retention. Various processes can preferentially destroy smaller craters. The precision and accuracy of the lunar crater production function was validated when it accurately predicted the absolute ages of samples returned by the Apollo and Luna missions (Neukum et al., 2001). The addition of radiometric dates for samples from known locations correlates the lunar production function to absolute time. An adaptation of the lunar production function for use on Mars took into consideration the unique Martian orbital, physical, and atmospheric parameters, allowing a way to scale crater SFD’s quantitatively (Ivanov, 2001). The lunar production function fails to account for episodic large impacts and this can alter results.

A SFD determines the placement of the unit within the timeline of the production function. The number of craters, their associated diameters, and the area of the units are tabulated, and the areal densities of craters are then normalized to 1 or $1\times10^6$ km$^2$. This yields the number of craters per million km$^2$ and provides a correlation to “type locality” surfaces within the relative time scale (e.g., Hartmann and Neukum, 2001). Two types of crater counts, cumulative and incremental (binned), are used to develop a relative time scale. Herein, this study presents only the cumulative crater count method. This method counts the number of craters larger than a certain diameter, reported as $N(X)$, where $X$ is the chosen diameter. For example, a selected geologic unit of 100,000 km$^2$ that has 5 craters greater than 5 km in diameter would have a crater density represented by $N(5)=50 \pm Q$, where $Q$ is the error reported as the square root of number of craters divided by the area normalized to $1\times10^6$. The number of craters of a certain diameter per km$^2$ (Moon) or
10^6 km^2 (Mars) led to the division of time scales based on areas of significantly different counts.

Scott and Carr (1978) introduced the Martian divisions of the geologic time scale, partitioning it into three geochronologic periods: Noachian, Hesperian and Amazonian, oldest to youngest, respectively. The relative ages of map units in Noachis Terra (Noachian), Hesperia Planum (Hesperian) and Amazonia Planitia (Amazonian) were the basis of these divisions. Noachis Terra (Figure 2.2a) is an intensely cratered highland region in the southern hemisphere. Hesperian Planum (Figure 2.2b) is a ridged, smooth, moderately cratered region near Terra Tyrhenna and the Tyrhenna Patera volcanic edifice. Amazonia Planitia (Figure 2.2c) is a low-lying, regionally horizontal, sparsely cratered region of the northern plains. Tanaka (1986) further subdivided the periods into eight epochs based on geographic regions and N(2), N(5), and N(16) cumulative crater statistics, and tied the relative time scale to published absolute age models (Table 2.2). The scheme was applied to subsequent geologic mapping and other topical studies of Mars (e.g., Scott and Tanaka, 1986)

Craters with diameters ≥16 km define the Early to Late Noachian (Tanaka, 1986). For the Late Noachian, Hesperian and Early Amazonian, the best record is within the N(5) range (Tanaka, 1986). Craters with diameters between 1 and 2 km are the most useful in distinguishing the Late Hesperian and Amazonian surface ages (Tanaka, 1986). Unlike the Moon, samples from known locales on Mars do not exist, thus hindering the application of absolute ages to geologic units. Hartmann and Neukum (2001) fit the crater statistic ages of Tanaka (1986) to a model of absolute time using improved production functions (Table 2.2).
Figure 2.2: Colored shaded-relief images of the three regions used to divide the Martian geologic time scale based on crater densities. A) A portion of Noachis Terra showing densely packed craters ranging in size from ~200 km to ~5 km in diameter. B) Hesperia Planum (central portion of the figure) ridged volcanic plains with fewer craters than Noachis Terra and no craters >100 km in diameter. The smooth plains of Amazonia Planitia (C) display a paucity of craters. The color scale is applicable for all panels. Simple Cylindrical projection of Mars Orbiter Laser Altimeter (MOLA) data (3.7 km/pixel).
Table 2.2: Crater SFD’s and absolute ages for Martian time scale boundaries.

<table>
<thead>
<tr>
<th>Epoch</th>
<th>N(2)</th>
<th>N(5)</th>
<th>N(16)</th>
<th>Absolute Ages$^2$ (Ga)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Late Amazonian</td>
<td>&lt;40</td>
<td>&lt;0.6</td>
<td>-0.3</td>
<td>$&lt;0.6 - 0.3$</td>
</tr>
<tr>
<td>Middle Amazonian</td>
<td>40-150</td>
<td>&lt;25</td>
<td></td>
<td>2.1 to 1.4 - 0.6 to 0.3</td>
</tr>
<tr>
<td>Early Amazonian</td>
<td>150-400</td>
<td>25-67</td>
<td></td>
<td>3.3 or 2.9 - 2.1 to 1.4</td>
</tr>
<tr>
<td>Late Hesperian</td>
<td>400-750</td>
<td>67-125</td>
<td></td>
<td>3.6 - 3.3 or 2.9</td>
</tr>
<tr>
<td>Early Hesperian</td>
<td>750-1200</td>
<td>125-200</td>
<td>&lt;25</td>
<td>3.7 - 3.6</td>
</tr>
<tr>
<td>Late Noachian</td>
<td>200-400</td>
<td>25-100</td>
<td></td>
<td>3.82 - 3.7</td>
</tr>
<tr>
<td>Middle Noachian</td>
<td>&gt;400</td>
<td>100-200</td>
<td></td>
<td>3.93 - 3.82</td>
</tr>
<tr>
<td>Early Noachian</td>
<td>&gt;200</td>
<td></td>
<td></td>
<td>&gt;3.93</td>
</tr>
</tbody>
</table>

$^1$Modified from Tanaka, 1986; $^2$Hartmann and Neukum, 2001

Hartmann and Neukum (2001) identified several notable limitations to the crater statistics method: 1) fitting SFD’s to the production function using either the least squares or polynomial method have only ~10% accuracy rate, 2) the extent of certain geologic units is subject to interpretation, thus multiple counts may exist, 3) removal of craters from a surface through erosion or burial may produce a younger age, 4) as data resolution increases, so does the ability to define smaller geologic units. The large diameter craters used to define the oldest ages on Mars may not exist on these smaller surfaces, also skewing the interpretation towards a younger age. A pitfall with using smaller craters (typically <100 meters in diameter) for SFD’s is the presence of secondary craters. Secondary craters form when an initial impactor ejects competent materials that create smaller craters radial or circumferential to the original structure. These impacts are sometimes difficult to separate from original small impacts.

An examination of the global distribution of material ages reveals a relationship between the ages of materials and physiography. The southern high-elevation region is predominately Noachian to Hesperian age materials with small areas of Hesperian to Amazonian age volcanic materials (Figure 2.3). The northern low-lying region is mostly
Figure 2.3: Global ages of locally dominant materials. Yellow hash line represents the dichotomy boundary. L = Late, M = Middle, E = Early, A = Amazonian, H = Hesperian, and N = Noachian. Molleweide Projection with an MDIM2.1 base map. Modified from Nimmo and Tanaka (2005).
Hesperian to Amazonian with older materials cropping out near the HLB (Figure 2.3). Craters and small domes, mostly occurring on Noachian age materials, remain poorly constrained (Tanaka et al., 1988).

**Geologic History**

Using geologic characteristics of the Martian surface, crosscutting relationships, crater densities, and a working knowledge of planetary evolution, we can piece together the broad-scale evolution of Mars. Most of the information discussed in the following section is summarized from papers whose topics include early crustal formation (e.g., Nimmo and Tanaka, 2005; Frey et al., 2002), process history (e.g., Tanaka et al. 1988; Head et al. 2001), and stratigraphy (Tanaka, 1986).

The geologic history of Mars includes intense cratering and long-lived volcanic, fluvial, periglacial, and eolian resurfacing (Figure 2.4). The period of “late heavy bombardment” occurred between ~4.1 and ~3.85 Ga, and affected the terrestrial planets. The Earth underwent the same amount of cratering as the rest of the terrestrial planets, but with its active tectonics and consistent erosional processes, the record is not as prevalent as those of Mars and the Moon. Heavy bombardment was unrivaled in the rate and the size of impactors, and the largest Martian impacts occurred late in the barrage. The impact events of early Mars shaped the surface leaving behind large basins and mountainous regions from eroded impact structures. Alteration of a surface by impact processes created megaregolith, a zone of impact ejecta with volcanic and sedimentary deposits that overlies fractured bedrock throughout the southern highlands (MacKinnon and Tanaka, 1989, Squyres, et al., 1992). Volcanic events changed the shape of the Martian surface by forming large rises, edifices, numerous smaller constructs, and by
Figure 2.4: Resurfacing processes on the surface of Mars. The browns correspond to the volcanic and impact processes that dominated the resurfacing of the planet. The greens are the volcanic and eolian processes, whereas the blues represent the fluvial and periglacial processes. Adapted from Tanaka, et al. (1988) using the geologic units from Scott and Tanaka (1986) and Guest and Greeley (1987). Mollweide projection with an MDIM2.1 base.
resurfacing large areas of the planet’s surface. Dense valley networks and massive outflow channels substantiate that fluvial processes were active throughout early geologic time, although gullied crater walls suggest recent fluvial activity. Fluvial processes include incision of single valleys or networks of valleys, and erosion and transport over long distances within flood-initiated outflow channels. Patterned ground, creep-like landforms, and knobby and etched terrains are all attributed to periglacial processes. Eolian processes dominate the current erosional system on Mars (Greeley et al., 1992 and 2001). Similar to Earth, wind moves material by suspension, saltation, and creep (Greeley, et al., 1992 and 2001). These processes form sand sheets, dunes, ridges, yardangs (wind sculpted ridges), and wind streaks (Wilson and Zimbleman, 2004). Often, these processes acted together to produce the current observable surface.

The following is a brief synopsis of the geologic history framed within the three major time divisions: Noachian, Hesperian, and Amazonian Periods.

Noachian Period

The crust of Mars formed during the Early Noachian Epoch (>3.93 Ga, \(N(16) \geq 200\)), at ~4.5 Ga. This time overlaps with the period of heavy bombardment when large bolides impacted the Martian surface, forming Hellas, Argyre, Isidis, Utopia, and Chryse basins. The HLB formed at roughly the same time as the internal dynamo shut down. The HLB formation remains enigmatic, and both exogenic (several large impacts) and endogenic (i.e., subduction and convection) processes have been proposed (Head, et al., 2001). The formation of the HLB during this period suggests that the topographic disparity between north and south also formed during this epoch. Volcanic and minor periglacial processes resurfaced small areas of Mars (Tanaka et al., 1988).
During the Middle Noachian Epoch (3.93 – 3.82 Ga, [100<N(16)<200]), intense bombardment continued, which furthered the evolution of the cratered highlands. Major fracture systems, termed fossae (see Appendix A), located in the southern hemisphere began to form during this time. Volcanic and impact resurfacing rates peaked, periglacial activity remained static, and fluvial and eolian processes began (Tanaka, et al., 1988).

Deposition of embaying materials decreased in the cratered highlands as highland volcanism and impact processes began to wane during the Late Noachian Epoch (3.82 – 3.7 Ga, [25<N(16)<100 or 200<N(5)<400]). Eolian processes intensified, periglacial processes remained steady, and dense valley networks formed as small-scale fluvial processes peaked.

**Hesperian Period**

Extensive lowland and highland areas, including the south polar region (Tanaka and Kolb, 2001), were resurfaced by volcanic ridged-plains during the Early Hesperian Epoch (3.7 – 3.6 Ga, [125<N(5)<200]). The ridges are thought to be either compressional features or remnants of volcanic edifices (Scott and Tanaka, 1986). Rifting at Valles Marineris (probably initiated in the Late Noachian) and fracturing radial to the Tharsis rise continued. Plains materials of Dorsa Argentea in the south polar region were deposited (Tanaka and Kolb, 2001). Fluvial, eolian, and impact processes decreased as volcanic resurfacing increased, and periglacial processes remained steady and minor (Tanaka, 1986).

The northern plains were resurfaced by volcanic, eolian, and fluvial materials (Tanaka, et al., 2005) during the Late Hesperian Epoch (3.6 – 3.3 or 2.9 Ga, [67<N(5)<125 or 400<N(2)<750]). Fracturing continued radial to the Tharsis rise, and
layered materials, likely volcanic are deposited in Valles Marineris. Fluvial processes peaked during this time due to the onset of large-scale evacuation of volatiles from the subsurface forming the chaotic terrain and outflow channels. Supposedly, eolian and volcanic processes decreased, periglacial processes peaked, and impact processes leveled off during this epoch (Neukum, et al., 2001).

Amazonian Period

The Amazonian period is marked by the deposition of the Vastitas Borealis units in the northern plains (Tanaka et al., 2008). During the Early Amazonian (3.3 or 2.9 – 2.1 to 1.4 Ga, [150<N(2)<400]), volcanic activity increased near the Elysium rise and on the Tharsis rise, with fracturing near both regions. The largest volcano in the Solar System, Olympus Mons, was growing during this time and gained most of its volume by the end of the epoch. The floors of Valles Marineris were being mantled, and headward erosion created large side canyons. Basal material of the north polar plateau and the surrounding plains were formed during this period (Tanaka et al., 2008). Volcanic and eolian processes increased, fluvial and periglacial diminished, and impact decreased slightly (Tanaka, 1988).

During the Middle Amazonian (2.1 to 1.4 – 0.6 to 0.3 Ga, [40<N(2)<150]), volcanism was restricted to the Tharsis and Elysium rises, and along the HLB between the two constructs. Additionally, periglacial or pyroclastic materials were deposited along the dichotomy boundary. Mass wasting deposits and eolian processes increased during this epoch, fluvial and volcanic activity declined, and impact processes remained steadily low.
During the Late Amazonian (0.6 to 0.3 Ga – Present, [N(2)<40]), volcanic activity and fracturing occurred on the Tharsis rise and Olympus Mons. The north south polar regions were covered in layered deposits of ice and dust (Kolb and Tanaka, 2001; Tanaka, et al., 2008). Both poles are subsequently covered by residual water-ice and seasonal CO$_2$ frost. Fluvial processes continue to form valleys near Elysium rise, and form gullies on steep slopes. Volcanic, impact, tectonic, and periglacial processes all decreased during this most recent epoch. Eolian and mass-wasting processes increased and are currently the most active erosional processes on Mars (Greeley, et al., 2001).
CHAPTER 3

METHODOLOGY FOR MAPPING

The study presented herein used multiple, remotely-sensed datasets in an attempt to constrain the geologic history within the study area. This chapter includes a detailed presentation of the software, datasets, and processing techniques used in the mapping portion of this study. The software programs were used for processing, mapping, and analysis of the datasets. The datasets were used to interpret the geologic history of the study area. This chapter includes a detailed description of the processing techniques used to get an image from its original raw format into a map projected and GIS-compatible format. These processes are documented in detail in the interest of repeatability, as well as to provide a single source for the methodologies.

Software Programs

This study utilizes a wide range of both proprietary software and Free Open Source Software (FOSS) programs to complete the image processing, mosaicking, mapping, and graphing. The descriptions of the software programs and the websites or companies that maintain the software are listed below:

- Environmental Systems Research Institute’s ArcGIS (v. 9.2, ©1996-2006, Redlands, CA) is a geographic information system (GIS) which includes the ArcMap, and ArcCatalog applications. The functionality of these applications was also extended using the 3D analyst and Spatial Analyst add-on extensions. ArcMap is a graphical user interface (GUI) that allows for vector mapping and raster display in projected space, graphic and tabular spatial data analysis, and cartographic production. ArcCatalog is a file management program specific to the file-types created for GIS
software, which allows the user to create, manage, view, and manipulate vector and raster data outside of map-projected space. The 3D Analyst extension for ArcMap creates shaded relief, slope, and contour maps. This extension contains the ArcScene application, which is a three-dimensional data viewer allowing the user to examine raster and vector data with respect to elevation data. The Spatial Analyst extension for ArcMap allows for raster-based, basin analysis (e.g., flow direction, watershed, stream order, etc.), and vector- and raster-based spatial statistics (e.g., nearest neighbor, clustering, etc.). Both ArcMap and ArcCatalog include a suite of analysis or geoprocessing tools called ArcToolbox. The utility of ArcToolbox is in its ability to run programs with specialized functions, i.e., raster conversion, projection scripts.

- Global Mapper desktop GIS from Global Mapper Software, LLC, (v. 9.03, © 2002 - 2008, Olathe, KS) is a desktop GIS that handles large image datasets and creates mosaics. Global Mapper also handles the manipulation and viewing of large topographic datasets. The advantage to Global Mapper over ESRI’s software is its ability to quickly view and mosaic large raster datasets.

- United States Geological Survey’s (USGS) Integrated Software for Imagers and Spectrometers (ISIS) is a FOSS image processing software package (URL: http://isis.astrogeology.usgs.gov). This program is engineered specifically to process spacecraft data for planetary missions. Nearly all of the datasets used in this study undergo some amount of ISIS processing to prepare them for the GIS environment.

- Geospatial Data Abstraction Library (GDAL; URL: http://www.gdal.org) FOSS is a suite of applications to help convert between different geospatial raster formats. It also has the capability to reproject raster datasets to many map projections.
• International Telephone & Telegraph’s (ITT) Interactive Data Language (IDL, v. 6.3, ®, Boulder, CO) and the IDL based Environment for Visualizing Images (ENVI, v. 4.3, ®, Boulder, CO) image processing software was used for multi-spectral image processing, band mixing, mosaicking, and viewing.

• Golden Software’s Surfer (v. 7, ®1993-1999, Golden, CO) data visualization program and Grapher (v. 5, ®1992-2004, Golden, CO) graphing program are used for topographic data visualization and crater statistic plotting. Surfer 7 allows for the concatenation, calculation, and analysis of large sets of X-, Y-, and Z-space data points. Grapher 5 was used to compile and plot crater statistics.

• Adobe’s Photoshop (v. 8.0, ®1990-2003) is an image manipulation program used to recalibrate the pixel values within the image. This software is also used to remove portions of the data where errors exist.

Datasets

The image data for Mars research are typically available online and hosted at NASA’s Planetary Data System (PDS, URL: http://pds.nasa.gov) archive. The PDS is an established standard for the file structure for planetary data, and the PDS products include universal header files that contain the image information (i.e., cell size, projection, solar azimuth, spacecraft orientation, etc.) and a raw (unprocessed) raster image. The raster format is a series of squares (picture elements or pixels), arranged in columns and rows, with one value per pixel. The range of pixel values, or the digital number (DN), varies between datasets and the total number of allowable values is dependent on the bit value. Bit values are typically 32-, 16-, or 8-bit, which corresponds to the maximum sum of DN values. The number of bits corresponds to the power to which the number 2 is raised,
i.e., 8-bit (2\(^8\)=256 DN values). Excluding the 32-bit (~4.3x10\(^9\) possible DN values) CRISM multi-spectral data, and 16-bit MOLA topographic data, the images used for this study are 8-bit.

The mapping portion of this project included interpretation from topographic, image, and spectrometer datasets. The Mars Global Surveyor’s (MGS) Mars Orbiter Laser Altimeter (MOLA) instrument produced the elevation points, which were gridded into a ~463 m/pixel, digital elevation model (DEM, Figures 1.1 and 1.2, Table 3.1). These data provided the information to produce slope and aspect raster datasets, and a vector contour map. The DEM is also useful for describing the texture and physiography, and performing quantitative analyses. The global coverage of MOLA provided regional context when coupled with global image datasets.

Table 3.1: Martian topographic, image, and spectrometer datasets used in this study.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Res. (m/pix)</th>
<th>λ (μm)</th>
<th>Band</th>
<th>Figures</th>
</tr>
</thead>
<tbody>
<tr>
<td>MOLA(^a)</td>
<td>463</td>
<td>N/A</td>
<td>N/A</td>
<td>1.1, 1.2</td>
</tr>
<tr>
<td>MDIM2.1(^b,c)</td>
<td>231</td>
<td>0.450 - 0.590</td>
<td>Visible</td>
<td>2.3, 2.4</td>
</tr>
<tr>
<td>THEMIS daytime IR(^c)</td>
<td>100</td>
<td>6.78 - 14.88</td>
<td>Thermal IR</td>
<td>3.1</td>
</tr>
<tr>
<td>THEMIS nighttime IR(^c)</td>
<td>100</td>
<td>6.78 - 14.89</td>
<td>Thermal IR</td>
<td>3.1</td>
</tr>
<tr>
<td>THEMIS DCS(^d)</td>
<td>100</td>
<td>6.78 - 12.57</td>
<td>Thermal IR</td>
<td>3.3</td>
</tr>
<tr>
<td>CRISM(^e)</td>
<td>160 &amp; 85</td>
<td>0.37 3.92</td>
<td>Visible, Near IR</td>
<td>3.2</td>
</tr>
<tr>
<td>THEMIS Visible(^c)</td>
<td>17</td>
<td>0.425 - 0.860</td>
<td>Visible, Near IR</td>
<td>3.4a</td>
</tr>
<tr>
<td>MOC(^f)</td>
<td>1.4 - 11</td>
<td>0.5 - 0.9</td>
<td>Visible, Near IR</td>
<td>3.5a</td>
</tr>
<tr>
<td>CTX(^g)</td>
<td>6</td>
<td>0.5 - 0.7</td>
<td>Visible</td>
<td>3.4b</td>
</tr>
<tr>
<td>HiRISE(^h)</td>
<td>0.25</td>
<td>0.570 - 0.830</td>
<td>Visible, Near IR</td>
<td>3.4c, 3.5b</td>
</tr>
</tbody>
</table>

\[^a\] Neumann, et al., 2003; \[^b\] Archinal, et al., 2003; \[^c\] Christensen, et al., 2004; \[^d\] Rogers, et al., 2005; \[^e\] Pelkey, et al., 2007; \[^f\] Malin, et al., 1992; \[^g\] Malin, et al., 2007; \[^h\] McEwen, et al., 2007

The base for mapping is the 100 m/pix Mars Odyssey Thermal Emission Imaging System (THEMIS) daytime infrared (IR) dataset (Figure 3.1A, Table 3.1). This dataset provides albedo, adequate resolution, and nearly complete coverage within the map area. This dataset uses the thermal band to image relative surface temperatures, which provide
Figure 3.1: THEMIS IR mosaics showing brightness temperature signatures from (A) daytime and (B) nighttime. (A) The mosaic of daytime surface temperatures is commonly close to true albedo and is the base used for mapping. The red boxes indicate the location of figure 3.2 – 3.5. (B) The nighttime surface temperatures are a proxy for the size and competence of the material. Brightness (white) indicates warmer temperatures relative to cooler (dark) surfaces within the individual image strips in this non-calibrated, image mosaic. The brighter surfaces are indicative of materials that are more competent and/or have larger grain sizes. These materials have a higher thermal inertia (resistance to temperature change) which allows them slowly lose heat from daytime temperatures. The darker materials are low thermal inertia materials and usually indicate smaller grain sizes and/or less cohesion.
qualitative data on the thermophysical properties of the materials. As a compliment to this base, the THEMIS nighttime IR data (Figure 3.1B) provide nearly complete coverage of the map area, and provide additional detail about the thermophysical properties of the materials. In contrast, the Viking Orbiter 1 & 2 image mosaics provide global coverage, but at a reduced resolution, ~231 m/pix. The Viking data were used by the USGS to create the global Mars Digital Image Model 2.1 (MDIM 2.1) product (Figure 2.3 and 2.4). The MDIM2.1 was used in the few locations where gaps existed in the higher resolution base maps of the area.

As a complement to the base maps, moderate resolution images provide additional details about surface textures, landforms, and geologic relationships. The THEMIS visible range images (VIS, Figure 3.2A, Table 3.1) are ~18.5 km wide, 17 m/pix images, which provide more detail than the map base yet still provide context for the surrounding geology. The abundance of THEMIS VIS data in the map area made them an indispensable part of the mapping process. The Mars Reconnaissance Orbiter (MRO) Context Camera (CTX) provides ~28 km wide, visible band images at 6 m/pix (Figure 3.2B, Table 3.1). This is a newer dataset, so there are fewer images available in the map area. Where available, these images provide unprecedented detail over large areas giving context to the information from the scattered, higher resolution images.

Two high-resolution cameras provide detailed information at local-scales. The MGS Mars Orbiter Camera (MOC) narrow angle, visual range images are <4 km wide and, typically ~5 m/pixel (Figure 3.3A, Table 3.1). Additionally, the Mars Reconnaissance Orbiter’s High Resolution Imaging Science Experiment (HiRISE) visible
Figure 3.2: Image frames of the same feature at differing resolutions. This figure stresses the point that while high-resolution images are helpful in discerning what is happening at smaller scales, without context it is difficult to determine the geologic setting. (A) Portion of a THEMIS VIS image (18 m/pix, V14717005) giving a context for a feature on the bottom of a sinuous valley within the highland terrain. White box indicates the location of frame B. (B) Portion of a CTX camera image (6 m/pix, P02_002008_1558_XI_24S009W) showing mesas and knobs that are embayed and partially mantled by a smoother material, which contains small craters. White box indicates the location of frame C. (C) Portion of a HiRISE image (25 cm/pix, PSP_002285_1560) showing that the ridge (white arrow) is almost indistinguishable from the surrounding smooth unit. The increased resolution also shows more small craters on the smoother unit and a higher density of craters on the ridge. The ridge material is a competent material that is eroding into 1–2 meter-sized boulders (near the black arrow) that lie atop the smooth unit. North is at the top of all of the images and the orange arrows indicate the illumination direction.
Figure: 3.3: Portions of the two highest resolution Martian datasets showing the strengths of high-resolution datasets. (A) A portion of a MOC image (~3 m/pix, M07053871) with ridges, knobs, mesas, and small craters surrounded by darker materials and a bright plain. Box indicates the location of frame B. (B) A portion of a HiRISE image (25 cm/pix, PSP_007559_1570) shows an irregularly shaped, boulder-covered mound surrounded by radial eolian ridges. The bright materials from frame A (at lower left of the red box) correspond to the fractured, competent surface underlying the mounds and eolian mantle. North is at the top of all of the images and the orange arrows indicate the direction of illumination.
range images are ~5 – 7 km wide at 25 cm/pixel (Figure 3.2c and 3.3b, Table 3.1). Both of these datasets provide unparalleled insight into the geologic history of the map area. However, these images have smaller footprints, and are sparsely distributed throughout the map area. These data are used to provide details about the units they intersect. This is in contrast to the low and moderate resolution datasets, which are better suited for delineating the geologic relationships.

A related set of data used for mapping include the coarse mineralogy provided by orbiting spectrometers. The Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) multi-spectral pushbroom (~160 m/pix, 54 bands) and window (~80 m/pix, 54 bands) datasets provide this coarse mineralogy (Figure 3.4, Table 3.1). The available bands in CRISM are mixed to produce various summary parameters, which yield spectra similar to those for known mineralogies from laboratory experiments (Pelkey et al., 2007). In addition to CRISM, three, 3-band THEMIS daytime IR images [bands: 9-6-4 (12.57μm, 10.21μm, 8.56μm, respectively), bands: 6-4-2 (10.21μm, 8.56μm, 6.78μm, respectively), and bands: 8-7-5 (11.79μm, 11.04μm, 9.35μm, respectively)] provide mineralogy using a decorrelation stretch (DCS, Figure 3.5, Table 3.1). The DCS extracts unique spectral units and displays those units using differences in color; which provides coarse mineralogy (Rogers, et al., 2005). Unfortunately, both CRISM and THEMIS DCS cannot penetrate the extensive dust mantle in the southern highlands. The mostly heterogeneous mantle makes it difficult to determine the mineralogy of the underlying materials.

In addition to spectroscopic, topographic, and image data sets, previously produced geologic maps, of this region include the 1:5,000,000 scale geologic map of the
Figure 3.4: THEMIS daytime IR mosaic overlain by CRISM olivine summary parameter signature within the threshold for absorption. Note that most of the concentration is primarily on the crater floor except on the eastern crater wall. The relative paucity of olivine in the rest of the crater wall may indicate that the high concentration on the eastern wall is a false reading. Mass wasting deposits on the crater floor to south have only small portions showing an olivine signature indicating that the source of the olivine is likely from the floor of the crater, possibly flood basalts. North is at the top of the image.

Margaritifer Sinus quadrangle of Mars (0° and 30° south and 0° to 45° west; Saunders, 1979). Due to the resolution of the Mariner 9 image base (~1 km/pix) and the map scale, it lacked detailed unit delineation, unit descriptions, and stratigraphic correlations. The map does provide an overview of the resurfacing history within the region. The
Figure 3.5: THEMIS decorrelation stretch, 3-band images showing differences in mineralogy. (A) Image showing bands 6, 4, and 2, where green corresponds to olivine-rich, low plagioclase materials, blue corresponds to basaltic materials, and pink is silicic materials. (B) Image showing bands 8, 7, and 5, where purple corresponds to olivine-rich, low plagioclase materials, orange corresponds to basaltic materials, and yellow is silicic materials. (C) Image showing bands 9, 6, and 4, where purple corresponds to olivine-rich, low plagioclase materials, pink corresponds to basaltic materials, and yellow is silicic materials. The images appear diffuse because of the mantling deposits giving it a grainy appearance. The color discrepancies between images are due to the amount of mantling materials and the ability to see through the mantle based on different band wavelengths. Each image is 30 km wide and north is at the top of all of the images.
1:15,000,000 scale geologic map of the western equatorial region of Mars (Scott and Tanaka, 1986) and the 1:15,000,000 scale geologic map of the eastern equatorial region of Mars (Guest and Greeley, 1987) provide a global context of geologic units and their correlative stratigraphic relationships. Lastly, two 5°x5° (17.5°-27.5°S and 345°-350°E), 1:1,000,000 scale geologic maps of the Paraná-Loire Valles quadrangles adjacent to the western boundary of the study area (Grant, et al., in press) provided the starting point for delineating geologic units, making stratigraphic correlations, and detailed descriptions of some of the map units in the current study area.

Image Processing

With the exception of MDIM2.1 and MOLA data, processing is necessary to produce map-projected images from all Martian datasets. Standardized global map-projected products of MDIM2.1 and MOLA are available for download, and only require clipping out of the region of interest. The MOLA derived shaded relief, slope, and contours are generated using the 3D Analyst package in ArcMap. Image data uploaded from the PDS must go through ISIS processing to convert from PDS format to GIS-ready, map-projected format. There are three levels of processing: (1) Level 0, converts the image from PDS to an ISIS format; (2) Level 1, applies the corrections for atmosphere, sun angle, and sensor noise in the image to remove errors; and (3) Level 2, projects the data into a specified map-projection. Images in the standard PDS format are stored parallel to the orbit of the spacecraft, but when processed and map-projected, the images are rotated to match north-south directionality. This process creates an image collar that maintains the rectangular shape of the raster image. The collar contains a unique DN value to designate it as no data; however, the collar area often requires a
cosmetic, post-processing step to display the images *sans* collar. Below is a description of the processing methodologies for each of the datasets detailed above from the PDS format to GIS-ready format:

- CRISM images are the most labor intensive to process and require the most attention when processing. These images utilize an add-on to the IDL ENVI software package, called the CRISM Analysis Tool (CAT). CAT converts the multi-spectral PDS format into the ENVI format, atmospherically corrects the data, destripes the data, sets a projection, and concatenates the absorption bands into summary mineralogic information. The transition from PDS to ENVI format makes the data easier to manipulate within the program. The atmospheric correction uses two images of Olympus Mons, one at the caldera, the highest point on Mars, and one at the base, ~21 km lower, to calculate the atmospheric disturbance coefficient, which is then subtracted from each image band. The destriping routine removes the linear artifacts generated by the camera during data acquisition. Conversion of the data into projected space using the header file information, places the data into a mapping context. Finally, the 54 multi-spectral bands of data from the instrument detectors are processed. Particular combinations of mixed bands and band ratios produce valuable mineralogic information referred to as summary parameters. For multi-spectral pushbroom and window datasets, there are 34 summary parameters per image. Pelkey et al. (2007) provide a detailed overview of these parameters and the band ratios used to derive the parameters. These data are compiled into a mosaic, which can be loaded in a GIS. This produces a very colorful dataset, but ultimately inaccurate, because no detector thresholds have been applied. Application of the
threshold is a tedious process in ENVI. A region of interest (ROI) is selected for each band of the mosaic, one band at a time. Detector thresholds are applied to regions of interest and the specified summary parameters are exported as ESRI ArcGIS shapefile formats. Analysis of the exported data ArcGIS produces map views and statistical information.

- THEMIS daytime and nighttime IR data processing converts the Reduced Data Record (RDR) radiometrically corrected spectral file into an 8-bit, projected Tagged Image Format File (tiff). Images are processed using the Arizona State University (ASU) THEMPROC website (URL:http://themproc.mars.asu.edu) which utilizes ISIS in an online graphical user interface. The processing uses the THEMIS-specific ISIS subroutines to remove spacecraft wobble, the collar, and striping artifacts, and it corrects the radiance, sets the projection, and removes the collar. Projection information, image extent, resolution, and export type are set prior to export. This process exports a band-9 (λ=12.57μm), 32-bit image containing the projection information and a band-9, 8-bit, grayscale, 100 m/pix image with a running standard deviation stretch. The running standard deviation re-stretches statistical subsets of the image DN values to create a cosmetically superior product versus a simple standard deviation stretch, which re-stretches the image based on statistics compiled from the entire image. The web interface also creates the THEMIS daytime IR DCS images for mineralogy, the only difference is during band selection for these image types, three bands are chosen and the DCS-stretch option is applied. After the images have been processed and uploaded, the USGS’s isis2world.pl (PERL programming language script), available at the Planetary Interactive GIS-on-the-Web Analyzable
Database (PIGWAD, URL:http://webgis.wr.usgs.gov), extracts the projection information from the 32-bit cube and creates a worldfile, a text file containing the location and pixel size information, to display the image data in the GIS program properly. Adobe Photoshop allows selection of the backplane and sets its DN value equal to zero. Inverting the selection allows the image data to be set to a DN range of 1 to 255. This allows the image to display sans collar in the GIS. Additional georeferencing to the MDIM2.1 map base fine-tunes the image location, which is necessary for all THEMIS data due to inexact targeting information. Georeferencing is the process of manually matching locations on an image base to the image that needs to be spatially adjusted. After georeferencing, the individual image strips are compiled in Global Mapper GIS software and a full-resolution mosaic is generated for display in ArcGIS.

- THEMIS VIS images processing uses the “USGS THEMIS VIS HTML Loader Toolbox” add-on (available from PIGWAD), which was written for ESRI’s ArcGIS ArcToolbox. This tool downloads a projected image and the ISIS image that contains the projection information from ASU. This ISIS image is used to create a projection file (prj). A typical prj file contains the projection information and has following format:

```
PROJCS["Mars_Sinusoidal_clon2",GEOGCS["GCS_Mars_2000_Sphere",DATUM["D_Mars_2000_Sphere",SPHEROID["Mars_2000_IAU_IAG_Sphere",3396190.0,0.0],PRIMEM["Reference_Meridian",0.0],UNIT["Degree",0.0174532925199433]],PROJECTION["Sinusoidal"],PARAMETER["False_Easting",0.0],PARAMETER["False_Northing",0.0],PARAMETER["Central_Meridian",-2.0],UNIT["Meter",1.0]]
```

An ArcToolbox script, “Project Image Using PRJ file”, transfers the projection information from the prj file to the image. Projection of the image within ArcGIS creates a 16-bit image, which is converted back into an 8-bit format using another
ArcToolbox script, “Covert 16- to 8-bit”. As with the THEMIS daytime and nighttime IR images, the collar is removed, the image is georeferenced, and the images are mosaicked.

- MOC narrow-angle processing uses MOC-specific ISIS commands. The process is similar to those run on THEMIS daytime and nighttime IR images. The ISIS routine, MOCLEVALL, processes the data from PDS format to map-projected tiff format, stepping from level 0 through level 2, as mentioned in the above section. Cosmetic and location post processing are necessary to display the images in a GIS.

-CTX images use ISIS subroutines and Geospatial Data Abstraction Library (GDAL) scripts to convert from PDS format into a projected Joint Photographic Experts Group 2000 (JP2) format. The JP2 format compresses the file size, which is useful when dealing with high-resolution, large-area, contextual images. The ISIS subroutines convert from PDS to ISIS format, calculate statistics, and generate level 0, 1, and 2 ISIS files. The GDAL script converts from the ISIS level 2 cube into a lossless compressed JP2. These images do not require any post processing because of better-quality targeting information provided by the spacecraft. The backplane for these images is assigned a value of zero during processing, negating the need for cosmetic corrections.

- HiRISE images use a GDAL script to create a prj file for the downloaded, preprocessed, JP2 image. The same steps for applying the projection information from a prj as those for THEMIS VIS, complete the processing. These images do not require any location or cosmetic corrections.
It is important to note that data processing was critical for this project and took a significant amount of time. Incremental processing occurred throughout the project as the various missions released images to the public. It should also be noted that additional datasets do exist, but were not used for this study due to insufficient resolution at the mapping scale, no coverage in the map area, or a newer dataset rendered them obsolete.
Chapter 4

GEOLOGIC/GEOMORPHOLOGIC MAPPING

Geologic maps provide scientists with a mechanism to convey physiographic context, geologic history, geomorphic processes, and temporal relations between adjacent units. For Mars, geologic maps are produced using (primarily) remote-based datasets and (comparatively) small map scales (i.e., large map regions). As such, the methods that are commonly employed to ascertain geologic and geomorphologic characteristics and relations, to describe and interpret discrete geologic units, and to construct a temporal framework are considerably different from those applied on Earth. I outline the regional geologic history (and variants thereof) of the region of interest using the geologic map, unit descriptions, and correlation of map units.

Mapping Methods

Geologic mapping methods included a variety of techniques. In the following section, I describe the methods employed in unit delineation, naming, and grouping, as well as mapping limitations and symbol usage that appropriately and efficiently convey the map content. Additionally, this section includes a discussion of the map projection, the crater statistics, and the development of the correlation of map units (COMU).

Unit Delineation

I identified discrete geologic units in this region of Mars based primarily on superposition relationships and perceived primary emplacement morphologies, where available. In addition, perceived secondary characteristics were used to describe and divide units. Post-depositional erosion or tectonic processes typically cause the secondary characteristics, which form geomorphologies unique to a particular unit. The
interpreted depositional environments and emplacement processes provided a means to group the geologic units. Thermophysical properties and mineralogy, where discernable, were used to describe the units further.

Superposition relationships include embayment relationships, where a younger unit onlaps or buries another unit, providing a temporal gauge that shows the “burying” unit is younger than the “buried” unit. Also helpful to these analyses are crosscutting relationships with formational landforms and tectonic structures. For this region, I found crosscutting relationships between geologic units and fluvial geomorphologies, tectonic ridges, and crater ejecta particularly helpful in discerning the temporal history of these features. Primary emplacement features are generally landforms that can be attributed to the deposition of materials or formation of the materials by cratering processes (i.e., floor, rim, ejecta formation). The lengthy exposure to surface processes results in overprinting of primary features by characteristics of secondary, post-emplacement processes, including tectonic deformation (e.g., wrinkle ridges).

Secondary processes impart geomorphologies that allow for further characterization of the megaregolith (see chapter 1). The megaregolith covers most of this study area and lacks distinct contacts, which makes it difficult to identify lenses or blankets of material clearly. It is possible to delineate the boundaries of these units based on the geomorphology imparted by post depositional processes. In the map area, secondary processes expose stratigraphically lower, and possibly older, sediments and rocks, allowing for the more detailed characterization of regional stratigraphy.

The thermophysical properties of different units in the map area are examined using the THEMIS nighttime IR dataset. It is often difficult to differentiate whether these
properties are primary or secondary in nature. As such, the thermophysical properties of
the geologic units are used primarily to subdivide units with uncertain superposition
and/or geomorphologic relationships. Some geologic units do not have informative
thermophysical properties; these are noted in individual unit descriptions.

Although spectral data alluding to mineralogy do exist, the presence of an
extensive, thin sedimentary mantle hinders the designation of mineralogic or lithologic
descriptors for the geologic units (i.e., olivine-rich, basalt, carbonate, etc.). Units were
not delineated based on the mineralogic data, but the mineralogy was described wherever
possible. These occurrences were uncommon because of dataset limitations, extent of
mantling, and heterogeneity of the highland materials. The THEMIS DCS data provide
nearly complete coverage but the data typically show diffuse mixtures of material
compositions due to the heterogeneity of the mantling dust (Figure 3.5). Rarely in this
mapping area do the THEMIS DCS show distinct color variation between adjacent units
of differing mineralogy and/or lithology in areas where no mantle is present. The
CRISM dataset (Figure 3.4) does not cover the entire mapping area and has similar
problems with the mantling deposits. Moreover, the detection of mineral types often
varied considerably over adjacent or overlapping images and datasets depending on the
seasonal, atmospheric, and orbital parameters.

Unit Names

Unit name ages are based on the relative positioning of the units and the crater
ages. Unit name descriptors are based solely on depositional setting and stratigraphic
relationships. Similar with Tanaka et al. (2005), the processes that erode the unit are not
reflected in the naming convention. Craters designations are based on relative
degradation of the rim and ejecta material. Although very coarse mineralogic information exists, there is no reference to rock-type or mineralogy in the name.

Naming Groups and Symbols

Depositional environments are the principal means of grouping the geologic units. Stratigraphic position within these groups further subdivides these units. Three different groups exist in the mapping area: (1) the megaregolith group; (2) the basin group; and (3) the crater group. The megaregolith and basin groups are subdivided based on their stratigraphic relationships. The crater group is based on the relative degradation of the rim, floor concavity, and the continuity of the eject blanket. In locations where the craters are breached, crater floors were included in the basin group. The crater group is diachronous and therefore is not assigned a chronologic symbol.

The symbols for the name abbreviations are: (1) the chronologic period(s); indicated by large capital letters (N = Noachian, H = Hesperian, and A = Amazonian); (2) the depositional setting; indicated by a lower case letter (m = megaregolith, b = basin, and c = crater); and (3) the relative stratigraphic position; indicated by a subscript, sequential numbering system (1 = oldest unit). Accordingly, the symbol HNm\(_1\) indicates the Noachian Hesperian aged, lower megaregolith unit.

Determining Relative Ages

Crater measurements were compiled on the THEMIS daytime IR mosaic (100 m/pix) using a crater counting tool developed for the Mars crater cataloguing project (Hare et al., 2006). This add-on to ArcMap allows the user to measure craters using a three-point circle and stores the dimensional information and user-defined morphologic attributes in a tabular format. Crater counts were exported to a spreadsheet program for
analysis and plotting. The crater counts are separated by their associated geologic unit, sorted in descending order according to diameter, and then sequentially numbered from largest to smallest, starting with the number 1. Next, the area of the unit is scaled to 1,000,000 km$^2$. This scaler is also applied to the sequential numbering, giving the cumulative number of craters of that diameter, per 1,000,000 km$^2$. Finally, the error is calculated by taking the square root of the sequential number and dividing it by the inverse of scaling factor (Appendix B). This quotient is the amount of error based on the relative position within the cumulative list of sizes. The largest crater diameter in the population always has a 100% error, and error decreases with decreasing crater diameter.

To determine the relative ages, the N(2), N(5), and N(16) numbers are compared to the ranges presented in Table 2.2. The calculated values and errors were used to determine a range of time the unit has been exposed to impact processes. The length of time is determined by the range of ages of adjacent units, their superposition relationships, and the crater flux rate. From this, a correlation of map units was generated with units arranged with respect to time and each other. It is important to note that the “range” of geologic time estimated for a particular unit does not inherently indicate that the geologic processes were active across the entire time range. Rather, crater counts provide a temporal bracket for the primary emplacement of a particular unit and are intended to be used interpretively along with other temporal relationships and knowledge of regional to global geologic events. Units that are stratigraphically older with lower crater densities do exist in the map area, and it is important to use all of the available data to determine the period in which these units were emplaced.
Map Base

*Symbol Usage and Mappable Limits*

The 1:1,000,000 map scale for this study equates to 1 mm on paper to 1 km on the Martian surface. Thus, the smallest features mapped are craters with diameters $\geq 2$ km ($\sim 3$ km$^2$, 2mm on paper) because of their importance to the relative chronology. Other than craters with diameters of 2 - 12 km, all of the geologic polygons exceed an area of 100 km$^2$. Linear features are typically $>10$ km in length and are represented with line symbols.

*Projection*

The projection used for the raster and vector map data was a simple cylindrical projection. This projection creates an equal area rectangle using a common central parallel, in this case, the equator. The distortion of the map increases away from the equator, and the map area in this study is centered 22º from the equator. The accuracy of area and distance measurements decrease proportionally with distance from the equator, but ArcMap is able to produce accurate measurements from the projection settings. Seidelmann, et al. (2002), defined the Mars 2000 spheroidal datum, which uses the equatorial radius to project the data about a sphere. This is a simplified model compared to the actual ellipsoid shape of Mars with the radius decreasing towards the poles. It is important to note that the U.S. Geological Survey standard for Martian maps is the Martian Transverse Mercator (MTM) projection. The square shape of the simple cylindrical projection makes it ideal for displaying the data.
Figure 4.1: Topography of the map area with basins (b) and valleys (v) denoted. The blue dashed line represents the divide between areas that drained to the north and those that carried sediments to the west into Margaritifer Terra. The black dashed line roughly represents the divide between Noachis Terra and Arabia Terra, and closely mimics the MOLA zero contour. The basins of Noachis Terra are mostly formed by craters, whereas the basins of Arabia Terra are broad, flat depressions that appear unassociated with craterforms. MOLA 128 pixels/degree colored shaded-relief.
Results

Physiography and Geologic Setting

The study region (17.5° to 27.5° S, 0° to 10° W; Figure 4.1) slopes southeast to the north-northwest and coincides with the regional physiography across Arabia Terra. The total relief in the map area is ~4,400 meters, including deep crater floors. The relief of the inter-crater plains is ~2,800 m. The region includes Noachis Terra and Arabia Terra ("terrain", pl., terrae), Peta and Newcomb craters, and a portion of Paraná Vallis ("valley", pl., valles), as defined by USGS and IAU (see Figure 4.1 and Appendix A). The terrae are large areas of ancient highland terrain characterized by landforms of fluvial, impact, and volcanic origin. In the map area, Noachis Terra and Arabia Terra contain rises, crater rims, and local slopes that are radial to the larger craters. These local slopes typically dictated the drainage directions. Newcomb crater is an ancient, flat-floored impact basin located along the eastern boundary of the map region. Newcomb crater has a high-standing western rim, which forms a drainage divide between channels that flow towards the interior of Newcomb crater and those that indicate flow away from the crater rim. The majority of sediments and rocks that define the floor deposits of Newcomb crater were likely sourced from the east where roughly one-third of the crater rim is embayed. The channels that occur along the western rim of Newcomb crater indicate flow into Noachis basin and north into Arabia Terra.

Noachis basin (Figure 4.1) is a ~350 km diameter inner ring of a dual-ring impact basin. The 400-kilometer-diameter outer ring is topographically subdued and located outside the map region (Schultz et al., 1982). Noachis basin appears to have been the depocenter for the majority of the upland material transported through valley networks.
located in both the south and southeast part of the map region as well as off the western flank of Newcomb crater. The floor of Noachis basin slopes to the northwest, following the same regional trend observed across the map region though at slightly lower slopes. The rim of Noachis basin, where discernable, has been densely incised by fluvial activity and overprinted by subsequent impacts. Depressions and craters to the south and southwest of Noachis basin collected sediment during fluvial activity from three, east-northeast-trending valley networks. A single outlet located along the northwest rim of Noachis basin, and several adjacent valleys, formed in a U-shaped (planview), highly degraded region; and deposited materials in a basin to the northwest (“b” in the upper left corner of Figure 4.1).

West of Noachis basin, a north-south trending rise separates the valleys of Paraná Vallis (which likely transported sediments into Paraná basin, located to the west of the map region) and other unnamed valley systems that indicate flow toward several small basins south of Noachis basin. Northwest of Newcomb crater and northeast of Noachis basin, local slopes are to the north. An unnamed, small, flat-floored basin (“b” in the upper right corner of Figure 4.1) forms at the termination of valley systems, which indicate basin-filling sediments, were deposited from north-trending flow.

The valleys and valley networks that dominate the topographic character of the map region consist of small canyons that incise near-surface rocks and sediments of the Martian megaregolith. In general, the valleys have wide (100’s of meters to kilometers), flat-floors and steeply sloping walls in the medial and distal reaches. The floors in proximal reaches are typically filled with colluvium deposited through mass wasting from over-steepened walls. Scalloped valley rims, sometimes with debris apron below,
are present throughout the valleys, but typically increase in frequency in proximal reaches. The valleys are consistently U-shaped in plan view and have little variation in width from the proximal to distal portions. Valleys generally range from ~10 km to >300 km in length and are characteristically digitate (finger-like) with few tributaries. However, the area south of the main inlet into Noachis basin has a “classic” terrestrial dendritic pattern with many tributaries feeding a large trunk valley. The depth of most valleys increases towards the distal reaches with depths typically <1 km.

Impact craters occur throughout the map region, and range in size from <1 kilometer to >350 km in diameter. The craters show contrasting preservation or erosional states ranging from “pristine” to “highly-degraded”. Pristine craters have well-defined rims that are elevated above surrounding terrains, circular (or near circular) ejecta blankets, concave floors, and layers exposed in the crater walls. Highly-degraded craters have poorly-defined or nonexistent rims, no discernible ejecta blanket, horizontal floors, gullied interior rims, no exposed layers, and colluvium along the base of the interior crater walls. Typically, highly-degraded crater rims are incised by valleys and/or embayed by rocks and sediments of the intercrater plains.

Geologic/Geomorphic Units

The geologic/geomorphic units of the study area are divided into the megaregolith, basin, and crater-related units. The megaregolith units include materials that are interpreted to have been emplaced through ancient impact and volcanic processes. The basin units include materials associated with crater formation and subsequent erosion through (chiefly) fluvial activity. The crater-related units include materials identified based on preservation or erosional state. Descriptions,
Figure 4.2: Map of study area with symbols and correlation of map units. Full-scale version in Appendix C.
interpretations, and type localities are detailed in the following section. A reduced-size, complete map is presented in Figure 4.2, and a full-scale version of the map is presented in Appendix C.

**Megaregolith Group**

**Noachian megaregolith unit 1** (Nm₁, \(N(16) = 86 \pm 22\), Table 4.1): Forms broad plains that contain fluvial landforms. This unit is exposed in the scarp walls of the valleys and valley networks. This unit contains ridges and mesas of the overlying unit (HNm₂) of insufficient size to map. Primary emplacement features are not generally recognizable in this unit, although some layers are observed in the interior walls of impact craters in this unit. Secondary features are dominantly of a fluvial origin, with overprinting by more recent aeolian processes. Thermophysical properties contain a diffuse mixture of high (bright) and low (dark) digital number (DN) values in THEMIS nighttime IR. The unit consists of a mix of mafic minerals, although THEMIS DCS and CRISM images suggest a mélange of rock types and/or mantling (Rogers et al., 2005).

**Interpretation:** Megaregolith emplaced primarily through impact processes though intercalated to various degrees with volcanic rocks and sediments and possibly localized fluvial sediments or colluvium. The layering within crater walls indicates the presence of stratified rock or sediments, although this may be due to the impact processes. If these materials include volcanic flows, they are likely flood basalts because no source vent is present in the map region. The valleys located in the unit expose older deposits with increasing depth, though age variation within the exposed deposits cannot be confidently ascertained. The thermophysical properties suggest that the material contains a mix of grain sizes and/or a range of induration. The bright areas suggest
relatively more induration and/or lithic blocks than stratigraphically younger megaregolith, perhaps as the result of gravitational compaction, contact metamorphism with volcanic materials, and/or chemical cementation imparted by the fluvial or groundwater processes. The primary thermophysical features of this material may be overprinted or enhanced by recent eolian processes. The mafic mineralogy dominates the unit and implies that the composition is a combination of material derived from older, crust-forming volcanic materials and younger, differentiated volcanic materials (Mustard, et al., 2005). Age was determined by a combination of the crater density and the stratigraphic position with the HNm2 unit. The value for N(5) crater density is less than the HNm2 unit. This discrepancy is possibly due to the burial by the HNm2 unit, or by resurfacing due to fluvial processes. 

*Type Locality: 26°S, 3°W, Figure 4.3.*

### Table 4.1: Area, crater count, and superposition of mapped units.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Area (km²)</th>
<th>N(16)²</th>
<th>N(5)²</th>
<th>N(2)²</th>
<th>Superposition relations¹b</th>
</tr>
</thead>
<tbody>
<tr>
<td>m1</td>
<td>185145</td>
<td>86 ± 22</td>
<td>319 ± 42</td>
<td>859 ± 68</td>
<td>~b1, b2; &gt;m2, b3, b4</td>
</tr>
<tr>
<td>m2</td>
<td>86835</td>
<td>81 ± 30</td>
<td>288 ± 58</td>
<td>955 ± 105</td>
<td>&lt;m1; ~b1; &gt;b2, b3, b4</td>
</tr>
<tr>
<td>b1</td>
<td>53880</td>
<td>55 ± 32</td>
<td>260 ± 69</td>
<td>668 ± 111</td>
<td>&lt;m1; ~m2; &gt;b2, b3, b4</td>
</tr>
<tr>
<td>b2</td>
<td>2315</td>
<td>NA</td>
<td>NA</td>
<td>864 ± 611</td>
<td>&lt;m1, m2, b1; ~b4?; &gt;b3, b4</td>
</tr>
<tr>
<td>b3</td>
<td>80163</td>
<td>87 ± 33</td>
<td>286 ± 60</td>
<td>948 ± 109</td>
<td>&lt;m1, m2, b1; ~b4; &gt;b4</td>
</tr>
<tr>
<td>b4</td>
<td>59487</td>
<td>NA</td>
<td>NA</td>
<td>NA</td>
<td>Present in c1, c2, and c3 craters</td>
</tr>
</tbody>
</table>

²N(x) = total number of craters >x km in diameter per 10⁶ km²

¹ < is younger than; ~ is overlapping in time; and > is older than

**Hesperian Noachian megaregolith unit 2 (HNm2, N(16): 81 ± 22, Table 4.1):**

Forms relatively smooth and areally-expansive surfaces that contain north-south-trending narrow ridges or scarps, typically ~100 m in relief. The ridges are commonly curvilinear and typically have east-facing concavities. The ridges are usually steeper on the western face with gentle slopes on the eastern side. Similar ridges occur on the floors of Noachis basin and Newcomb crater. This unit has average thermal inertias displaying mid-range DN values within the THEMIS nighttime data. However, the ridges sometimes have
Figure 4.3: (A) Geologic map showing the type location of dissected megaregolith Nm\textsubscript{1} terrain. Note the dendritic tributaries on the eastern margin of the map area versus the broad morphology of the trunk valley in the west-central area. (B) THEMIS nighttime IR mosaic (100 m/pix) of the tributary system that is likely fluvial systems because of the proximity of the head of the valleys to the unit HNm\textsubscript{2} divide. The dense valleys are darker than the surrounding terrain, indicating less consolidated material deposits on the valley floors. Brighter materials at the distal portions of the valleys may indicate a more consolidated material possibly through chemical cementation or precipitant. (C) A portion THEMIS Visible range image V15790004 (17 m/pix) showing the dendritic and digitate (finger-like) morphologies.
relatively lower thermal inertias. This may be due to the material properties or the accumulation of fine-grained materials emplaced through aeolian processes. Unit mineralogy is indistinguishable from that of the Noachian lower megaregolith unit. 

*Interpretation:* Megaregolith emplaced by similar processes as the megaregolith unit 1 (Nm1), though it is younger based on stratigraphic relationships and exposures in the walls of exposed valleys and craters. The ridges are crosscut by valleys in some locations and in other locations ridges are crosscut the valleys, suggesting coeval and/or long-term contribution to unit development as a secondary characteristic. The ridges are likely tectonic (wrinkle) ridges formed by lateral shortening. Although Mars has been relatively tectonically quiescent compared to the Earth (Banerdt, et al, 1992), these features are hypothesized to be compressional tectonic features (Watters, 2004). These features may exploit preexisting subsurface fractures that developed during impact events. The ridges are present on basin floors and on the inter-crater plains units. The age range was determined using the superposition of this unit with the Nm₁ unit. The age range was determined using the N(16), N(5), and N(2) statistics. *Type Locality:* 24°S, 8.5°W, Figure 4.4.

**Basin Materials Group**

Crater floors are commonly buried by colluvium shed from the interior crater walls. Floor materials appear strikingly brighter than surrounding material in THEMIS nighttime IR images and slightly darker than surrounding materials in THEMIS daytime IR images. In THEMIS visible range images, the surface appears denuded compared to the adjacent materials.
Figure 4.4: (A) A portion of the geologic map showing the ridged terrain of the HNm$_2$ unit. The Nm$_1$ unit is exposed by fluvial downcutting through the upper megaregolith to expose the older deposits. (B) THEMIS visible range image V17737007 (17 m/pix) of the ridges where erosion has exposed multiple scarps and benches within the HNm$_2$ unit wrinkle ridge, possibly indicating layering within the HNm$_2$ unit. Black square indicates the location of (C) a portion of MOC image E04007951 (~5 m/pix) of the scarp face with the up (U) and down (D) portions of the ridge are indicated. The scarp has a gentle slope that has been impacted but appears to have been exposed for less time than the surrounding plains.
Noachian basin unit 1 (Nb₁, N(16): 55 ± 32, Table 4.1): This unit consists of angular plates typically <100 m² often separated by meter-scale fractures that are filled with low albedo material. The unit contains the lowest thermal inertia values of any material in the map region. Ridges, similar to those in the HNm₂ unit are present. The THEMIS nighttime IR data indicates the unit consists in some areas of admixtures of low and high thermal inertia materials. The mineralogy appears diffuse in THEMIS DCS and CRISM, this may be because materials similar in composition to that of the megaregolith (MacKinnon and Tanaka, 1989; Squyres et al., 1992), mantle most of the area.

Interpretation: Brecciated basement rocks related to the formation of impact basins and may represent the original crater floor. Fractures on the surface are exposed portions of the crust, which was fragmented upon impact. This unit is the bright material found on the floors of most of the larger diameter craters in the map area. This brightness indicates that the material is more competent or consists of larger grain sizes than the surrounding or overlying material (Christensen et al., 2003). The mixed thermal inertia values for this unit may be attributable to the material filling the narrow fractures. The mineralogy is indistinct but is likely similar to the source of impact-emplaced megaregolith giving it similar characteristics. The age of this unit was determined by its stratigraphic position as compared to the Nb₂ unit. The Nb₂ unit appears to be on the surface of the Nb₁ unit. The N(2) statistics for unit Nb₂ indicates a Noachian origin, thus the Nb₁ unit must also be Noachian aged. The crater statistics discrepancy could be due to resurfacing of the Nb₁ surface by fluvial and lacustrine deposits. Type Locality: 22.5°S, 5.5°W, Figure 4.5.
Figure 4.5: Portion of the map centered at 22.5°S, 5.5°W showing materials on the floor of Noachis basin. (B) A portion of HiRISE image PSP_007559_1570 (25 cm/pixel) showing the fractured, platy texture of the ancient basin floor of the Nb₁ unit. This image is not in frame A, but is representative of unit Nb₁. (C) A portion of MOC image E10045971 showing the remnant floor material of the Nb₂ unit with linear to curvilinear dikes, of either sedimentary or volcanic origin, forming both the scarp edges and internal structures within the material. (D) A portion of THEMIS VIS image V08115004 (17 m/pix) showing the exposed Nb₁ unit with linear features interpreted as remnant dikes remaining after erosion of Nb₂ floor unit. (E) A portion of MOC image M02039171 (~5 m/pix) of a well-preserved crater rim. The crater rim shows multiple layers, some possibly due to the impact processes but the lower stratigraphy hinting at the underlying stratigraphy of ancient crust.
Noachian basin unit 2 (Nb₂, N(2): 864 ± 611, Table 4.1): This unit forms scarp-bounded blocks and islands of materials with hummocky surfaces. Within the islands, small linear to curvilinear ridges, similar to those that form the scarp margins of the islands, protrude out of the surrounding material. Typically, aprons of mass wasting materials surround the base of the landforms. These materials have a higher thermal inertia than the surrounding deposits of basin unit materials. The mineralogy is indistinct compared to the surrounding terrain.

*Interpretation:* This unit is exhumed crater floor deposits possibly a mélange of breccia from the original impacts and materials from the ejecta of Newcomb crater. The boundary scarps and the internal ridges are volcanic or sedimentary dikes formed by materials filling the fractures. An alternative hypothesis, which cannot be confirmed through current spectral mineralogy, is a hydrothermal origin. The contact with the subjacent basin unit 1 is not observed, perhaps due to masking by debris aprons and colluvium. The higher thermal inertias of the blocky and island material compared to the basin 1 unit may be from the original material, or imparted from trapped sediments mantling the deposits. The age for these materials was determined because this unit overlies the Nb₁ unit. In portions of the Nb₁ unit, remnants of the dikes common in this unit are present. At the margins and within this unit where it is eroded, the Nb₁ unit is present. The stratigraphic relationship is more important because of the low crater population and small area of this unit. *Type Locality:* 23.2°S, 5.5°W, Figures 4.5 & 4.6.

Noachian Hesperian basin unit 3 (HNb₃, N(16): 87 ± 33): Forms large fans at the mouths of the valleys of the southeast, east, and northwest portions of Noachis basin.
In addition, this material forms smooth surfaces in smaller basins located southwest, northeast and northwest of Noachis basin. Unit often preserves inverted relief valley morphologies in the distal portions of the deposits. In THEMIS nighttime IR data, these deposits have average DN values, which correspond to average thermal inertias. Mineralogically, the deposits are similar to the high standing materials from which they are sourced. At the margins of this unit in the southeast portion of Noachis basin, a more resistant capping material erodes ~2 m diameter blocks onto the Nb1 unit. This area forms knobs, mesas, buttes, and ridges of material as secondary features. Radial ripples typically surround the structures. This unit likely contains at least two layers because resistant top layer is not seen in direct contact with the underlying Nb1 unit. The thermophysical properties of this area are twofold, with the geomorphologic structure displaying above average DN values in THEMIS nighttime IR data, and the eroded material showing a low thermal inertia (high DN values). The mineralogic data is difficult to discern, likely due to mantling deposits on the tops of the structures, steep scarps with minimal plan view exposure, and the small area of the features versus the pixel size of the spectrometer data.

**Interpretation:** Mostly fan sediment emplaced as valley networks debouched in Noachis basin. In the western portion of Noachis basin, the unit likely includes some Newcomb crater ejecta material at its base. The average thermal inertia of the material is likely due to poor consolidation of finer-grained material than the underlying blockier materials corresponding to fluvial transport. This unit includes some amount of the basin 2 unit at its base, and it is a gradational contact dependent on erosion. **Type Locality:** 28.9°S, 4.3°W, Figure 4.6.
Figure 4.6: (A) Portion of the map centered at 23.2ºS, 5.5ºW showing the margins of the HNb3 flow deposits at the southeast breach of Noachis basin. (B) A portion of a THEMIS nighttime IR mosaic (100 m/pix) of the same area shown in A, with the bright materials corresponding to possibly less-mantled or more-consolidated flow deposits of the HNb3. (C) A portion of THEMIS VIS image V23490002 (17 m/pix) with a segment of the main stem with a portion of positive relief in the center of the channel. To the southwest, the channel is exposed negative relief; this may be the transition from subareal to subaqueous deposition on an ancient fan. (D) A portion of THEMIS VIS image V08065005 (17 m/pix), showing the retreating front of the deposit forming knobs, buttes and mesas. This erosion exposes the underlying Nb1 unit and mantles it with a brighter material. (E) A portion of HiRISE image PSP_007559_1570 (25 cm/pix) of one of the small buttes shows the blocky nature of the upper layer of the HNb3 materials and the eolian reworking of the sediment shed from these landforms. Due to the low density of the THEMIS visible range and HiRISE datasets, the locations of frames B, C, and E are not noted on frame A but are within the units of interest near the type locality.
In the southeast portion of Noachis basin, the primary emplacement processes are likely similar to that of the megaregolith with some degree of material either deposited or partially cemented \textit{in situ} by lacustrine processes. Eolian processes, evidenced by the radial ripples surrounding the landforms, may strip material from the base of the structures causing backwasting of the resistant cap material. Small “erratic” blocks litter the surface of the Nb1 unit suggesting that this cap rock material was once more laterally extensive in the region. A chemical precipitant acting as a cementing agent may form a dura-crust. Additionally, eolian processes could have scoured the surface, creating a desert pavement. The crater statistics for this unit place the unit in the Late Noachian. This unit overlies the Nb1 and the Nm1 units. This unit is rarely in contact with the Nm2 unit because it is usually deposited in topographic lows in the region. \textit{Type Locality:} 23.7\textdegree S, 4.7\textdegree W, Figure 4.6.

\textbf{Amazonian Noachian basin unit 4 (ANb4, Present on the floors c1, c2, and c3 craters):} Forms the smooth floors of craters through non-fluvial processes. This unit has higher DN values (low thermal inertia) in THEMIS nighttime IR. In some locations, ridges are present at the margins of the floor, near the mass wasting deposits of the crater walls. These deposits are sometimes surrounded by a moat-like depression with somewhat different thermal inertia. The mineralogy is indistinct in most locations, although, in Peta crater, the material is high in olivine or similar iron bearing minerals.

\textit{Interpretation:} This is a volcanic or hydrothermal resurfacing material of the crater floors. Because of the thermal differences between the rim and floor materials, it is likely that flood lavas or upwelling, volatile rich fluid, sourced inside the craters, possibly
exploiting the fractures imparted by the impact processes, emplaced this material. The olivine mineralogies are linked to differentiated magmas of a younger age than those that were not differentiated that formed the crust. Not all of the small basins have indicative CRISM signatures. The age range of this unit is dependent on the degradation of crater in which it is deposited. Because the thermal signatures of this unit are similar, although mineralogies may vary, it is probable that these undergo similar processes in their formation. Type Localities: Type Locality: 26.7ºS, 8.45ºW.

**Crater Group**

The crater units are designated based on their degradational states. The amount of relief between the rim crest and the surrounding plains is a good indicator of the relative age of the crater.

If the crater rim has been breached by valleys, or if the crater rim is partially embayed by another unit, the crater is designated as c1 (likely the oldest). If the crater has an intact rim with only small breaches, but lacks a distinct ejecta blanket, it is designated as moderately degraded and c2. If the crater rim is significantly higher than the surrounding terrain, it has an ejecta blanket or has crater rays, it is designated as fresh or slightly degraded and labeled as c3 (probably younger).

The relative ages of the craters are time transgressive due to the dependency of preservation state on location. The material properties of the units into which the craters form are important for the preservation of the impact form. The location of the crater is important because of what erosional processes acted upon the crater. For example, a crater in an area of fluvial activity is more likely to be breached and eroded than one in an area that has only undergone eolian erosion. The materials are important because an
impact in more competent material will better preserve a crater than one impacted into soft sediment. The large ancient craters may be discernable because they were impacted into the ancient crust and not into a less competent material like the megaregolith as most of the subsequent craters were.

The c1 crater group is focused in the Noachian Age but extends into the Hesperian Age. The c1 unit has the smallest population of mapped craters due to extended exposure to surface processes including breaches and infilling due to fluvial processes in the Late Noachian through the Early Hesperian. The larger diameters of craters in the c1 unit also correspond to sizes more common during the Late Heavy Bombardment. Additionally, the largest diameter c1 unit craters (Newcomb crater, Noachis basin, and Parana basin) were the locations of the basin group deposition.

The majority of the mapped craters fall into the c2 unit (Noachian through the Amazonian), which is focused in the Hesperian Age. This is likely because these craters formed near the end of or after the peak fluvial erosion of the Late Noachian and Early Hesperian, and were not significantly modified by other erosional processes that were active from the Late Hesperian. The c2 unit craters have flat floors from gravity driven erosion from the rim materials sluffing off into the floor, or air-fall deposition from volcanics or dust storms. These are smaller diameter craters and are present on all units.

The smallest population of craters is within the c1 unit, which spans the Hesperian Age to present. These craters have intact rims, mostly circumferential ejecta blankets, and some have concave floors because of a lack of gravity driven erosion. Erosion of the c3 unit craters is likely limited to eolian and some degree of slumping due to gravity.
**Geologic History of the Map Area**

This section is graphically summarized in Figure 4.7 showing the duration of the major events and processes that shaped the map area. *Pre-Noachian and Noachian Period (>~3.7-3.5 Ga)*

- The ancient crust of Mars formed in the Pre-Noachian. During the Early Noachian, the late heavy bombardment continued to emplace large amounts of ejecta material, and volcanic processes, likely airfall deposition given the distance of the map area from volcanic constructs, forming the unit Nm$_1$. 

Figure 4.7: Chart showing the duration of major events and processes in the study area. Black oval is the formation of the early crust. Brown circle represents the Parana basin, Noachis basin, Newcomb crater formations near the end of the Late Heavy Bombardment. Yellow wedge indicates the waning cratering rate throughout time. Red oval is the contraction (wrinkle) ridge formation preserved in the HNm$_2$ and Nb$_1$ units. Blue oval represents the fluvial process active on the surface with morphologies preserved in the Nm$_1$ unit. Purple oval represents the possible hydrothermal and volcanic processes that may have been active in producing dikes in Noachis basin and later altering the floors of some of the areas craters. The beige oval represents the eolian processes active in the study area. The eolian processes may have been active in Noachian and Early Hesperian, but the evidence for this has either been buried or eroded.
• During the Middle to early Late Noachian, Paraná basin formed, west of the map area, followed by the formation of Noachis basin as a multiple-ring impact basin and unit Nb₁ was formed as the floor of the impact basin. Newcomb crater formed with a floor similar to that of Noachis basin (Nb₁). Newcomb ejecta deposits were emplaced on the floor of Noachis basin. The eastern flank of Noachis basin was overprinted by the rim of Newcomb crater, coinciding with the weakening or partial removal of the southeastern Noachis basin rim material. These three large impacts added to the thickness of the Nm₁ and the preserved HNm₂.

• During the Late Noachian, contractional (wrinkle) ridges began forming during this time in the HNm₂ and Nb₁ units. Impact rates began to decrease and most of the larger craters were emplaced. Unit Nb₂ was likely emplaced as volcanic and impact airfall materials. Volcanic upwelling, sediment infilling, and/or hydrothermal mineralization in Noachis basin likely is partially made up of the ejecta from Newcomb crater. The fractures in unit Nb₁ served as conduits for the material that formed the more resistant dikes in unit Nb₂.

• Valleys were beginning to form and to incise the loose megaregolith materials. The HNb₃ deposits began forming in Noachis basin as valleys transported material from the western flank of Newcomb crater and the plateau surface of Noachis Terra. The weakened or possible already breached southeastern rim of Noachis basin becomes the main conduit for water and sediment transported from the highlands into Noachis basin. The western and northern flanks of Newcomb crater were heavily dissected during this time stripping the ejecta material from
the area where the rims of Noachis basin and Newcomb crater would have intersected, and exposed a large scarp. Parana Valles formed during this time and began to erode headward toward the north-south-trending rise that was likely formed by a combination of the impacts that formed Noachis and Parana basin. On the western flank of the rise, several small valleys began to incise the megaregolith and transport water and sediment through a series of small basins before finally debouching into Noachis basin.

- Water began to pond in Noachis basin, and likely in the smaller crater basins during this time. Because of the proximity of the remnant fans to the crater rims, the transported sediment settled out near the mouths of the valleys, beginning to form fan morphologies on the basin floor. It is likely that the U-shaped basin to the northwest of Noachis basin was beginning to undergo a degree of erosion due to groundwater. The groundwater was likely being transmitted from Noachis basin down the regional slope using the radial and circumferential fractures of the impact that formed Noachis basin as conduits.

- During the Late Noachian, at least the western portion of Noachis basin was filled with standing water, evidenced by the paucity of linking valleys between the eastern and southern portions of Noachis basin and its single outlet on the northwestern flanks.

- A ~35 km diameter crater formed in northwest Noachis basin. This crater breached the rim of Noachis basin and rapidly began to fill water and spill over into a smaller crater to its north, forming a small fan of HNb₃ unit. This likely triggered a flood event(s) that removed a large amount of the Nm₁ unit from the
northwest portion of the map area. The smaller basins to the southwest, northwest, and northeast of Noachis basin also began amassing fluvially transported sediment.

**Hesperian Period (~3.7-3.5 Ga – ~3.3-2.9 Ga)**

- The emplacement of the HNb$_3$ and AHb$_4$ units continued into the Early Hesperian. Fluvial dissection and headward erosion continued into the Hesperian as deposition into the basins peaked. Cratering during the Hesperian might have interrupted fluvial systems or buried those that were already extinct. In some areas, groundwater may have kept some systems active in the northern portions of the map area.

- During the Late Hesperian, widespread fluvial activity ceases, Noachis basin empties. The lack of water does not allow the valleys to react to the change in base level leaving stranded valleys along the scarp to the west of Newcomb crater. Gullied interior walls of some of the c2 unit craters indicate that surficial water activity may have continued into the Late Hesperian.

**Amazonian Period (~3.3-2.9 Ga – Present)**

- Eolian and impact processes become the dominant processes on the surface.

- Basin units began to erode with the material of the HNb$_3$ unit being differentially stripped from the interior of Noachis basin. The fan deposits of the HNb$_3$ unit are stripped of upper surface materials exposing the well-cemented and preserved negative relief portions of the valley floors.

- Eolian materials are organized into thin sheet-like mantles over most of the map area. Eolian ripples and small dune forms form in some craters and transverse
eolian ripples form in some valley bottoms. Regionally, eolian materials, which typically have a very low thermal inertia, usually occur in the valleys that trend east-west. Whereas the valleys oriented more north-south are typically free of eolian ripples although sediment does accumulate. The preferential orientation of valleys trending east-west may indicate the prevailing wind direction during, at least, the Late Amazonian.
Chapter 5

HYPSOMETRIC ANALYSIS OF MARTIAN AND TERRESTRIAL BASINS

Introduction

To determine the dominant process forming the valleys in the mapping area, I quantitatively compared the topography and area from selected Martian basins to those of the tributary basins of the distal reaches of the Escalante River, southeastern Utah. These valleys provide a close morphologic analog to those seen throughout Mars. To examine whether overland flow or groundwater processes dominated the formation of valleys in the study area, I used basin hypsometry to compare quantitatively basins on the Earth and Mars using topographic data. Hypsometry is a technique used to compare three-dimensional surfaces statistically, regardless of scale differences. In terrestrial studies, hypsometry has been shown to discriminate between groundwater sapping (referred to as sapping in the remainder of the thesis) and fluvially dominated valleys based on their basin morphology (Luo 2000, 2002).

Planetary scientists use analogous terrestrial landforms to try to constrain formational/erosional processes active on other planetary surfaces (e.g., Howard and Kochel, 1988; Luo, 2000). Analog studies are either qualitative or quantitative dimensional analyses. Qualitative analogs use similarities in shape, structure, or depositional, tectonic, and latitudinal setting to form viable hypotheses about the formational and/or erosional processes. Quantitative analogs typically involve comparisons of the dimensional aspects of one feature to another (i.e., height, width, depth, radius, slope, etc.). For both qualitative and quantitative analyses, Mars-Earth analogs rarely have a 1:1 correlation because of (1) differences in physical parameters
(i.e., gravity, surface pressure, etc.; see Table 2.1); (2) differences in orbital parameters (i.e., distance from the sun, obliquity, etc.; see Table 2.1); (3) differences in scale; (4) similar landforms with multiple possible origins; and (5) poorly constrained characteristics of the Martian subsurface (i.e., rheology, structures, etc.).

This chapter presents a description of the geologic settings, morphologic similarities and differences, methods, results, and a discussion of the implications of this quantitative analog study.

**Geologic Setting**

This section describes of the geologic setting of the Escalante River and Escalante Arm of Lake Powell of the terrestrial study area. A discussion of the similarities and differences of the Martian study area is included at the end of each subsection. For a more thorough discussion of the geology of the Martian study area, see chapter 4.

The tributary valleys along the east side of the Escalante Arm of Lake Powell, in southern Utah provide an excellent qualitative terrestrial analog for the valleys of Mars. The morphology of these canyons is very similar to those presented in the Mars map area discussed in chapter 4 (see Figure 4.2 and Appendix C). Past studies of the stratigraphy, structure, hydrogeology, and geomorphology of the Escalante region provided data and hypotheses about how the valleys were forming within the basins.

*Stratigraphy and Structure*

In their distal portions, the canyons cut into the sedimentary rocks on the east side of the Escalante arm of Lake Powell (Figures 5.1a and 5.1b). Two canyons flow south from the Water Pocket Fold, and debouche directly into Lake Powell (until 1963, the
Figure 5.1: (A) Shuttle Radar Topography Mission (30 m/pix) color shaded relief map overlain on Digital Ortho Quadrangle (8 m/pix) showing the distal reaches of the Escalante River and the Escalante Arm of Lake Powell. (B) Lake Powell area in southeastern Utah (light blue area). The location of frame A is indicated by the red box. The stratigraphy (C) in the study area gently dips to the southwest from the Water Pocket Fold/Circle Cliff Anticline until reaching the Fifty-Mile Creek Syncline fold axis and the dip direction changes to the northeast. The dip of the bedding changes again along the Bridge Anticline axis near the base of Fifty-Mile Mountain. The valleys on the west side of the Escalante River exhibit typical fluvial morphologies, whereas the valleys on the east side have morphologies corresponding to a sapping dominated environment. Frame C modified from Hintze, 1988.
canyon of the Colorado River), east of the Escalante Arm (Figure 5.1a). The canyons range in length from ~0.1 km to 4 km and are ~30 to 500 m wide. Mostly sub-horizontal lying sandstones and shales of Paleozoic and Mesozoic age underlie this region of the Colorado Plateau [Howard and Kochel, 1988]. I discuss the stratigraphy of the Glen Canyon Recreation Area (Figure 5.1c) from oldest to youngest. The lowest unit is the Late Triassic Owl Rock Member of the Chinle Formation. It is composed of siltstones, mudstones, limestones and poorly sorted very fine grained sandstones [Kochel and Riley, 1988; Laity, 1988a]. The Owl Rock is overlain by the moderately well sorted, very fine-grained Early Jurassic Wingate Sandstone [Kochel and Riley, 1988; Laity, 1988a]. The interbedded sandstone, siltstone and mudstone of the Early Jurassic Kayenta Formation underlies the layer of primary interest for sapping morphology, the eolianitic late Early Jurassic Navajo Sandstone [Kochel and Riley, 1988]. The Navajo is dominated by aeolian cross bedding but contains limestone lenses, and mudcracked and structureless sandstones [Kochel and Riley, 1988]. Between cross-bedded units, silty sandstone deposits are common, with increasing frequency towards the base of the Navajo Sandstone [Kochel and Riley, 1988]. The Navajo Sandstone is overlain by sandy siltstone and limestone of the Middle Jurassic Carmel Formation that is hypothesized to interfinger with the Entrada Sandstone [Laity, 1988a]. The Page Sandstone is not present this far to the east of Page. All of the canyons examined in this field study are predominately incised into the Navajo Sandstone, with some incision into the underlying Kayenta Formation.

Although nearly flat lying, the local stratigraphy gently dips (~1°-2°) to the west-southwest due to the Water Pocket Fold monocline and Circle Cliff Anticline that trend
north-northwest in the eastern portion of the study area. To the west of the Escalante River, the dip direction reverses at the axis of the Fifty-Mile Creek Syncline (Doelling and Willis, 2008). The most pervasive structural features are joints and fractures within the layered rock [Howard and Kochel, 1988 and Laity, 1988a]. The orientation of the regional fractures in southeastern Utah trends N65° – 75°E with its lesser prominent pair trending N25°W [Laity, 1988a]. Exfoliation jointing parallel to the canyon wall in the Navajo Sandstone is likely due to pressure release from erosion of the canyon [Laity, 1988b]. These joints in conjunction with undercutting of the canyon walls by seepage facilitate collapse of large slabs of sandstone, and subsequently control their steep-walled morphology [Laity and Malin, 1985 and Laity, 1988b].

In the Martian study area, the stratigraphy includes the megaregolith, a mélange of fractured crustal and impact ejecta materials, and ancient crust (Figure 5.2). The stratigraphy/materials surrounding craters are uplifted with steeper dips nearer the rims. The dips decrease into the regional dip trend away from the rim. These dips may be similar to the gentle dips of the lower Escalante River area. Because the dips are radial to the rim crest, this could help explain why there are sapping morphologies near crater rims on Mars. The structures on Mars include wrinkle ridges, which are thrust faults associated with crustal cooling and fractures associated with impact processes (Figure 5.2). There is still debate on whether the wrinkle ridges represent thin- or thick-skinned deformation but it is hypothesized that thin-skinned (décollment) deformation would penetrate to a depth of 3 to 5 km (Watters, 2004), whereas thick-skinned deformation would extend to 10’s of km (Golembeck et al., 2001). Rodriguez et al. (2007) hypothesized that wrinkle ridges may have enhanced permeabilities. Rodriguez et al.
Figure 5.2: Hypothesized cross-section of stratigraphy and structures of the Martian subsurface. Grey is the megaregolith, black is ancient crust, and fractured crustal material, and the white lines are the fractures created by impact processes. From Hanna and Phillips (2005).

(2005) concluded that the fractures produced by impact processes extend 1/3 to 1 crater diameter from the crater rim. These radial impact fractures could have provided the conduits for enhanced subsurface transmission and may control the location of subareal water flow. The Parana basin, Noachis basin and Newcomb crater impacts could have filled the study area with an interconnected system of subsurface fractures. The presence of wrinkle ridges on the floors of both Newcomb crater and Noachis basin, as well as their presence within the megaregolith, could have provided subsurface conduits through which water was transported. These ridges and fractures may be analogous to the regional jointing in the of the Escalante River region of the Colorado Plateau.

Hydrogeology

The groundwater of the Colorado Plateau plays a role in the development of the current morphology of the valleys. The aquifer properties within the formations of the Colorado Plateau are discussed in order of decreasing depth. The Chinle Formation forms a barrier to flow for the overlying Wingate Sandstone aquifer [Laity, 1988a]. The
Kayenta Formation is the bounding layer between the Wingate and Navajo Sandstone aquifer system and is fractured and permeable in local areas, allowing transport from the Navajo into the Wingate [Laity and Malin, 1985; Laity, 1988a]. The Wingate Sandstone is a less transmissive sandstone than the Navajo Sandstone [Jobin, 1956], and together they comprise the N-aquifer.

The permeability of the Navajo Sandstone is affected by the grain size changes, diagenetic secondary porosity, fracturing, and bounding surfaces [Laity, 1988a]. For the purposes of this discussion, I focus on the effects of the bounding surfaces on permeability. The first-order surfaces (Figure 5.3) typically impede vertical migration of water due, and are formed by fine-grained layered sediments forming in the interdune areas (Kochel and Riley, 1988). The second-order bounding surfaces (Figure 5.3) formed between sets of crossbeds and represent layering between individual migrating dunes [Kochel and Riley, 1988, Kocurek, 1988]. Third-order surfaces (Figure 5.3) represent reorientation or scouring of the dune slip face during dune migration (Kocurek, 1988). Kochel and Riley (1988) stated that permeability is higher horizontally than vertically in eolian sandstones, and significantly greater parallel to the strike of the crossbeds than perpendicular.

Spring discharge measurements obtained from the eastern tributaries to the Escalante River varied between 0.001 - 0.002 m$^3$/s [Irwin et al., 2009]. The water in these channels appeared to be devoid of any bed or suspended load, indicating that the volume of spring flow was incapable of moving visible grains in any significant volume as to remove material from the valley. Measurements in the fluvial channels above the headwall scarps of the sapping valleys were typically >10 m wide, >1 m deep.
Additionally, imbricated rocks that had been deposited during flood events typically had A- and B-axes of 10’s of centimeters [Irwin et al., 2009]. Using the Manning relationship (a combination of the area from paleoflood surface cross-section and the entrained grain sizes to calculate discharge), it was determined that these valleys experience flood events with discharges from $1 - 10$ m$^3$/s, at least three orders of magnitude higher than the seepage undercutting the headwalls. These flood events are likely important for the removal of material from the sapping valley.

![Diagram of the typical eolian bedding with bounding surfaces and foreset bedding](image)

**Figure 5.3:** Diagram of the typical eolian bedding with bounding surfaces and foreset bedding. The first-order bounding surfaces are the interdune deposits. The second-order bounding surfaces are the separate sets of crossbeds. The third-order bounding surfaces are reactivated surfaces on the lee slope of the dune. Modified after Kochel and Riley (1988).

Martian hydrogeology is mostly hypothetical because of Mars’ current atmospheric conditions make liquid water unstable on the surface. Tanaka and Golembek (1989) suggested that near the end of the Late Heavy Bombardment Mars had a warmer climate and groundwater was nearer the surface than at present Martian
conditions (see Table 2.2). Presently, a ground ice layer stretches from a depth of 100’s of meters to a depth of ~1km in the equatorial region of Mars, and groundwater may be present under this cryosphere (Tanaka and Golembek, 1989).

MacKinnon and Tanaka (1989) modeled the Martian crust as 10 km of fractured bedrock overlain by a 1 – 2 km thick sequence of ejecta materials. The MacKinnon and Tanaka model predicted that the bedrock had a permeability of ~1000 darcies near its top, decreasing with depth to zero. The permeability of the ejecta layer in this model is dependent on grain size and packing, with a maximum permeability of 10^{-2} darcies. Rodriguez et al. (2007) suggested that wrinkle ridges breached the cryosphere and transmitted water along the subsurface faults and impact fractures. Rodriguez et al. also suggests that the faults associated with wrinkle ridges may have intercepted normal groundwater flow and focused it into other areas. Hanna and Phillips (2006), proposed a fractured megaregolith aquifer model, similar to MacKinnon and Tanaka (1989), for crater-saturated, Noachian-aged rocks and surface materials of the southern highlands. This model suggests that water percolates to depths of 15 km using impact-induced fractures in the megaregolith and in the upper basement rock. The permeabilities of this model were also similar to those found by MacKinnon and Tanaka (1989).

The valleys in the study area incise the megaregolith wherein water could be transmitted along the impact-induced fractures. The amount of the material removed from the valleys in the region, suggests higher seepage discharges or flood events capable of moving materials from the valley and depositing them 10’s to 100’s of kilometers from the source.
Valley Morphology

Most terrestrial valleys are formed by a combination of sapping and runoff [Howard, 1988b; and Ritter et al, 1995 after Schumm and Phillips, 1986]. Although overland flow is the dominant process in terrestrial drainages, sapping processes likely play a role in their initiation and maintenance [Ritter et al, 1995]. Morphologic and morphometric differences in runoff valleys and sapping valleys can indicate which process dominates the system [Mars Channel Working Group, 1983; Laity and Malin, 1985; Luo 2000]. Laity and Malin [1985], proposed characteristics differentiating sapping dominated valleys from those dominated by runoff (Figures 5.4A & B): (1) amphitheater heads of tributaries, (2) relatively constant width from head to mouth, (3) steep and high walls in the valleys, (4) persistent structural control, (5) and common hanging valleys. Groundwater sapping as a process includes: (1) emergence of groundwater at a seep face in the rock; (2) removal of material from the seep face (3) undermining of the headwall, creating an alcove; (4) collapse of the overhanging material; and (5) backwasting headward, increasing the length of the valley. Above the headwall, run-off, typically due to flooding, provides an adequate volume of water to transport smaller material from the floor of the valley.

Well-integrated runoff dominated valleys are V-shaped in both planview and cross-section with a tapered head, and increasing width with increasing distance from the head [Figures 5.4C; Malin and Laity, 1985; Howard and Kochel, 1988; Ritter, et al, 1995]. Sapping valleys are U-shaped in planview (relatively constant width from head-to-mouth) and in cross-section. Amphitheater-shaped headwalls and alcoves form due to
Figure 5.4: (A) THEMIS daytime IR mosaic (100 m/pix) showing the sapping valleys of Parana Valles. These valleys have a consistent width from head-to-mouth over 10’s to 100’s of km. (B) DOQQ photomosaic (8 m/pixel) of the distal portion of the Escalante River and a portion of the Escalante Arm of Lake Powell showing the amphitheater heads of the valleys, U-shape in planview, and stubby tributaries. (C) DOQQ photomosaic (8 m/pix) image of the area west of the Escalante River with incision from fluvial valleys with tapered heads, multiple tributaries, widening from head to mouth. North is up in frames A-C. (D) Perspective image of the alcove head in East Bowns Canyon. In addition, the impact and capacity of fluvial channels feeding the valley from the upper plateau surface is clear in this image. (E) Small exfoliation fractures at the headwall with some infilling by calcite. (F) The aquifer is likely recharged by pools in the fluvial channel above the headwall.
undermining at the aquifer-boundary layer interface (Figure 5.4D). Subsequent collapse of the overhanging scarp occurs due to lack of basal support [Howard and Kochel, 1988]. Groundwater is an effective mechanically weakening agent in the Navajo Sandstone because amorphous calcite evaporates out of the water and into pore spaces [Laity, 1983]. These deposits separate the interlocking sand grains weakening the sandstone [Laity, 1983; Laity, 1988a; Howard and Kochel, 1988]. The constant width and steep high walls of sapping valleys are maintained by undermining and collapse along exfoliation joints as well as by surface runoff [Laity, 1988; Howard and Kochel, 1988].

Sapping valleys are prone to form along structural weaknesses because of increased porosity and conductivity along fractures [Laity and Malin, 1985; Laity, 1988a]. This is evidenced by the constancy of tributary junction angles with the main trunk stream, consistent angles over a large region and parallel tributaries (Figures 5.1a). Runoff valleys are also structurally controlled on the Colorado Plateau but also form networks with topographic controls [Laity and Malin, 1985]. In addition, stratigraphic control on the development of sapping and runoff valley development is dependent on the direction of the scarp face compared with groundwater flow direction and dip of the bedding [Laity and Malin, 1988; Howard and Kochel, 1988]. If the scarp facing in the same direction of groundwater flow and down dip, a tributary sapping valley is likely to form along the joints of that side. If a scarp is facing into the opposite direction, up dip and into the direction of groundwater flow, groundwater will recharge the aquifer instead of discharging [Laity and Malin, 1985].

Sapping valleys cut into the Colorado Plateau have hanging valleys due to perched aquifer seepage and deficient downcutting due to armoring. The permeability
differences, decreased vertical migration, and possible traps, created by the two types of boundary layers in the Navajo Sandstone, produce perched aquifers seeps above the Navajo-Kayenta boundary. Seeping water from these perched aquifers and springs create terraced alcoves and subsequently produce hanging valleys [Laity and Malin, 1985]. Hanging valleys persist because of insufficient groundwater flow, and small catchment areas do not supply a sufficient volume of water to cut through bedrock, down to base level [Laity and Malin, 1985]. The upper portions of the eastern basins are characterized by a system of plunge pools connected by shallow and narrow channels (Figure 5.4F). The plunge pools have variable widths and depths, and typically form at the confluence of multiple fractures. The plunge pools seem the most likely source of recharge given that (a) the proximity of the headwall scarps to the drainage divide at the Circle Cliff Anticline; (b) the dip direction of the bedding away from the divide; (c) multiple fractures intersect at the plunge pools, which focuses water into these areas; and (d) the plunge pools are the only standing water in the area, which allows for infiltration.

Martian valleys display similar morphologies to both run-off- and sapping-dominated terrestrial valleys. The Martian fluvial valleys have typical terrestrial fluvial dendritic morphologies. The fluvial valleys are typically shallower than the sapping valleys. Additionally, distances from the head of the fluvial valley to its mouth are shorter than those of the sapping valley morphologies. Fluvial valley morphologies are typically nearer the drainage divide than valleys with sapping morphologies. The sapping valleys are U-shaped in planview (different from cross-section shape, meaning they hold a relatively constant width from head-to-mouth), are steep-walled (U-shaped in cross-section, unlike fluvial valleys which are sloped and V-shaped in cross-section), and
have very little relief from the headwall to the mouth. The majority of the slopes in the Martian southern highlands are created by impacts that are associated with radial and circumferential fracturing created by impact processes. These fractures enhance the transmission of groundwater, help valley formation, and give them their radial appearance.

Methods

Hypsometric Analyses

For this study, the extent of basin morphometry is limited to relief and area measurements. On Mars, other measurements of drainage density, stream order, and channel length are inhibited by the temporal extent of exposure to resurfacing process, i.e., impact interruptions of valleys, mantling of the channel characteristics, and insufficient resolution to map valleys in adequate detail everywhere. The area and relief measurements allow for a thorough determination of the hypsometry statistics, and can be used for a quantitative comparison with little effect from resurfacing.

Hypsometry is a measure of the ratio of basin relief to basin area to quantify the amount of material that remains in a basin after erosion (Figure 5.5). This analysis produces a curve that represents the erosion in all parts of the basin, and is indicative of the processes forming the basin. In addition to the curve, the integral of the area underneath the curve, and the skewness and kurtosis of both the hypsometric curve and first derivative of the curve can also indicate which processes formed the basin (Table 5.1). Assuming that channels initiate through runoff in areas with gently sloped environments, the hypsometric integral is a quantitative way of determining the amount of material that remains after erosion. The skewness (difference from a normal
Figure 5.5: (A) Perspective and planview showing the slices of relief and area for the hypsometric analyses. (B) Sample graph created from the sample data in frame A of this figure. The hypsometric curve is used to calculate the hypsometric integral (percent material remaining after erosion), and the hypsometric skewness (headward erosion) and hypsometric kurtosis (erosion in the central portion of the basin). (C) Upper graphs showing typical hypsometric curves from a fluvial system with increased headward erosion (left) and a more typical sapping morphology with less headward erosion (right). The lower graphs show the typical slopes of the curve derived from the best-fit polynomial for the hypsometric curve. Again, the left graph is typical of fluvial valleys, and the right graph is typical of sapping valleys. Modified from Luo (2000).
Table 5.1: Definition and Typical Values of Hypsometric Analyses.

<table>
<thead>
<tr>
<th>Calculated Parameter</th>
<th>Parameter Definition</th>
<th>Sapping Values</th>
<th>Fluvial Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hypsometric Integral</td>
<td>Material remaining after erosion</td>
<td>&gt;0.5</td>
<td>&lt;0.5</td>
</tr>
<tr>
<td>Hypsometric Skewness</td>
<td>Amount of headward erosion; upper basin</td>
<td>Lower</td>
<td>Higher</td>
</tr>
<tr>
<td>Hypsometric Kurtosis</td>
<td>Erosion in the central region of the basin</td>
<td>Similar</td>
<td>Similar</td>
</tr>
<tr>
<td>Density Skewness</td>
<td>Slope change over the entire basin</td>
<td>Negative</td>
<td>Positive</td>
</tr>
<tr>
<td>Density Kurtosis</td>
<td>Slope change in the central region of the basin</td>
<td>Higher</td>
<td>Lower</td>
</tr>
</tbody>
</table>

Adapted from Luo (2000)

... of the hypsometric curve indicates the amount of headward erosion in the upper portions of the basins, and these values are typically low for sapping valleys. Values of kurtosis (difference in the peak of the hypsometric curve from the statistical normal peak distribution) indicate erosion in the middle of the basin, and these numbers are usually lower for sapping valleys. The density skewness and density kurtosis values are derivatives of the curve and indicate the changes in overall basin slope and middle basin slope, respectively. The values for density skewness are typically negative for sapping values and positive for fluvial networks, and the density kurtosis values are typically higher for sapping valleys.

These values were calculated for the basins with visible drainages within the mapping area. The DEM was not reconditioned to fill in sinks so that the craters that are breached in the basin are included in the analysis. The methods for drawing, extracting, and calculating the statistics are described below.

Datasets

The MOLA DEM (~463 m/pix) was used to define the Martian basins, and the values for relief were extracted from the topography. An automated program in the Spatial Analyst extension of ArcMap defined the polygonal basins. The basins were...
adjusted manually to include only areas where visible valleys existed, and excluded the areas where water would have ponded (Figure 5.6).

For the terrestrial basins, ArcGIS files containing the basin polygons and the stream linework were downloaded from the USGS National Hydrography Database (URL: http://nhd.usgs.gov/). The polygonal basin data were trimmed to the features of interest, and further subdivided using ArcGIS. The 1/3 arc second (~10 m/pixel) National Elevation Dataset (NED) was downloaded from the USGS National Map Seamless Server (URL: http://seamless.usgs.gov) and projected into Universal Transverse Mercator (UTM) zone 12 North using the North American Datum (NAD) 1983 (Figure 5.7).

Calculations

For both sets of basins, the polygons were used to clip the basins from DEMs to create basin-discrete DEMs for input into the ArcGIS script. This script prompts the user for the starting, ending, and interval elevations. The starting and ending elevations describe the relief of the basin, and the interval is amount of relief per area slice. For this study, the intervals were determined by dividing the relief by $x$ (where $x$ is a variable number depending on the total relief) to produce $x$ number of slices for each basin. The script calculates the projected area of the basin by multiplying the longitudinal cell size by the cosine of the latitude, and then multiplying the latitudinal cell size. The area of the slice to the total area is calculated and the relief of a slice versus the total relief is calculated. The values for this are determined per slice, and for this study 20 points (hypsometric moments) were used to define the hypsometric curve. The script exports the values for the hypsometric moments to create the hypsometric curve, the
Figure 5.6: Location map of the 14 basins used to determine the dominant valley forming process within the map area. The red lines indicate the boundaries of the individual basins. See table 5.2 for the hypsometric values associated with each basin. Base image is the shaded-relief generated from the 473 m/pix MOLA topography. The basins were clipped out of the 473 m/pix MOLA grid fit polynomial is calculated by the script, and the derivative is taken to determine the density skewness and density kurtosis values (See Appendix D for equations).

hypsometric integral value, and the values for the hypsometric and density skewness and kurtosis. These values are plotted to generate the curve, and the eight possible graphs of hypsometric possible graphs of hypsometric attributes against each other.
Figure 5.7: Tiled DOQQ (8 m/pix) location map of the 21 basins (outlined in red) used in the terrestrial hypsometry study. The Escalante River and the Escalante Arm of Lake Powell separate the basins into eastern and western sets. The western basins have typical fluvial valley morphologies in the proximal reaches, and typical sapping morphologies in the distal reaches. The eastern basins contain classic sapping valleys with headwall alcoves, consistent widths, stubby tributaries, and gentle slopes. North is toward the upper left corner.
Results

Escalante

The results for the five hypsometric variables for each of the terrestrial basins are in Table 5.2. The results are mapped to show the distribution of the results in Figure 5.8.

The hypsometric curve, moments, and variable results, as well as a perspective view of the topography of each individual terrestrial basin are available in Appendix E. The hypsometric integral values range from 0.20 – 0.74. The hypsometric skewness values range from 0.11 to 1.04, and hypsometric kurtosis values range from 1.46 and 3.47. The density skewness values range from -0.77 to 1.37 and density kurtosis values range between 1.19 and 3.35. The curves for each basin, the hypsometric moments, and a 3-dimensional perspective view are located in Appendix E.

Table 5.2: Results of the Terrestrial Hypsometric Analysis

<table>
<thead>
<tr>
<th>Basin Name</th>
<th>Hypsometric Integral</th>
<th>Hypsometric Skewness</th>
<th>Hypsometric Kurtosis</th>
<th>Density Skewness</th>
<th>Density Kurtosis</th>
<th>Dominant Process</th>
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<td>Bowns Canyon</td>
<td>0.51</td>
<td>0.33</td>
<td>1.97</td>
<td>-0.03</td>
<td>1.33</td>
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<tr>
<td>Long Canyon</td>
<td>0.64</td>
<td>0.32</td>
<td>2.04</td>
<td>-0.52</td>
<td>1.87</td>
<td>Sapping</td>
</tr>
<tr>
<td>Cow Canyon</td>
<td>0.62</td>
<td>0.33</td>
<td>2.05</td>
<td>-0.51</td>
<td>1.86</td>
<td>Sapping</td>
</tr>
<tr>
<td>Fence Canyon</td>
<td>0.64</td>
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<td>1.95</td>
<td>-0.43</td>
<td>1.55</td>
<td>Sapping</td>
</tr>
<tr>
<td>Small Mouth Canyon</td>
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<td>0.26</td>
<td>1.92</td>
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<td>1.34</td>
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</tr>
<tr>
<td>Explorer Canyon</td>
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<td>1.86</td>
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<td>1.30</td>
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</tr>
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<td>East Side 01</td>
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<td>2.11</td>
<td>-0.67</td>
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<td>East Side 02</td>
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<td>0.01</td>
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<td>1.88</td>
<td>0.04</td>
<td>1.21</td>
<td>Sapping</td>
</tr>
<tr>
<td>Minimum</td>
<td>0.20</td>
<td>0.11</td>
<td>1.46</td>
<td>-0.77</td>
<td>1.19</td>
<td></td>
</tr>
<tr>
<td>Maximum</td>
<td>0.74</td>
<td>1.04</td>
<td>3.47</td>
<td>1.37</td>
<td>3.35</td>
<td></td>
</tr>
<tr>
<td>Average</td>
<td>0.55</td>
<td>0.33</td>
<td>2.04</td>
<td>-0.13</td>
<td>1.77</td>
<td></td>
</tr>
</tbody>
</table>
Results for the five hypsometric variables for each of the Martian basins are in Table 5.3. The results are mapped to show the distribution of the results in Figure 5.9. The hypsometric curve, moments, and variable results, as well as a perspective view of the topography of each individual Martian basin are available in Appendix F. The hypsometric integral values range from 0.32 – 0.65. The hypsometric skewness values range from -0.01 to 0.78. The hypsometric kurtosis values range from 1.83 to 2.55. The density skewness values range from -0.50 to 1.05. The density kurtosis values range between 1.17 and 2.94. The curves for each basin and a 3-dimensional perspective view are located in Appendix F.

Table 5.3: Results of the Martian Hypsometric Analysis

<table>
<thead>
<tr>
<th>Basin Name</th>
<th>Hypsometric Integral</th>
<th>Hypsometric Skewness</th>
<th>Hypsometric Kurtosis</th>
<th>Density Skewness</th>
<th>Density Kurtosis</th>
<th>Dominant Process</th>
</tr>
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<tbody>
<tr>
<td>Newcomb01</td>
<td>0.41</td>
<td>0.47</td>
<td>2.04</td>
<td>0.66</td>
<td>1.87</td>
<td>Fluvial</td>
</tr>
<tr>
<td>Newcomb02</td>
<td>0.35</td>
<td>0.62</td>
<td>2.20</td>
<td>1.05</td>
<td>2.70</td>
<td>Fluvial</td>
</tr>
<tr>
<td>Newcomb03</td>
<td>0.36</td>
<td>0.78</td>
<td>2.55</td>
<td>1.05</td>
<td>2.94</td>
<td>Fluvial</td>
</tr>
<tr>
<td>Newcomb04</td>
<td>0.33</td>
<td>0.58</td>
<td>2.12</td>
<td>1.00</td>
<td>2.45</td>
<td>Fluvial</td>
</tr>
<tr>
<td>Newcomb05</td>
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<td>0.57</td>
<td>2.11</td>
<td>0.88</td>
<td>2.17</td>
<td>Fluvial</td>
</tr>
<tr>
<td>Noachis01</td>
<td>0.55</td>
<td>0.39</td>
<td>2.11</td>
<td>-0.31</td>
<td>1.70</td>
<td>Sapping</td>
</tr>
<tr>
<td>Noachis02</td>
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<td>0.33</td>
<td>1.94</td>
<td>0.13</td>
<td>1.31</td>
<td>Fluvial</td>
</tr>
<tr>
<td>Noachis03</td>
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<td>1.84</td>
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</tr>
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<td>Noachis05</td>
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<td>1.41</td>
<td>Fluvial</td>
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<td>-0.14</td>
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<td>Sapping</td>
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<tr>
<td>Arabia02</td>
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<td>-0.01</td>
<td>1.88</td>
<td>-0.50</td>
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<td>Sapping</td>
</tr>
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<td>0.43</td>
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<td>2.00</td>
<td>0.27</td>
<td>1.43</td>
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<td>Parana</td>
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<td>1.52</td>
<td>Sapping</td>
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<tr>
<td>Minimum</td>
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<td>-0.01</td>
<td>1.84</td>
<td>-0.50</td>
<td>1.18</td>
<td></td>
</tr>
<tr>
<td>Maximum</td>
<td>0.65</td>
<td>0.78</td>
<td>2.55</td>
<td>1.05</td>
<td>2.94</td>
<td></td>
</tr>
<tr>
<td>Average</td>
<td>0.47</td>
<td>0.39</td>
<td>2.06</td>
<td>0.28</td>
<td>1.79</td>
<td></td>
</tr>
</tbody>
</table>

Discussion

These results from the Martian study suggest that nine of the basins (Newcomb: 01-05, Noachis: 02, 03, and 05; and Arabia: 03) are fluvially dominated basins. The combination of the low integral values, higher hypsometric skewness values, and positive
Figure 5.8: Escalante region basin map showing the results of the hypsometric study. Basins are categorized as fluvial (F), sapping (S), and combination (C). The blue color corresponds to basins with hypsometric statistics that indicate a fluvially dominated formation. The green color corresponds to basins with a sapping dominated formation. The yellow color corresponds to basins where the hypsometric statistics were indicative of a possible combination of fluvial and sapping formation.
Figure 5.9: Martian basin map showing the results of the hypsometric study. Basins are categorized as fluvial (F), sapping (S), and combination (C). The blue color corresponds to basins with hypsometric statistics that indicate a fluvially dominated formation. The green color corresponds to basins with a sapping dominated formation. The yellow color corresponds to basins where the hypsometric statistics were indicative of a possible combination of fluvial and sapping formation.

values of density skewness trend towards criteria established by Luo (2000) for terrestrial values, low hypsometric skewness values, and negative density skewness values, which are within the established range for sapping-dominated morphologies (Luo, 2002).
Newcomb01, Noachis01 and Noachis02 had integrals above 0.50, but had nonconforming values in hypsometric and density skewness. The range in these values may be due to a combinative formational process that was a balance of fluvial and sapping, leading to values representative of both categories. Another hypothesis is that a secondary process, e.g., subsequent burial, high occurrence of impacts, development of wrinkle ridges, infilling, etc., could play a role in these values.

The results from the terrestrial study indicate that three of the basins (Davis Gulch, Willow Gulch, and Fifty-Mile Creek) are fluvially dominated basins. The combination of the low integral values, higher hypsometric skewness values, and positive values of density skewness trend towards criteria established by Luo (2000) for terrestrial valleys with overland flow dominated basins. The remaining eighteen basins had high integral values, low hypsometric skewness values, and negative density skewness values, which are within the established range for sapping-dominated morphologies (Luo, 2000).

Laity and Malin (1985) hypothesized three of the western basins (Indian Creek, Clear Creek, and West Side01) to be fluvially dominated valleys; however, these results suggest that they are sapping-dominated valleys (Figure 5.8). Indian Creek and Clear Creek both had high integral values, low hypsometric skewness, and negative density skewness values. Whereas, West Side01 has an integral value of only 0.52, low hypsometric skewness, and positive density skewness, which are attributes from both fluvial and sapping. The structural controls (change in dip at the axes of the Fifty-Mile Syncline and the Bridge Anticline) on the west side of the Escalante river may have created an environment that allowed for at least the distal reaches of these basins to form sapping valleys while still preserving a large fluvial-type catchment area.
Past studies of Martian hypsometry by Luo (2002), Grant and Fortezzo (2003, 2004) suggested that both fluvial and sapping dominated basin morphologies were present in the Martian study area and in the surrounding regions. This study differs from both of the previous Martian studies by excluding areas beyond the mouth of the valleys. Luo (2002) examined basins within the current study area, but used computer-generated basins that concatenated multiple smaller basins. Luo (2002) also included the area beyond the mouth of valley, which would have skewed the hypsometric integral into lower numbers. Grant and Fortezzo (2003) examined basins on Mars, the Earth, the Moon, and Venus. The Martian basins were outside the current study area and included areas beyond the mouths of the valleys. Using bodies other than Mars and the Earth yielded an interesting result: Bodies with no known fluvial modification yielded similar results based on the initial topography. This led to the hypothesis that cratered terrains may “precondition” surfaces resulting in a fluvial signature from the hypsometric analysis. Grant and Fortezzo (2004) examined the longitudinal profiles of terrestrial fluvial basins and the longitudinal profiles of both Martian and terrestrial crater flanks. The results of that study showed striking similarities in the longitudinal profiles of two systems with vastly different origins and erosional histories.

Conclusions

The results of the hypsometry study indicate that the qualitatively similar valleys on Mars and on the Earth are quantitatively similar. Although the basins shapes and sizes are not similar, the erosional pattern of the valleys within the basins is analogous. Thus, it is possible to state that the formational mechanisms were likely similar. Returning to the caveats from the introduction to this chapter, analogs are rarely 100% correlative. On
both Mars and Earth, the structures, stratigraphy, hydrologic variables, climate variations, water chemistry, and gravity may cause differences in the way these basins are eroded and the process by which the valleys incised.

These results suggest that both precipitation run-off and groundwater processes were likely in the Martian past. Geomorphic evidence for both processes include valleys initiated at or nearby (fluvial), and far from drainage divides (sapping); and both terrestrial-type dendritic fluvial valleys and sapping morphologies. Precipitation in the region may have been a function of increased atmospheric density increasing the amount of available volatiles in the atmosphere, elevation changes within the region to affect the air temperature and produce moisture, and/or due to the amount of ponded water available for evaporation. There is evidence for an active groundwater system in parts of the study region. The valleys with heads far from the basin divides are likely a result of water percolating into the subsurface with seepage occurring at intersections of the Martian surface and the water table. The discussion on which process, fluvial or groundwater, dominated the hydrological regime in the Martian past may be moot, as it appears to be a basin-to-basin distinction, much the same as terrestrial basins.

This is the first study to examine the sapping valleys of the Escalante Arm of Lake Powell and in the study area for Mars. Grant and Fortezzo (2003, 2004) studied the basins to the west of the Martian mapping area with the differences noted in the discussion section. This study is the first to compare the basins and valleys of Mars with long established terrestrial analogs. Although other factors may influence the type of valleys that form in a particular watershed, basin hypsometry can be applied to determine the dominant process based on basin morphology.
Chapter 6  
SUMMARY OF CONCLUSIONS, AND RECOMMENDATIONS

Conclusions from Mapping

The geologic units within the area mostly developed in the Noachian and Early Hesperian, with the exception of some crater floors. Alteration of the surface through impact, hydrologic, volcanic processes were most active during the Noachian Age. Water ponded in basins during the Late Noachian through the Early Hesperian, typically formed by impact processes, and led to infiltration and, in some cases, formed outlets, which led to desiccated portions of the megaregolith. Valleys began to form in the Middle to Late Noachian and continued to develop into the late Early Hesperian (~4.0 – ~3.5 Ga). Valley morphologies indicated both sapping and fluvial valley development. Small impact and eolian processes have been the most active processes from the Late Hesperian to present.

Hypsometric Analysis Conclusions

Both fluvial and sapping processes form valleys in the distal reaches of the Escalante River. Sapping processes in the Escalante region are dominant in areas where the recharge zone is up-dip from base level, allowing the seep faces to undermine the rock and enhance headward erosion.

Sapping and fluvial dominated basins are both present on Mars and are dependent on the local topography. Pluvial events likely occurred on the surface of Mars during the Noachian and waned during Early Hesperian. Impact processes on Mars may have created conditions ideal for the subsurface transmission of water, and the formation of
sapping-dominated basins. This is similar to the terrestrial study area where warping of the stratigraphy created conditions favorable to groundwater sapping processes.

This study confirms that hypsometry is a useful tool in establishing a quantitative method for categorizing morphologies on both the Earth and Mars. For Mars, the hypsometry analysis indicates that both fluvial and sapping processes were actively forming valleys in the Late Noachian and Early Hesperian. Although hypsometry is being used to determine the dominant process, other geologic controls (e.g., dip, dip direction, sedimentation, original topography, etc.) are also important factors to consider when interpreting the results of this type of study.

**Recommendations for Future Work**

Within the Martian study area, volume measurements for basins where ponding occurred are needed to determine the minimum amount of water present. This number could then be compared to reasonable discharge rates from valleys flowing in and out of the basin. This would further constrain the duration of activity of water in this region of Mars. Additionally, further morphometry measurements from the fluvial basins above the headwall would allow for further constraint of the degree to which structure, slope, shape, and topography control whether the basins have a fluvial or sapping hypsometric signature.

For the terrestrial basins, more information is needed on the structural constraints on the formation of sapping valleys. There are more examples of these morphologies on the Colorado Plateau, and it would be interesting to compare the dip directions, channel slopes, structural controls and recharge in other locations to the Escalante region. To constrain further the seasonal variations, it would be ideal to set up permanent flume sites
to establish base flow, and monitor the sediment load of the seepage faces. Additionally, hydrologically-triggered cameras set up in areas of confined flow, with known cross-sectional areas in the fluvial channels above the sapping valleys would allow for further constraint on the volume and frequency of flood events spilling into the sapping valleys.
References


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