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Morphologic and structural mapping of layered central uplifts on Mars

Anna M. Nuhn  
The University of Western Ontario

Supervisor  
Dr. Gordon Osinski  
The University of Western Ontario

Joint Supervisor  
Dr. Livio Tornabene  
The University of Western Ontario

Graduate Program in Planetary Science

A thesis submitted in partial fulfillment of the requirements for the degree in Master of Science

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MORPHOLOGIC AND STRUCTURAL MAPPING OF LAYERED CENTRAL UPLIFTS ON MARS

(Thesis format: Integrated Article)

by

Anna Nuhn

Graduate Program in Geology: Planetary Science

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science

The School of Graduate and Postdoctoral Studies
The University of Western Ontario
London, Ontario, Canada

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Abstract

Central uplifts within complex impact craters on Mars and Earth have been studied for many decades. Nevertheless, their formation is still poorly understood (regarding target rock weakening mechanisms), as well as the impactites found within, them including their extent and distribution. We have cartographically approached these questions by mapping morphologies and structures within three impact craters on Mars, utilizing high-resolution images taken by the Mars Reconnaissance Orbiter combined with other data sets, while making comparisons to terrestrial field-based observations and measurements. Betio (281.38° E, -23.15° N), Byala (293.54° E, -25.75° N), and Halba (303.91° E, -26.03° N) impact craters are all ~30 km in diameter, located within the Hesperian-aged Ridged Plains unit in Thaumasia Planum, Mars, and contain central floor pits that expose megablocks of layered bedrock. A comparison of three craters of similar diameters, target lithologies, and geographical regions allows us to investigate the morphologic and structural similarities and differences found between the three craters. Our mapping reveals a variety of faults, folds (likely radial transpression ridges), multiple interpreted breccia dykes, in addition to different types of interpreted impactites (e.g., breccias and melts –clast –free and –rich, pitted materials and uplifted bedrock (i.e., parautochthonous bedrock). Through structural mapping we have found that smaller (60–300 m in diameter) blocks with high dips of ~ 45° to 85° are present proximal to the relatively flat floor pit and larger (>800 m in diameter) blocks with shallow dip angles of ~ 5° to 15° occur in the outer sections of the floor pit, similar to terrestrial observations. Deformation mechanisms that aided in the uplift and collapse include brittle deformation, seen increasing towards the centre of the crater, a decrease in block size closer to the centre, and the presence of fault-bounded blocks.

Keywords

Mars, Impact Cratering, Central Uplifts, Central Floor pits, Layered Bedrock
Co-Authorship Statement

Chapter 1 was a compilation of the existing literature relevant to the topic of this study. Dr. G. R. Osinski and Dr. L. L. Tornabene contributed to the editing of the chapter.

Chapter 2 is an article currently in review for publication. The title of the article is “Morphologic and Structural Mapping of the Central Uplift of Betio Crater, Thaumasia Planum, Mar”, and co-authors of the paper are Dr. G. R. Osinski, Dr. L. L. Tornabene (Western University, London, ON), and Dr. A. S. McEwen (University of Arizona, Tucson Arizona, USA.). All data for both Chapters 2 and 3 was acquired by A. Nuhn. Dr. L. L Tornabene and Dr. G. R. Osinski provided input into interpretations and guidance as to the overall structure and content of the both chapters, along with editorial input.

Chapter 4 was a summary on the contributions of the study. Dr. G. R. Osinski and Dr. L. L. Tornabene contributed to the writing and editing of the chapter.
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Chapter 1

1 Introduction

Impact cratering is one of the most common geological events in our solar system that affects almost all surfaces of satellites, planets, and asteroids. The impact cratering process involves projectiles (comets or asteroids) that collide with a target body at high velocities. During these events, the surface materials experience variable shock pressures, many orders of magnitude greater than pressures rocks experience during regular, endogenous metamorphism (e.g., Stöffler, 1971). This results in permanent changes within the target rock that range from brittle deformations to melting and vaporization.

It is well known that impact craters display two main characteristic morphologies: simple and complex. Whereas, simple craters comprise a relatively simple bowl-shaped form, complex craters possess an uplifted rim, terraced walls, down-faulted annular trough containing impact melt rocks and/or breccias, and an uplifted structure on the crater floor referred to as a central uplift (e.g., Dence, 1968; Melosh, 1989). The central uplifts of complex craters are of considerable scientific interest. Central uplifts exhume deep-seated bedrock from the subsurface that is often not otherwise exposed at the surface. They also preserve morphologic and structural features from both the pre-impact target and those formed during the impact process. In addition, central uplifts provide habitats for life through the exhumation of altered materials or via impact-generated hydrothermal systems (Cockell et al., 2013; Tornabene et al., 2013; Osinski et al., 2013). On Earth, central uplifts may also contain potential economic mineral and hydrocarbon deposits (Grieve, 2013). On Mars, central uplifts are known to be morphologically diverse with these features not only being comprised of a peak(s), but also pit morphologies including summit pits and central floor pits (see Barlow, 2010). Despite their importance, much remains unknown about the detailed mechanics and processes of central uplift formation.

On Earth, very few central uplifts are exposed and accessible and only a fraction of those are geologically and structurally mapped in detail (Kenkmann et al., 2005). The high levels of erosion (from ice, water and wind), volcanic activity, vegetation, and plate
Tectonics throughout Earth’s history make it challenging to locate and observe the bedrock, deposits and structures that occur in well-preserved impact craters. Fortunately, studies of impact craters on Earth and other planets compliment each other. Terrestrial craters provide critical ground-truth, as they can be mapped in the field, drilled, sampled, and analyzed, while the structure and morphology of the best-preserved craters on other planetary bodies can be observed in great detail through increasing availability of higher resolution orbital data, including: visible images, digital elevation, thermal, and spectral data from past and ongoing missions.

This study focuses on the morphologic and structural mapping of central uplifts in three nearly equal sized craters (~30 km diameter), similar target materials, and similar regions south of Valles Marineris in the Thaumasia Planum region of Mars. Detailed mapping permitted the morphological interpretation of pre-, syn- and post-impact deposits and the structural characteristics of the uplifts of the craters Betio, Byala, and Halba to be discerned. By assessing these interpretations, it has been possible to discuss topics such as potential weakening mechanisms aiding in uplift formation, central floor pit formation mechanisms and the various stages of preservation, erosion, and degradation occurring within each crater. This current chapter, Chapter 1, provides a detailed review on the geology of Mars and specifically the Thaumasia Planum region, the impact cratering process, central uplift formation, central pit formation, and crater related impactites. Chapter 2 focuses on the morphologic and structural mapping of Betio crater, providing detailed observations and interpretations that aid in better understanding possible weakening mechanisms and structural deformation occurring during the modification stage that result in the central uplift (floor pit morphology). Chapter 3 focuses on three mid-sized craters (Betio, Byala, and Halba) and their central floor pits. It focuses on the morphological and structural similarities and differences along with evaluating the various floor pit formation models, based on observations we made in the three craters and in the surrounding region. Chapter 4 brings together Chapters 2 and 3 and discusses the results of this thesis in the field of planetary surface processes, specifically impact cratering, along with some concluding remarks.
1.1 Mars

Mars (Fig. 1.1) is the forth-closest planet to the sun in our solar system, with a surface gravity of 3.71 m s\(^{-2}\), a mean radius (3389.5 km) of about half the size of Earth’s, and a very thin atmosphere comprised mainly of CO\(_2\) (95%) with various other trace elements including: N\(_2\), Ar, O\(_2\), and H\(_2\)O making up the remaining 5%. Mars has two small satellites, likely captured asteroids, Phobos and Deimos. Mars is thought to have shared a similar earlier history to Earth, that was once quite geologically active with volcanoes and tectonics and possibly a water-rich environment. However, now the planet is dominated by rock and dust with minimal visual obstructions (e.g., vegetation) and geologically slower erosional events (compared to Earth), allowing for geological structures to be relatively well preserved.

![Mars](http://astrogeology.usgs.gov/maps/mars-viking-hemisphere-point-perspectives)

Figure 1.1: Mosaic of the Cerberus and Schiaparelli hemispheres of Mars projected into point perspective. Image from: http://astrogeology.usgs.gov/maps/mars-viking-hemisphere-point-perspectives.

1.2 Remote sensing in planetary studies

Remote sensing data sets are essential in planetary studies and, even in some cases, regarding remote places on Earth to understanding the geology of surfaces, particularly when directly accessing surface rocks and when conducting laboratory analyses is not
possible. Common types of remote sensing data sets include the measurement of the interaction of photons of various energies from different ranges of the electromagnetic spectrum (ultraviolet, visible, infrared, x-ray, etc.) with matter (i.e., in this case, planetary surfaces and their geologic materials). Using a combination of these data sets and their derived data products (e.g., digital elevation models (DTM), surface temperature, thermal inertia, etc.) enables us to characterize and map geological units and structural features located on other planetary surfaces.

1.2.1 Martian datasets

Orbital and landed spacecraft’s have been collecting data from Mars for more than 50 years and have enabled us to study the planet remotely and in situ. The data from the cameras and sensors onboard the Mars Reconnaissance Orbiter (MRO), Mars Express, Mars Odyssey, and Mars Global Surveyor (MGS) (Table 1.1) are used in this study. Each of these provide various levels of detail and scale to ensure full coverage of the central uplifts and context for the craters that contain them. Images from the High Resolution Imaging Science Experiment (HiRISE) onboard the MRO with a resolution up to ~0.25 m/pixel [McEwen et al., 2007, 2010]) are capable of resolving meter-scale features on central uplifts that enable detailed morphological and structural mapping. These HiRISE images were supplemented by various other data sets and derived data products (Table 1.1) including, elevation data (Digital Terrain Models [DTMs] derived from the Mars Orbiter Laser Altimeter [MOLA~462 m/pixel]; [Smith et al., 2001], High Resolution Stereo Colour Imager [HRSC~150–175 m/pixel] [Neukum & Jaumann, 2004]), thermal infrared (~100 m/ pixel) images from the Thermal Emission Imaging System (THEMIS) (Christensen et al., 2004), and context camera (CTX) images (~6 m/pixel) (Malin et al., 2007). In addition, we utilized THEMIS-derived thermal inertia measurements (Fergason et al., 2006; Putzig & Mellon, 2007), which represents the resistance to change in temperature of the upper few centimeters of the surface throughout the day, independent of local time, latitude, and season to understanding the near-surficial geology and recent processes that are potentially still active today. This data product is sensitive to surfaces containing variable grain sizes and varying degrees of induration/lithification and, thus, is used to distinguish between bedrock and fine-grained unconsolidated units.
Table 1.1: Orbital datasets used in this study.

<table>
<thead>
<tr>
<th>Camera</th>
<th>Orbiter</th>
<th>Mission Duration</th>
<th>Spatial Resolution</th>
</tr>
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<tbody>
<tr>
<td>HIRISE</td>
<td>Mars Reconnaissance Orbiter (MRO)</td>
<td>2006-Present</td>
<td>~25 - 32 cm/pixel</td>
</tr>
<tr>
<td>HIRISE DTM</td>
<td>Stereo-pair Product</td>
<td>2006-Present</td>
<td>~2 m/pixel</td>
</tr>
<tr>
<td>CTX</td>
<td>Mars Reconnaissance Orbiter (MRO)</td>
<td>2006-Present</td>
<td>~6 m/pixel</td>
</tr>
<tr>
<td>CRISM</td>
<td>Mars Reconnaissance Orbiter (MRO)</td>
<td>2006-Present</td>
<td>~18.4 m/pixel</td>
</tr>
<tr>
<td>HRSC DTM</td>
<td>Mars Express</td>
<td>2003-Present</td>
<td>150–175 m/pixel</td>
</tr>
<tr>
<td>THEMIS VIS + IR</td>
<td>Mars Odyssey</td>
<td>2001 - Present</td>
<td>~100 m/ pixel (IR)</td>
</tr>
<tr>
<td>MOLA</td>
<td>Mars Global Surveyor</td>
<td>1997 - 2006</td>
<td>~462 m/pixel</td>
</tr>
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1.3 Geologic history of Mars

On Earth, a geologist would interpret the geology of a region equipped with knowledge of the lithology, chemistry, and mineralogy, and knowledge of the subsurface. Unfortunately for most of Mars we have information only on properties that can be determined remotely, such as morphology, elevation, and spectral signatures. While interpretation of surface morphology can be straightforward (e.g., large volcanoes, impact craters, etc.) it is more often not, so many aspects of the geological history of Mars are poorly understood.

The geologic history of Mars is divided into three main time-stratigraphic eras (Fig. 1.2): Noachian (older than ~3.9 Gy), Hesperian (~3.9–3.0 Gy), and Amazonian (~3 Gy to present) (Tanaka, 1986). The Noachian is thought represent the time of the heavy bombardment, when impact rates were much higher than they were for the rest of the planet’s history. During this time the majority of Martian impact craters and large impact basins (e.g., Hellas) formed, now seen in the heavily cratered terrains in the southern hemisphere (Fig. 1.3) (Nimmo & Tanaka, 2005). There is also the assumed presence of surface water (Carr, 2006), showcasing branching valley networks found mainly in the Noachian cratered uplands, thought to have formed when surface conditions were wetter and warmer then they are today. The Hesperian era refers to the oldest surfaces that postdate the end of the heavy bombardment. This era is characterized by slowed rates of impact events, weathering, and erosion compared to the Noachian, and high rates of volcanic activity (which drops off by a factor of 10 during the Amazonian age) (See Carr & Head, 2010). Characteristic plains areas such as, Tharsis and Elysium showcase this
volcanism. The rates of volcanism on Mars are much lower than those on Earth. Greeley and Schneid (1991) estimated that roughly $6 \times 10^7$ km$^3$ of lava have accumulated on the Martian surface since the end of the heavy bombardment and that the average extrusion rate was 0.016 km$^3$ yr$^{-1}$, for comparison it is roughly 30 km$^3$ yr$^{-1}$ on Earth for the last 180 Myr (Sclater et al., 1980). The youngest era is called the Amazonian, characterized by low rates of meteorite and asteroid impacts and by cold, hyperarid conditions broadly similar to those on Mars today (Tanaka, 1986; Carr, 2006).

Figure 1.2: Mars geological timescale versus Earths geological time –scale (http://palaeos.com/precambrian/planetary.html).
1.3.1 Martian Hesperian ridged plains stratigraphy

The impact craters in this study are located within the Hesperian ridged plains (Hr) geologic unit of Thaumasia Planum (Fig. 1.4), originally defined by Tanaka (1986) and Greeley & Guest, (1987), which is characterized by smooth plains between regularly spaced compressional wrinkle-ridges. The commonly accepted method of their formation involves contractional deformation (e.g., Plescia & Golombek, 1986; Watters, 1988) that affects the underlying and surface stratigraphy resulting in differing layer strengths, as when a stronger lava layer overlies weaker sediments (Golombek et al., 2001). This unit has been mapped over extensive regions of Mars and has been interpreted to be volcanic flood lavas, formed by the massive volcano-tectonic complex well known for prolonged volcanic episodes of flood basalts and eruptions from shield volcanoes, such as Olympus Mons (Tanaka, 1986; Greeley & Guest, 1987; Keszthelyi & McEwen, 2006). Some of the best exposures of layered volcanics can be found in Valles Marineris (e.g., Beyer & McEwen, 2005; Caudill et al., 2012).
Figure 1.4: Mars geologic map (Scott & Tanaka, 1986) with a cropped close-up of the Tharsis region, showing the Hesperian ridged planes (Hr) geological unit in pink and stars indicating the location of Betio, Byala, and Halba craters.

This geologic unit is interpreted to be analogous to terrestrial Large Igneous Provinces (LIPs) (e.g., Caudill et al., 2012), where sequence stratigraphy of lava flows with intervening volcanic ash, sediment, and soil layers are known as cooling units (Nichols, 1936; Self et al., 1997). These stratigraphy sequences suggest temporal separation in emplacement and subsequent cooling of lava flows. There is one impact crater found in LIPs on Earth – Lonar Crater, India (Fig. 1.5) – however, this crater is a simple crater only ~2 km in diameter and does not contain a central uplift.
1.4 Impact cratering on Mars

Impact craters were initially identified on Mars by the Mariner 4 spacecraft in 1965, with early observations showing that Martian craters are generally shallower and smoother than lunar craters, indicating that Mars has a more active history of erosion and deposition than the Moon (Carr, 2006). As previously mentioned, the majority of the craters present on Mars have survived since the heavy bombardment, which ended roughly 3.8 billion years ago. More sparsely cratered terrains, e.g., the plains (volcanic in origin) contain younger craters from the Hesperian and Amazonian.

1.5 Mechanics of impact crater formation

The formation of an impact crater begins when a projectile (comet or asteroid) first contacts a planetary surface (target) and ends with the final motions of debris around the crater. There are three main stages (Fig. 1.6) leading to the formation of an impact crater on a surface (e.g., Gault et al., 1968; Melosh, 1989): (i) contact and compression of the projectile into the target; (ii) excavation and formation of a transient cavity; and lastly (iii) crater modification. The entire process of crater formation is instantaneous in a geological context and the transition between each stage is a continuum. The following sections describe each of the impact cratering stages in additional detail.
Figure 1.6: Series of schematic cross-sections depicting the three main stages in the formation of impact craters. This multi-stage model accounts for melt emplacement in both simple (left panel) and complex craters (right panel). For the modification stage section, the arrows represent different time steps, labelled ‘a’ to ‘c’. Initially, the gravitational collapse of crater walls and central uplift. (a) Results in a generally inwards movement of material. Later, melt and clasts flow off the central uplift. (b) Then, there is continued movement of melt and clasts outwards once crater wall collapse has largely ceased (c) Modified from Osinski et al. (2011) caption and figure from Osinski and Pierazzo, (2013).
1.5.1  Contact and compression

During the initial contact and compression stage, the projectile penetrates into the target, from as much as one to two times the diameter of the projectile (Kieffer & Simonds, 1980; O’Keefe & Ahrens, 1982). High-pressure shock waves form, spreading through the projectile and into the target (i.e., the planetary surface) (Fig. 1.6). It is during this stage that the maximum impact shock pressures and temperatures are experienced by the target surface (Melosh & Ivanov, 1999). After the waves pass through the projectile and reach its rear surface, they are reflected back through the projectile (as rarefaction waves) vaporizing the projectile (Ahrens & O’Keefe, 1977; Melosh, 1989). Generally, the time it takes for this to occur depends on the size, composition, and velocity of the projectile, but typically lasts for a fraction of the entire crater forming process (Melosh, 1989). The area of target surface immediately surrounding the projectile is also vaporized or melted (Ahrens & O’Keefe, 1977; Grieve et al., 1977). With greater distances from the projectile (decreasing shock pressures and temperatures), the target materials are displaced and fractured due to the passage of shock waves (the displaced rocks are called “impact breccias”; Stöffler, 1971). More details about the types of deformations within target rocks are listed in section 1.7.1.

1.5.2  Excavation

The short-lived contact and compression stage grades immediately into the excavation stage. It is during this stage that the transient crater is opened up by complex interactions between the expanding shock and refraction waves and the original ground surface (Fig. 1.6) (Melosh, 1989). The projectile is surrounded by a roughly hemispherical envelope of shock waves that travel both downward and outward as well as upward and outward, which eventually intersects the original ground surface (i.e., free-surface) and reflects back downwards as rarefaction (release) wave (Melosh, 1989). As this shock wave expands it declines in strength, degrading to a plastic wave and finally an elastic wave (Melosh, 1989). The shock wave is detached, equivalent to a shock front, and maintains an approximately constant thickness as it weakens (Melosh, 1989). These complex processes drive the target rock upward and outward from the impact point, producing a symmetric excavation flow around the centre of the evolving structure. Exact flow
directions vary with location within the target rocks (Fig. 1.6).

In the upper level, target material moves dominantly upward and outward (excavated zone) and become ejected beyond the transient cavity rim to form the continuous ejecta blanket (see Section 1.9). In the lower levels, target material moves dominantly downward and outward (displaced zone) to form the base of the expanding cavity (Grieve et al., 1977; Stöffler et al., 1975). The majority of this displaced zone comprises target rocks that are shocked to relatively low to intermediate shock levels and they ultimately come to form the parautochthonous (moved but appear to be in place) rocks in central uplift structures, in the case of complex craters. The highest recorded shock levels are found in the melt-rich materials that line the transient cavity do not escape and, ultimately, form the allochthonous (formed elsewhere and moved to their current location) crater-fill deposits in simple and complex impact structures (Grieve et al., 1977; Melosh, 1989), and may also be caught by the uplift.

Eventually, the velocity of the cratering flow field diminishes to a point when it can no longer excavate or displace target rock and melt, respectively. For large hypervelocity impacts, including all those on Earth, gravity controls this limit so that excavation stops when insufficient energy remains to lift the overlying material against the force of its own weight. At the end of the excavation stage, a mixture of melt and rock debris forms a lining to the transient cavity. The obliqueness and the presence of layering and pre-existing structures in the target has also been shown to affect the excavation flow field (e.g., Anderson et al., 2004).

### 1.5.3 Modification

The final part of crater formation is the modification stage (Fig. 1.6). Depending on the size of the transient cavity, the gravitational effects of the planetary body, and the properties of the target rock, the resulting impact crater feature can have variable morphologies (Fig. 1.7) – starting with a simple bowl shape cavity (simple crater) and transitioning to a cavity with an interior uplift and terraced walls (complex crater) (Melosh, 1989).
1.6 Impact crater morphology

As previously mentioned, impact craters can be classified into two main groups: simple or complex (Fig. 1.7). The simple to complex transition size is different for each planetary surface and is predominantly governed by gravity (Pike, 1980a,b; Melosh, 1989). On Mars, simple craters—bowl-shaped depressions filled with impact melt deposits overlaid on brecciated rocks, and smooth walls lacking terraces (Pike, 1980a,b)—transition into complex craters containing terraced walls, flat floors, and central uplifts in the range ~5–8 km, as compared to Earth at ~2–4 km (depending on the target), and 15–25 km on the Moon (Pike 1980a,b). Lastly, basins with two or more concentric mountainous rings occur from ~150–1,200 km diameter on Mars. For the purpose of this study we will be focusing on mid sized complex craters with diameters of ~30 km.

Figure 1.7: Comparison of simple and complex crater morphologies. (a) Unnamed simple crater on Mars (38.7 N/316.1 E) displaying an elevated crater rim and steeply dipping upper cavity walls (HiRISE image). (b) The complex impact crater Aristarchus on the Moon, showing a central peak, a flat crater floor and an extensive slump terrace zone (Kaguya/SELENE image). Note the different scale bars in the two images (From Kenkmann et al., 2014).
1.6.1 Mid-sized complex craters central uplifts

As previously stated, complex impact structures display a complicated form, characterized by a centrally uplifted region, a generally flat brecciated and melt-covered floor, and extensive terraces around the rim (Fig. 1.7b) (Dence, 1968; Grieve et al., 1977; Pike, 1980a,b). The morphology of the central uplifts within complex impact craters on Mars manifest as a peak, summit pit, or a floor pit (discussed further in section 1.8). The bulk of the uplift is comprised of bedrock excavated from great depths beneath the centre of the transient cavity that rises during the modification stage. The uplift depth is typically ~1/10 the final crater diameter on Earth (Melosh & Ivanov, 1999; Grieve & Pilkington, 1996). A statistical relationship for the amount of structural uplift, derived from studies of well-constrained complex impact structures on Earth (Grieve & Pilkington, 1996), gave SU = 0.086 D^{1.03}. For mid-sized (~30 km) impact structures, these relations imply that the crustal rocks beneath the structure are uplifted vertically by ~3 km during the impact event. However, it should be noted that it is currently unknown as to whether this relationship is applicable to Mars or any other planetary object besides Earth (Grieve & Pilkington, 1996).

Both theoretical and field studies indicate that central uplifts form in only a few minutes, almost instantaneously by geological standards, even in the largest structures (Melosh, 1989), however, the details of this process are still the subject of debate. The formation of a central uplift is a complex process that occurs during the end of the excavation stage into the crater modification stage (Melosh & Ivanov, 1999; Kenkmann et al., 2014), where two competing processes occur; a downward-directed gravitational collapse of the inner rim along concentric faults and an inward and upward movement of large (~100 m – 1 km scale) fault-bounded blocks (Melosh, 1989; Melosh & Ivanov, 1999). To explain the magnitude of inward and upward motion of rocks, a temporary strength degradation must occur (e.g., Melosh, 1977; McKinnon, 1978). This momentarily reduces friction both within a rock mass and along fault zones causing the rocks to deform by localized brittle faulting. However, in impact events, these processes take place in solid rock and may operate over distances of tens to hundreds of kilometers. A recent review paper by Kenkmann et al. (2014) outlines the various possible weakening mechanisms operating
1.7 Central uplift formation: Weakening mechanisms during cavity collapse and target uplift

Our understanding of the impact process has significantly improved over the last decade with more complex numerical simulations (e.g., Melosh & Ivanov, 1999; Collins et al., 2004; Senft & Stewart, 2009). One of the best developed models to explain crater collapse is the block oscillation model, whereby impact induced pressure vibrations in the target causes the target material to temporarily oscillate similar to waves in a fluid, although the material deforms as blocks of rock (Ivanov & Kostuchenko, 1997). For the block oscillation model to be valid, “the sound speed of the matrix between blocks must be much smaller than that of the intact rock” (see Melosh & Ivanov, 1999), which translates to a breccia zone ~10–20 % of the block’s thickness (i.e., ~100–200 m of breccia for a 1 km size block) must be present. Another earlier model known as acoustic fluidization has also been suggested by Melosh, (1979) to explain this temporary weakening mechanism, whereby the strong shaking caused by the passage of impact-generated acoustic waves through fragmented rock debris reduces the overburden pressure, allowing fluidization.

Other weakening mechanisms include thermal weakening (see O’Keefe & Ahrens, 1999) due to localized frictional heating, and strain rate weakening. The latter model predicts that the dominant weakening mechanism along faults at high slip velocities in target impact craters (100-km) in crystalline lithologies is frictional melting (Senft & Stewart, 2009). The strength of rock drops considerably, as their temperature approaches the melting point (e.g., Stesky, 1974). Shock heating and the heat that remains in the rocks after adiabatic decompression are included in all numerical models in use today. Thus, only in very large impacts like the 200 km diameter Vredefort impact structure is shock and post-shock heating an important weakening mechanism, as most rocks that have been affected by a strong shock (>25 GPa) are within the excavation flow and become ejected (e.g., O’Keefe & Ahrens, 1999). Field observations of frictional melts at several large craters have been observed (e.g., Spray & Thompson, 1995). Like any numerical model, there are some issues when applying these models to real-world scenarios and verifying
their results. It is possible that multiple processes dominate in different lithologies to cause crater collapse.

1.7.1 Macroscopic brittle deformation during crater modification

1.7.1.1 Blocks

Evidence from studies of impact structures formed in sedimentary rocks, in which the actual uplift of key stratigraphic markers has been established through drilling and geophysical studies (e.g., Milton et al., 1972; Grieve et al., 1981; Grieve & Pilkington, 1996), has shown how the target underneath the crater floor is fragmented into blocks, in particular in the central uplift. These blocks are commonly internally deformed at the millimeter to decametre scale, rather than being entirely rigid and bounded on all sides by faults. An average block size of ~100 m was determined from the Vorotilovskaya Deep Borehole (5374 m), drilled through the central uplift of the 80 km diameter Puchezh-Katunki impact crater in Russia (Ivanov et al., 1996). Mapping at Upheaval Dome, U.S., (7 km diameter) (Kenkmann et al., 2005) and Waqf as Suwwan, Jordan, (6 km diameter) (Kenkmann et al., 2010) revealed block sizes of ~50 to 100 m, with evidence of both a lithological control on block size (smaller blocks were observed in limestone relative to chert).

1.7.1.2 Impact structures: faults, folds, fractures, dykes

As previously mentioned during the excavation stage, the attenuating shock wave damages and deforms the rocks developing a complex network of faults, folds, fractures, and dykes. Numerical models by Collins et al. (2004) and Kenkmann et al. (2012) show that the intensity of impact deformation increases from the rim to the center, growth and collapse of the transient cavity leads to an accumulated strain in the material underneath the crater (Collins, et al., 2004; Kenkmann et al., 2012). This is associated with a transition from localized brittle faulting to a more widespread cataclasis and granular flow. The inner crater shows an increase in brittle deformation and blocks that are smaller or more internally damaged. Extensive cataclasis and granular flow (grain size scale) is present between the blocks (Kenkmann, 2003), as observed at Upheaval Dome in the innermost strata (Kenkmann et al., 2005). Regardless of the specific weakening
mechanism, good agreement with observations indicates that motion along strain-rate weakened faults (see Senft & Stewart, 2009) is the dominant process controlling the collapse of large impact craters (Kenkmann et al., 2014).

1.8 Central pits

Central pit morphologies are defined as a depression either into the crater floor, or atop a central topographic rise (e.g., Wood et al., 1978; Barlow, 2010) (Fig. 1.8), which are referred to as “floor pits” and “summit or peak pits”, respectively. They are common on Mars, Callisto, and Ganymede and were first identified on Mars using Mariner 9 (Smith, 1976) and Viking images (Hodges et al., 1980; Hale & Head, 1981; Hale, 1982, 1983).

Pit craters are broadly distributed on Mars (~80° to 80 °N), though they are more numerous at low-to-mid-latitudes (~40° to 40 °N) (Barlow, 2010; Robbins & Hynek, 2012). Like all central uplift structures, their morphology is likely governed by the size of the transient cavity, and properties of the target rock lithology, including: strength, porosity, compositional variations and rheology (e.g., Melosh 1989; Grey & Burchell, 2003; Senft & Stewart, 2009; Garner & Barlow, 2012).

The formation mechanisms of central pits are not well constrained but multiple hypothesis have been suggested since their initial discovery. These formation mechanisms include: 1) Ice vaporization model (Kieffer, 1977; Smith & Hartnell, 1978; Wood et al., 1978); 2) Layered target and collapse model (Hodges, 1978; Passey & Shoemaker, 1982; Greeley et al., 1982; Whitehead et al., 2010); 3) Melt-drainage model (Croft, 1981; Bray, 2009; Senft & Stewart, 2009, 2011; Elder et al., 2012); and 4) Differential erosion of uplifted materials (e.g., Whitehead et al., 2010).
Figure 1.8: Summit pit vs. Floor pit morphologies of central uplifts. (A) Unnamed crater (~17 km in diameter) in Thaumasia Planum with a summit pit central structure (image ID: CTX G15_023951_1617_XN_18S055W). (B) Betio crater (~31 km in diameter) located in Thaumasia Planum with a floor pit morphology (image ID: CTX B18_016805_1567_XN_23S078W) please note that there is a ~4.5 over printing simple crater to the east of the floor pit. (C) 3-D model of unnamed crater (seen in figure A) showing the structural uplift with a pit on top (DTEED_027485_1610_026496_1610_A01 with ESP_026496_1610_RED draped over with a VE of 1.5x). (D) 3-D model of Betio crater (eastern side) as seen in figure B, showing the depression in the crater floor with a relatively flat bottom (DTEEC_01805_1565_017583_1565_A01 with ESP_016805_1565_RED_A_01_ORTHO draped over with a VE of 1.5x)
Figure 1.9: Floor pit schematic model showing the emplacement of crater fill and ejecta. Please note the location of crater fill within the floor pit (altered after Osinski & Pierazzo, 2013)

1.8.1 Floor pit formation models

1.8.1.1 Ice vaporization

Ice vaporization model suggests that central pit crater morphology on Mars may be due to the interaction of an expanding transient cavity with a sub-surface layer or zone of ice. The near adiabatic decompression and generation of heat during impact is hypothesized to result in the explosive decompression of sub-surface volatiles on release from this high pressure (Kieffer, 1977). The upper rock layers uplift to form a central peak at the same time as the volatile-rich materials at the core of the peak is lost via vaporization (Wood et al., 1978). For this model to be plausible, one would expect that the volume of central pits in Martian craters to be controlled only by the availability of ground ice and volatile content in the target rock. The amount of ground ice available on Mars is expected to vary both with latitude and with time (Baker et al., 1991; Byrne et al., 2009). Increasing the axial tilt of Mars up to 45°, which is representative of high obliquity, would control climatic variations responsible for the deposition of ice in the equatorial region (0°–30° latitude), as well as the depletion of ice reservoirs during periods of low obliquity (Levrard et al., 2004; Forget et al., 2007).
Evidence supporting this model comes from numerical modeling of impacts into pure ice (Senft & Stewart, 2008) and mixed ice-soil (20% ice, 80% basalt) (Pierazzo et al., 2005) craters 30 km in diameter, confirming that the region underlying the center of the transient cavity experiences temperatures high enough for volatile vaporization suggesting collapse occurs either during or shortly after crater formation. Global studies conducted by Barlow et al. (2010) show that most Martian central pit craters occur in the ~7 to 30-km-diameter range. Barlow (2010) utilized standard structural uplift estimates (Pinklington & Grieve, 2004) and found that these craters may excavate to depths between ~0.7 and 2.5 km, which, is within the region where subsurface ice is expected to exist on Mars. It should be noted that this calculation is based on observations of terrestrial impact craters and may not be applicable to Mars. Finally, the lack of up-domed floors (seen in more ice-rich targets such as Ganymede) in Martian floor pit craters (Kagy & Barlow, 2008), supports suggestions of Martian regolith ice content values of ~20%, which is the value used in the simulations conducted by Pierazzo et al. (2005).

1.8.1.2 Layered target model

The layered target model predicts that central pits are the result of collapse of the central uplift due to pre-existing weaknesses in the target in the form of stronger material overlying a weaker one (Greeley et al, 1982; Passey & Shoemaker, 1982). Greeley et al. (1982) performed a series of gas-gun experiments into differently layered targets (e.g., water, clay, sand and ice in a variety of different layering combinations), which produced some crater forms deemed analogous to central pit craters, supporting that layering within a target has a direct effect on crater morphology at laboratory scales. However it should be noted that no studies into a layered volcanic terrain have been conducted.

Global studies by Barlow, (2010) and Whitehead et al. (2010) noted a concentration of floor pit craters in units mapped as volcanics rather than sedimentary targets (Barlow, 2010). It should be noted that craters emplaced into lava plains are also better preserved than those into sedimentary terrains, which are easily eroded (Whitehead et al., 2010; Barlow, 2013). Contradictions to this model include, larger scale numerical modeling of
impacts into layered ice and water targets, which has not yielded any large final craters that have central pits (e.g., Bray, 2009; Senft & Stewart, 2009).

1.8.1.3 Melt-drainage model

The melt-drainage model originally suggested that temperatures reached under the centre of the transient cavity create a melt “plug” which, given the correct dimension and orientation of sub-crater fractures would allow for drainage of brecciated rock, ice and melt-water that could produce central pits in crater floors before the fractures “freeze shut” (Croft, 1981). However, drainage of large amounts of brecciated material could lead to pits in all large lunar craters, which is not observed (Bray, 2009). Therefore, more recent work restricts the draining material to actual molten material (Bray, 2009; Senft & Stewart, 2009; Elder et al., 2012). This model has since been used to explain central pits formations on Mars and Ganymede. For this model to be possible, we would need to assume that the fractures formed during the impact event were large enough to allow the melt to escape. It has been proven that the highest density of impact-induced fractures occur at crater centers (e.g., Kenkmann, 2002), likewise, the largest pockets of impact melt are concentrated in the central regions (e.g., Grieve et al., 1977) leading to a maximum potential for melt drainage at the center of craters (Bray et al., 2009).

Furthermore, if drainage of impact melt is a viable central pit formation mechanism, it must form pits in craters in ice or ice-rock mixtures but not those in volatile-poor targets (Elder et al., 2012).

Computer simulations by Bray (2009) and Senft and Stewart (2011) have both produced modeled craters with melt pools of widths approximately equal to the diameter of central pits on Ganymede. Neither work notes a central depression produced purely from central uplift collapse. Melt drainage model predictions are consistent with (pit diameter) Dp/ (crater diameter) Dc ratios and the presence of central pits in craters of only a certain diameter range, and are in the direction expected as target volatile content changes (i.e., larger Dc and Dp/Dc for higher ice concentrations [Alzate & Barlow, 2011]). To-date, there is neither direct observations supporting or disproving this theory, as one would need to be able to observe the subsurface.
1.8.1.4 Differential erosion of uplifted materials

The final possible floor pit morphology may be a consequence of differential erosion of weaker materials in the core of an uplift or exposing a pit with blocks that do not reach the elevation of the floor of crater (e.g., some central structures on Earth including Gosses Bluff, Australia [Hodges, 1978; Milton et al., 1972; Whitehead et al., 2010]). It is important to mention that global studies of Martian pit craters ranging in diameter from 5 to 157 km from Robbins and Hynek (2012) reported that 92% of central pit craters are qualitatively fresh, consistent with the interpretation that they are mostly relatively young and not likely the result of central peak erosion.

1.9 Crater impactites

Often, the identification of characteristic geological morphologies is required to determine the authenticity of an impact crater formed through an exogenic process. This includes identifying rocks that have been affected by the passage of shock waves, causing deformation known as crater impactites. Impactites are grouped according to the extent to that they have been moved from their original pre-impact location by the cratering flow-field and the subsequent modification and collapse of the transient cavity to form the final crater (Osinski et al., 2013) (Fig. 1.10). These are subdivided into autochthonous (formed in place), paraautochthonous (moved but appear to be in place) and allochthonous (formed elsewhere and moved to their current location). Allochthonous impactites can be further subdivided into those within and around the final crater (proximal) and those some distance from the final crater (distal). The classification of impactites and much of the current understanding of their formational processes comes from the observations of impacts on Earth, which provides ground-truth data on the lithological and structural character of impact structures (in three dimensions) (Grieve & Therriault, 2004).
1.9.1 Deformed layered volcanics

Detailed central uplift bedrock morphologies on Mars have been studied and catalogued globally by Tornabene et al. (2010; 2012a; 2014) and divided into three textural pre-existing bedrock categories: 1) Megabrecciated Bedrock (MBB); 2) Fractured and massive Bedrock (FMB); and 3) Layered Bedrock (LB). Spectral and morphologic analysis of MBB in Toro Crater (Marzo et al., 2010) and several FMB craters including Alga, Ostrov and Stokes (Skok et al., 2012), and Ritchey (Ding et al., 2013; Sun & Milliken, 2014) focused on addressing the regional crustal stratigraphy and geologic history of Mars. However, only limited structural analysis can be accomplished with these bedrock types (e.g., Ding et al., 2014). In contrast, LB central uplifts are particularly useful for structural mapping as the layers provide a frame of reference from which deformation can be recognized (e.g., Caudill et al., 2012; Wulf et al., 2012).

Layered Bedrock (LB) has been defined by Tornabene et al. (2010, 2012a, 2014), as LB uplifts as “a central uplift comprised of large megablocks (100s to 1000s of metres in diameter) of pre-existing relatively flat-lying layered bedrock that was exhumed from depth, tilted, and fractured during the impact process”. Moderately dust-free exposures of LB consist of alternating relatively dark-toned and light-toned, fractured, folded, and
faulted bedrock (Tornabene et al., 2010, 2012a; Caudill et al., 2012) (Fig. 1.11). The layers, once thought to be sedimentary in origin are more consistent with a volcanic origin, based on their combined morphologic and spectral attributes and the geologic setting and geographical distribution of LB central uplifts (Tornabene et al., 2010; Caudill et al., 2012; Quantin et al., 2012; Wulf et al., 2012). The concentration of LB central uplifts observed within the southwestern Tharsis Region was reported initially reported by Tornabene et al. (2010; 2012a) and it was subsequently estimated that ~90% of all known LB craters occur within the Hr unit defined by Scott and Tanaka (1986) (see section 1.3). Caudill et al. (2012) further noted that the Tharsis Region is known for a prolonged history of effusive volcanic activity that produced voluminous layered, cyclic volcanic deposits and that they were likely flat lying and relatively undisturbed prior to impact excavation an uplift by more than 20 complex craters in this region (recently revised to 34 by Tornabene et al., 2014). Spectral analyses via CRISM data of these LB central uplifts by Quantin et al. (2012) indicate that they are dominated by olivine and high-calcium pyroxene typical of basalts and are relatively poor in aqueous alteration phases, which is consistent with a volcanic interpretation.
Figure 1.11: North-eastern outcrop of uplifted LB megablocks in Betio crater. Aeolian beds, clast-free-poor impact melt (as seen from a 25 cm/pixel resolution) can be seen covering the floor of the crater, viewed viewed in ArcScene using (DTEEC_01805_1565_017583_1565_A01 with ESP_016805_1565_RED_A_01_ORTHO draped over with a VE of 1.5x).

1.9.2 Impact Melt Deposits

The bulk of our knowledge on impact melt comes from direct observations of terrestrial field base analysis of impact craters (e.g., Dence, 1968; Grieve et al., 1977), as mentioned above. During the 1960’s and 1970’s, at several Canadian impact structures (e.g., Mistastin and the twin Clearwater Lake structures), field and investigative studies were carried out and provided essential observational information regarding the character and distribution of impact-melted material in impact structures (e.g., Dence, 1971). Impact
melt rocks have a matrix of what was melted rock that formed by the passage of shock waves and rarefaction waves through the target body, particularly at shock pressures of 50 GPa – 100 GPa (Grieve et al., 1977). They generally contain both shocked and unshocked mineral and lithic clasts of target material and can be subdivided on the basis of clast content into clast rich, clast poor and clast free (Osinski et al., 2013). The melt matrix itself may be glassy or crystalline, depending on the cooling history of the melt rock. Within the same impact structure, impact melt rocks tend to have a common composition, as a chemical mixture of those target rocks melted (e.g., Grieve et al., 1977). They can, however, have a wide range of textures at the microscopic scale (e.g., Floran et al., 1978), as a result of a cooling history determined by geological location within the impact structure. These melt deposits form a considerable part of an impact crater and are some of the most characteristic features of hypervelocity impact events (Osinski et al., 2013).

Unfortunately, the entire original extent of impact melt-bearing outcrops is relatively rare on Earth, given their largely surficial nature and the erosional state of terrestrial impact craters (Grieve et al., 1977). Luckily while there is lack of understanding on their complete extent, supplemental observations of fresh craters on the Moon (e.g., Hawke and Head, 1977), and recently on Mars (e.g., Tornabene et al., 2010, 2012a,b; Marzo et al., 2010; Mouginis-Mark & Boyce, 2012) and continued research on Martian meteorites (e.g., impact melt clasts within meteorite NWA 7533 [Humayun et al., 2013]) can provide missing links. These Martian deposits share many similar morphological characteristics with lunar impact melt deposits; in particular, ponds, leveed channels and flow features. One notable difference between lunar and some Martian impact melt deposits is the widespread presence of characteristic pits covering the surface of the latter (Tornabene et al., 2007; 2012b). These are described in more detail below.

The amount of impact melt (Fig. 1.12) produced during an impact event depends largely on velocity, impact angle, and target properties (e.g., composition, rock porosities, etc.) (e.g., Cintala & Grieve, 1998; Osinski et al., 2011). As the magnitude of the crater event increases, the relative volume of melt generated also grows and more will be retained inside the rim of the crater or basin (Cintala & Grieve, 1998), due to the limiting effects
of gravity on cratering efficiency. There are currently no usable estimates for the amount of melt produced in Martian impact craters.

Figure 1.12: Calculated impact-melt volume as a function of modeled transient cavity diameter for the terrestrial case of chondritic projectiles and a granitic target (from Grieve & Cintala, 1992). Included are points representing terrestrial craters formed in crystalline rock. Error bars represent estimated uncertainties in melt volumes and cavity dimensions. Note the good agreement between the modeled and actual cases. This figure also shows the placement of our ~30 km Martian craters (however this does not reflect impact melt measurements for Mars).

1.9.2.1 Crater Related Pitted Material

Pitted materials (Fig. 1.13) were initially recognized in craters by Tornabene et al. (2007) and characterized globally (over 200 impact craters ~1 to 150 km in diameter) by Tornabene et al. (2012b). These pits are quasi-circular to circular or polygonal to irregular-shaped holes that are relatively shallow and generally lack raised rims and any sign of boulders or ejecta materials deposited around them. They can be found within
“ponds” within low-lying topographic areas, flow-like textures and continuous “sheets” on the crater floor similar to impact-melt rich units observed on the Moon and are, therefore, interpreted to be a volatile-rich equivalent of these primary crater deposits. The state and exposure of these pits ranges and some can appear to be quite degraded (see chapter 2 and 3). The current hypothesis is that these pits forms due to the interaction in hot impact melt-bearing impactites with volatiles (Tornabene et al., 2012b). It is still unknown whether these volatiles are from the impact melt deposits themselves – in which case these deposits could be similar to volatile-rich impact melt-bearing breccia deposits (i.e., suevite), which often contain so-called degassing pipes.

Figure 1.13: Crater-related pitted materials observed on the floor (i.e., the crater-fill) of Zumba Crater in Daedalia Planum from Tornabene et al (2012b) photo. This 3D perspective view was generated by draping the combined full resolution (25.2 cm/pixel) orthorectified HiRISE RED and IRB color composite image on a 1 m per post HiRISE stereo pair-derived DTM (DTEEC_002118_1510_003608_1510_A01).
1.10 References


Greeley, R., and Guest, J., 1987, USGS Map I-1802-B.


Chapter 2

2 Morphologic and Structural Mapping of the Central Uplift of Betio Crater, Thaumasia Planum, Mars

2.1 Introduction

Meteorite impact craters are among the most recognizable geological landforms on almost all terrestrial planets and rocky and icy satellites in our Solar System. They form during an early instantaneous event that alters the planetary surface and redistributes subsurface materials (see Melosh, 1989). It is well known that impact craters display two main characteristic morphologies: simple and complex. Whereas simple craters comprise a relatively simple bowl-shaped form, complex craters possess an uplifted rim, terraced walls, down-faulted annular trough containing impact melt rocks and/or breccias, and an uplifted structure on the crater floor referred to as a central uplift (e.g., Dence, 1968; Pike, 1980; Melosh, 1989). The central uplifts of complex craters are of considerable scientific interest. Central uplifts exhume deep-seated bedrock from the subsurface that is not otherwise exposed at the surface. They also preserve morphologic and structural features from both the pre-impact target and those formed during the impact process. In addition, they may provide habitats for life through the exhumation of altered materials or via impact-generated hydrothermal systems (e.g., Cockell et al., 2013; Osinski et al. 2013; Tornabene et al., 2013). On Earth, central uplifts may also contain potential economic mineral and hydrocarbon deposits (Grieve, 2013). On Mars, central uplifts are known to be morphologically quite diverse with these features not only being comprised of a peak(s), but also a pit morphologies including summit pits and central floor pits (e.g., Barlow, 2010). Despite their importance, much remains unknown about the detailed mechanics and process(es) of central uplift formation.

Of the ~184 terrestrial impact structures presently known, only ~40 contain central uplifts that are exposed and accessible, and only a fraction of those have been comprehensively studied and mapped in detail (e.g., Kenkmann et al., 2005; Osinski & Spray, 2005). As impact structures on Earth are extensively eroded to varying degrees, the well-preserved morphology of central uplifts are best studied on other planetary bodies. More detailed
studies and mapping efforts of crater central uplifts on other bodies are enabled by the increased availability of higher resolution orbital data, including visible images, digital elevation, thermal, and spectral from past and ongoing missions.

Detailed central uplift bedrock morphologies on Mars have been studied and catalogued globally by Tornabene et al. (2010; 2013). In addition, the recognition of different types of impactites (e.g., breccias and melts) and impact-related structures (e.g., fractures, faults, folds, and dykes), was also looked at. Based on the results of their global survey, uplifted bedrock was categorized into three textural categories: 1) Megabrecciated (MBB); 2) fractured massive-textured Bedrock (FMB); and 3) Layered (LB). Spectral and morphologic analysis of MBB in Toro Crater (Marzo et al., 2010) and several FMB craters including Alga, Ostrov and Stokes (Skok et al., 2012), and Ritchey (Ding et al., 2013; Sun & Milliken, 2014) focused on addressing the regional crustal stratigraphy and geologic history of Mars. However, only limited structural analyses can be accomplished with these bedrock types (e.g., Ding et al., 2013). In contrast, LB central uplifts are particularly useful for structural mapping, as the layers provide a frame of reference from which deformation can be recognized (e.g., Caudill et al., 2012; Wulf et al., 2012).

We carried out geologic mapping within a Geographic Information System (GIS) framework focusing on the morphological and structural analysis of the well-preserved complex Martian crater, Betio crater, in Thaumasia Planum. Herein, we provide a synthesis of our observations and interpretations of the relatively well-preserved central uplift units, and the various structures contained therein, to provide further insights into the impact cratering process. The implications of these results for understanding weakening mechanisms during the modification stage leading to central uplift formation, as well as mechanics resulting in a floor pit morphology, are discussed, along with analysis and interpretation of the associated morphologic units and an emphasis on formation of impact-generated dykes.

2.2 Background: Betio impact crater

This study focuses on Betio crater (281.38° E, -23.15° N), which is a mid-sized, nearly circular (~31.7 km diameter), well-preserved, complex crater located ~830 km south of
Valles Marineris, Mars. It occurs within the Hesperian-aged Ridged Plains material unit (Hr) defined by Scott and Tanaka (1986) on the boarder between Thaumasia Planum and Solis Planum (Fig. 2.1A). Based on a survey of central uplift bedrock morphologies, the greater Thaumasia and Solis Hr plains unit is interpreted by Caudill et al. (2012) to represent the thickest layered volcanic sequence on Mars ($\sim 8-17 \times 10^5 \text{km}^3$) comprised of alternating and cyclic layering of relatively thicker light- and relatively thinner dark-toned layers. This unit is characterized by a sporadically cratered and relatively flat plain marked by several widely-spaced sinuous wrinkle-ridges and is interpreted as a volcanic unit. Wrinkle ridges are commonly accepted to form when contractional deformation (Plescia & Golombek, 1986; Watters, 1988), which differentially affects the underlying and surface stratigraphy having differing layer strengths (Golombek et al., 2001). We have observed two wrinkle ridges near Betio impact crater, with one appearing to terminate in the SW quadrant at the rim (Fig. 2.1B) and one that has affected the western rim of the structure (Fig. 2.1C).

Betio crater is surrounded by a relatively symmetrical multiple layered ejecta (MLE) blanket that reaches as far as $\sim 54 \text{ km}$ from the rim. The crater rim sits at an average altitude of $\sim 2887 \text{ m}$, with a max altitude of $\sim 3086 \text{ m}$ (500 m above the surrounding terrain). The maximum rim height is found on the western portion of the rim, where there is an irregularity most likely caused the Betio-forming event intersecting a pre-existing wrinkle ridge in the target terrain. This wrinkle ridge appears to have expanded the diameter of Betio crater, increasing it by $\sim 3 \text{ km}$ (based on a best fit circle of $\sim 29 \text{ km}$). The relatively smooth crater floor sits at an average altitude of $\sim 1594 \text{ m}$ ($\text{Min-Max} = \sim 1648 \text{ m} - 1540 \text{ m}$) (Fig. 2.1D) giving it a depth of $\sim 1293 \text{ m}$ below the rim. Immediately after formation (based on crater scaling, the average depth should have been $\sim 1822 \text{ m}$ using the depth diameter equation ($d=0.302D^{0.52}$) for Solis Planum (Stewart & Valiant, 2006). The difference between the estimated depth and the actual depth is $\sim 529 \text{ m}$ making Betio crater quite shallower in comparison to most craters. However, if we take into account the wrinkle ridge on the western rim possibly affecting the apparent diameter of the crater and use the inner rim, putting Betio crater at a diameter of 28.7 the depth would be estimated at $\sim 1730 \text{ m}$ – closer to the depth of Betio crater we measured (difference of $\sim 437 \text{ m}$). Approximately 6.6 km from the terraces, there is an apparent
drop in elevation, which is interpreted to mark the start of the central pit. Central pit morphologies are defined as a depression either into the crater floor, or atop a central topographic rise (e.g., Wood et al., 1978; Barlow, 2010) that are referred to as “floor pits” and “summit or peak pits”, respectively. They are common on Mars, Callisto, and Ganymede and were first identified on Mars using Mariner 9 (Smith, 1976) and Viking (Hale & Head, 1981; Hale, 1982, 1983; Hodges et al., 1980) images. The central pit of Betio is asymmetrical ~10.8 km NW–SE and ~8.8 km NE–SW in diameter and covers an area of ~67 km². Approximately 500 m to the east of the floor pit there is a ~4.5 km diameter simple crater lacking a discernable ejecta blanket.
Figure 2.1: (A) Context view of Betio crater as seen in MOLA colourized elevation. (B) Betio crater as seen with THEMIS DAY VIS, noting the apparent multiple layer ejecta (MLE) and nearby wrinkle ridges. A pre-existing wrinkle ridge may have some effect on the western crater rim, making it higher in elevation and possibly widening the crater by ~3 km. (C) Betio impact crater seen in CTX image (B03_010647_1546_XN_25S078W) with location of associated cross section. (D) Betio crater cross-section showing the rim and floor pit elevation. The rim elevation and crater diameter may be have been affected by a pre-existing wrinkle ridge on the western side, as noted above.
2.3 Data and methods

For this study, we utilized images from the High Resolution Imaging Science Experiment (HiRISE) onboard the Mars Reconnaissance Orbiter (MRO). These images have a resolution up to ~0.25 m/pixel (HiRISE; [McEwen et al., 2007; 2010]) and are capable of resolving meter-scale features on central uplifts that allow detailed morphological and structural mapping. Mapping was conducted primarily on a combination of orthorectified (where available) and non-orthorectified HiRISE images (Table 2.1). The HiRISE images were augmented by various other data products (Table 2.1) including, elevation data (Digital Terrain Models [DTMs] derived from the Mars Orbiter Laser Altimeter [MOLA~462 m/pixel]; [Smith et al., 2001], High Resolution Stereo Colour Imager [HRSC~150-175 m/pixel] [Neukum & Jaumann, 2004]), visible (~18 m/pixel) and thermal infrared (~100 m/ pixel) images from the Thermal Emission Imaging System (THEMIS) (Christensen et al., 2004), and context camera (CTX) images (~6 m/pixel) (Malin et al., 2007). In addition, we utilized THEMIS-derived thermal inertia measurements (Fergason et al., 2006; Putzig & Mellon, 2007), which represents the resistance to change in temperature of the upper few centimeters of the surface throughout the day, independent of local time, latitude, and season making a valuable aid to understanding the near-surficial geology and recent processes that are potentially still active today. This data product helped to differentiate variable grain sizes, thus, distinguishing the difference between bedrock and fine-grained unconsolidated units, such as aeolian deposits.

The construction of a morphologic and structural map for Bečtio was accomplished by visual interpretation of the available datasets, with emphasis on HiRISE grayscale and colour images supported by CTX. All data were initially qualitatively inspected in the Java Mission-planning and Analysis for Remote Sensing GIS software package (JMARS; Christensen et al., 2009) and then downloaded from further analysis. The data were then imported and, in some cases, georeferenced (as needed) in ESRI ArcMap 10 to the MOLA base map, so they could be examined in a geographically accurate context. The remaining data were overlain in order of lowest to highest resolution, with HiRISE images representing the top-most layer. The first step in constructing the geologic map
involved the identification and characterization of distinct morphologic units and structural features within the study area. The image stretch was altered to ensure proper visibility in illuminated and shaded areas using the Dynamic Range Analysis (DRA) function within ArcMap. Multiple point, line, and polygon shapefiles of individual mappable units were created and digitized for the various geomorphologic and structural features observed in Betio. An Equirectangular_MARS projection was assigned to all images and shapefiles.

Table 2.1: Betio crater High-resolution Orbitsl Images and Topography Data IDs

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Orientation (i.e., strike) of the faults, fractures, and dykes were measured using the coordinate geometry (COGO) report tool, in the editor extension in ArcMap using the right hand rule. Strike and dip values for the layers within uplifted megablocks were geometrically calculated using the Layertools extension for ArcGIS (Kneissl et al., 2010) on a 2 m/pixel HiRISE DTM (Mattson et al., 2011), covering the eastern side of the central pit. Each block was measured ~20-40 times, depending on the size (i.e., more measurements for larger blocks) and a mean vector average was then calculated from multiple measurements for each sector. These values measured for each linear feature were plotted on an equal-area stereonets, as either planes or rose diagrams in OSX Stereonet v. 9.0.1. (Allmendinger et al., 2013). Based on a previous study by Wulf et al. (2012), the strike values determined with this tool are estimated to have an error of <10° for beds inclined at an angle of 75° or steeper, while the error for flatter beds is expected to gradually increase. Strike of sub-horizontal bedding (<15–25°) has a high degree of uncertainty. As such, it was important to visually confirm results with stereo-derived anaglyphs or a 3D model of Betio in ArcScene.

A detailed morphological map of Betio crater’s central pit is provided in Figure 2.4A. Units were defined based on surface appearance (tonality, surface textures, structures,
features (HiRISE image [~25 cm/pixel]), thermophysical properties of the surface, average THEMIS (~18-100 m/pixel) thermal inertia (Fig. 2.2C and Table 2.2) value (size of the particles), elevation maps outlining the units height (Fig. 2.2A) and their average slope (HiRISE DTM (~2 m/pixel) (Fig. 2.2B and Table 2.2) (steepness of a unit in degrees). Our mapped morphological units are described below.

2.3.1 Morphologic units

2.3.1.1 Layered dark and light-toned outcrops (*light-green*)

This unit is characterized by outcrops of alternating layers of generally thinner and higher-standing dark-toned layers and thicker and lower-standing light-toned layers (Fig. 2.3A), which are consistent with the observations of Caudill et al. (2012). When visibility permits it, particularly in the north-eastern quadrant on the eastern side of the uplift, the layers bound exposures of light-toned high thermal inertia (~422 J m\(^{-2}\) K\(^{-1}\) s\(^{-\frac{1}{2}}\)) bedrock that appears to have been fractured, folded, faulted, and uplifted into large megablocks (100s to 1,000s of km across). There is also some exposure on relatively flat portions of the pit floor towards the center of the central pit. The central uplift exhibits ~10 km\(^2\) (~15% of the total area) exposed layered bedrock. The best exposure is in the north-eastern sector of the central floor pit, where the majority of the exposed layered bedrock is on the west- and south-facing sides of discrete uplifted blocks.

2.3.1.2 Dark-toned smooth unit (*orange*)

This is a relatively smooth textured, dark-toned unit, lacking any visible fragments of light-toned rock (at the ~25 cm/pixel resolution), appears to preserve multiple overprinting craters, which suggests that it is coherent. We observe this unit generally in areas of low slopes (~1-13°), embaying and coating portions of the central uplift. Sometimes, it appears as though it has pooled or flowed from topographic highs to lows of the uplifted megablocks (eastern side [Fig. 2.3B]) and the crater floor. It is present throughout much of the central uplift covering ~ 34.5 km\(^2\) (comprising 51.5% of the total area), making it the most prominent unit in the central floor pit. It is important to mention that this estimate is conservative, as there are clearly some aeolian deposits on top of this unit (see section 2.4.1.7). This unit has a mid-range thermal inertia of ~230 J m\(^{-2}\) K\(^{-1}\) s\(^{-\frac{1}{2}}\).
2.3.1.3 Dark-toned smooth unit exposing layers (*dark-green*)

This relatively dark-toned unit has a similar appearance to the dark-toned smooth unit but has very clear linear layering texture that is similar to the dark-toned higher-standing layers in the megablocks (Fig. 2.3C). We observe this unit on the uplifted megablocks near the scarp wall consistently throughout the floor pit covering 7 km$^2$ (comprising 10.5% of the total area). There is minimal aeolian coverage on this unit and it has a mid-range average thermal inertia of $\sim$276 J m$^{-2}$ K$^{-1}$ s$^{-\frac{1}{2}}$.

2.3.1.4 Light-toned clast-rich unit (*red*)

This unit contains light-toned, angular and subrounded rock fragments (or clasts) embedded in a dark toned matrix (Fig. 2.3D). These clasts range in size (meter to decameter) and density. In the western outcrop, the clasts are densely packed and range from $\sim$6 - 45 m in diameter and some of the larger clasts can be observed to contain layers similar to the layering seen in the uplifted megablocks. Comparable sized clasts are observed in the northeast outcrops ranging from $\sim$4 – 26 m. Generally the clasts become smaller and less dense, when the elevation increases, and range from <1-5 m. Typically this unit is observed to be relatively clast-poor, when observed at higher elevations on the uplift, specifically near the outer rim of the pit; whereas, it becomes more clast-rich, where it borders or embays portions of the layered bedrock unit or uplifted megablocks. This unit occupies an area of $\sim$3 km$^2$ (4.5% of the total area) and the top edge of the scarp and between blocks in the northern region, as well as in a topographic low that lies between the scarp and blocks (Fig. 2.4a [red unit]), continuing around the sides of the megablocks bounding them. The average thermal inertia for this unit is $\sim$278 J m$^{-2}$ K$^{-1}$ s$^{-\frac{1}{2}}$.

2.3.1.5 Quasi-circular pit-bearing dark-toned unit (*aqua*)

This unit is dark-toned containing quasi-circular (honey-comb-like) clusters of subdued pits (Fig. 2.3E). This unit varies in preservation, with some containing aeolian deposits within the pits, and what appears to be light-toned boulders around the “rim” of the pits. The measureable pit diameters range from $\sim$51 m – 145 m, with the average diameter of the ten largest being $\sim$106 m. We observe this unit between the floor pit boundary and the
scarp with the best-preserved clusters in the S and NE covering ~3 km$^2$ and making up 4.5% of the total floor pit area. This unit located in relatively flat areas that are straigraphically higher than the dark-toned smooth unit. It has an average slope of ~4.5° and an average thermal inertia of ~285 J m$^{-2}$ K$^{-1}$ s$^{1/2}$.

2.3.1.6 Unconsolidated materials (brown)

This unit includes unconsolidated deposits with various sized light- and dark-toned boulders up to a few meters in diameter (Fig. 2.3F). We observe this unit on steep slopes (up to ~46° with an average of ~31°) in sections of the layered exposed bedrock but sometimes deposited in areas of lower slope. The unit covers ~1 km$^2$ (~1.5% of the total area). These materials have the second highest average thermal inertia of ~380 J m$^{-2}$ K$^{-1}$ s$^{1/2}$ relative to the other units.

2.3.1.7 Aeolian deposits (beige)

This unit is composed of linear deposits (typically trending NE-SW) of large accumulations of unconsolidated materials, exhibiting characteristic aeolian bedforms (i.e., dunes) (Fig. 2.3G). We observe this unit typically in topographic depressions throughout the central structure, covering the floor of the central pit depression, the depressions between the uplifted blocks and scarp on the western side, and flat depressions on the uplifted blocks typically in areas of low slope ~4.3°. This unit covers ~7 km$^2$ (~10.4% of the total area) and has the lowest thermal inertia average of ~205 J m$^{-2}$ K$^{-1}$ s$^{1/2}$. We should note that this estimate is likely higher, because the aeolian deposits overlie the majority of units with higher thermal inertias. We include, herein, a facies of this unit that contains aeolian deposits, with some exposure of layered bedrock that underlies it (Fig. 2.3H). These particular areas, found on pit floor, are slightly higher with thermal inertias of ~255 J m$^{-2}$ K$^{-1}$ s$^{1/2}$. 
Figure 2.2: Derived maps for Betio crater central floor pit. (A) HiRISE colourized DEM (DTEEC_016805_1565_017583_1565) for eastern side overlain on a MOLA Colorized digital elevation model where the stretch has been changed to max – min of 1495.87-1106.5 to better expose the difference in elevation of the blocks. Addition of 20 m interval contours. (B) Slope map generated using DTEEC_016805_1565_017583_1565 (C) THEMIS derived thermal inertia map using (image ID’s (Night: I01420006, Day: 01011700, 126556014, 01011706).
Figure 2.3: Betio crater morphological unit observations as observed in HiRISE (ESP_017583_1565 and ESP_027987_1565) (for context please refer to figure 4) (A) Layered dark and light-toned outcrops - alternating thinner, dark-toned and thicker light-toned layers typical of a layered megablock in Betio and other craters in the region. (B) Dark-toned smooth unit flowing down the side of a block on the
eastern side. (C) Dark-toned smooth unit exposing layers near the NW scarp. (D) Light-toned clast-rich unit in the central western portion of Betio, containing rounded to angular clasts. (E) The best-preserved outcrop of quasi-circular pit bearing dark-toned unit seen in the south near the scarp. (F) Unconsolidated materials on a steep slope megablock of alternating dark- and light-toned bedrock in Betio craters’ eastern quadrant. (G) Aeolian deposits in the bottom on Betio crater floor pit. (F) Exposed layers with aeolian deposits overtop seen in the central pit. (Please note the difference in scale in each image). Refer to Appendix A.1 for reference within Betio crater floor pit.
Table 3: Overview of the morphologic units in Betio crater

<table>
<thead>
<tr>
<th>Unit Name</th>
<th>Interpretation</th>
<th>HiRISE DTM Derived Slope</th>
<th>THEMIS Derived Thermal Inertia</th>
<th>Area (km²) coverage</th>
<th>Percentange(%) of total floor-pit area coverage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Layered dark- and light-toned outcrops</td>
<td>Layered Bedrock</td>
<td>25</td>
<td>3</td>
<td>13</td>
<td>7.5</td>
</tr>
<tr>
<td>Dark-toned smooth unit</td>
<td>Clast-poor impact melt</td>
<td>15</td>
<td>2</td>
<td>6.5</td>
<td>4</td>
</tr>
<tr>
<td>Dark-toned smooth unit exposing layers</td>
<td>Clast-poor impact melt</td>
<td>15</td>
<td>2</td>
<td>6.8</td>
<td>4.3</td>
</tr>
<tr>
<td>Light-toned clast-rich unit</td>
<td>Clast-rich impact melts or breccias</td>
<td>21</td>
<td>3</td>
<td>11.2</td>
<td>5.8</td>
</tr>
<tr>
<td>Quasi-circular pit bearing dark-toned unit</td>
<td>Pitted Material</td>
<td>6</td>
<td>0.3</td>
<td>2.8</td>
<td>1.8</td>
</tr>
<tr>
<td>Unconsolidated materials</td>
<td>Regolith</td>
<td>46</td>
<td>18</td>
<td>31</td>
<td>8</td>
</tr>
<tr>
<td>Aeolian deposits</td>
<td>Aeolian Deposits</td>
<td>8.8</td>
<td>1</td>
<td>4.3</td>
<td>2.8</td>
</tr>
<tr>
<td>Aeolian deposits with exposed layers</td>
<td>Layers with Aeolian Coverage</td>
<td>7</td>
<td>2</td>
<td>4.4</td>
<td>2.1</td>
</tr>
</tbody>
</table>
Figure 2.4: Geomorphologic map of Betio crater central floor pit. This map outlines Betio crater’s remotely observed morphologic units and their interpretation (discussed in section 2.5.1.). A higher resolution version of this map is included in Appendix B.
2.3.2 Structural assessment

2.3.2.1 General floor pit structure

A detailed structural map of Betio crater central floor pit is provided in Figure 2.14. There are three main structural units that make up the central floor pit – the floor pit rim, megablock scarp and the hourglass-shaped central depression (Fig. 2.5). We defined the extent of these features by utilizing high-resolution DTM (HiRISE and MOLA), in conjunction with high-resolution images (CTX and HiRISE). The apparent floor pit rim is more easily defined using topography data than in the visible images. We define this boundary, when the crater floor begins to drop in elevation, starting around ~7 km from the terraces and at an elevation of ~1450 m (Fig. 2.5). This boundary is quasi-circular and measures ~10.8 km NW-SE and ~8.8 km NE-SW in diameter and has an area of ~67 km². Within the floor pit, there is a very distinct scarp feature (Fig. 2.5), with pronounced edges, typically coated in the dark-toned smooth unit. The scarp also has a quasi-circular perimeter, giving it a sinuous appearance with an ~6.7 km diameter, sitting at an average elevation of ~1441 m. The scarp has a steep slope on the western and northern sides (~20-44°) but is much shallower in the south (~5-12°). The uplifted megablocks of layered bedrock are bounded by the scarp and notably do not occur within the elevations above it. Within the centre of the floor pit, we observe a relatively flat-floored, hourglass-shaped depression (Fig. 2.5) almost completely covered with the aeolian deposits unit and the dark toned smooth unit. This hourglass-shaped pit trends NW–SE, 4.5 km in length and has an ~1.3 km diameter for its NW-portion, which narrows to ~750 m near the centre, then expands back out to ~1.2 km in diameter in the SE bottom portion. The lowest depth of the pit is ~1110 m, which lies ~340 m below the pit rim. There is an apparent drop in elevation from the NW oval to the SE oval observed in topographic profiles (Fig. 2.4B).
2.3.2.2 Uplifted blocks

We define megablocks as exposed, coherent blocks of layered bedrock, ranging from ~80 to ~2.2 km across (orientation of the bedding), that have been uplifted, rotated, and deformed during the modification stage of crater formation. The average blocks size (diameter) measured ranged from 100-300 m (Fig. 2.7), similar to results found at the central uplift of the 80-km diameter Puchezh-Katunki impact crater in Russia (utilizing borehole samples) with an average block size of ~100 m (Ivanov et al., 1996). The blocks are relatively symmetrically arranged around an hourglass-shaped relatively flat
central depression (Fig. 2.5) and extend to the edge of the scarp but are never higher in elevation than the scarp. The uplifted blocks generally display dips anywhere from ~5 to 85°, similar to measurements made at the Jebel Waqf as Suwwan in Jordan (Kenkmann et al., 2010). It should be noted that dip measurements were only conducted for the eastern half of the floor pit, due to lack of coverage from high-resolution stereo images required to derive a digital terrain model for the western side.

It is apparent that the blocks increase in dip (~45–85°), as they become more proximal to the centre of the crater (Fig. 2.8A,B). The best exposures of bedrock are observed in the eastern side (Fig. 2.4, 2.8C,D). The larger coherent blocks, (>800 m in diameter and length), near the scarp edge (Fig. 2.10D), tend to have shallower dips (~5–15°). There is a noticeable trend in the blocks that appear to deform with the scarp, from ~ESE-WNW or E-W (i.e., concentric orientation in the north) to ~NE-SW or NNE-SSW (i.e., radial orientation in the west and east) (Fig. 2.6); however, more frequent intra-block strike changes occur towards the centre of the structure (Fig. 2.6C). There are depressions or a quasi-circumferential moat between the uplifted megablocks in the western portion of the central pit and the scarp; whereas, the blocks connect to the scarp everywhere else, when visibility permits it. We also observe that, in the southwest quadrant, the blocks tend to be less exposed, smaller, and quite variable in strike measurements, as well as at a lower elevation (Fig. 2.6, 2.14). In the south, almost no visible layering is found, due to coverage by the smooth dark toned unit.
Figure 2.6: Rose diagrams showing the orientation of blocks within Betio crater. Concentric circles represent the distance in strike measurement frequency. Strikes are plotted on equal area projections. (A) Northern section with an ~WNW – ESE trend. (B) Eastern section with a ~N-S trend. (C) Southwestern section ~NW-SE trend. (D) Western section ~NW-SE trend. There is a somewhat concentric pattern to the block orientations, except for the southwestern sector due to its highly disturbed orientations of the blocks (fig. 2.14).

Figure 2.7: Histogram of block sizes within Betio crater. The average diameter of the blocks ranges from 100 to 300 m; although there is considerable variation.
Figure 2.8: Strike and dip measurements (black dots) plotted on an equal area stereonet plotted as poles to planes for four blocks in Betio’s eastern half. The rainbow contour map represents the concentration of data points for strike and dip measurements. It is clear that beds closer to the centre are steeply inclined and, therefore, have higher dips (C and D) than the beds farther away closer to the scarp (A and B). It should be mentioned that it is possible to discern the orientation of the blocks for example stereonet “B” indicates that there may be a fold within the block. The image in the centre is an orthorectified HiRISE red mosaic image draped over the DEM (DTEEC_016805_1565_017583_1565) and viewed using ArcScene and given a Vertical Exaggeration (VE) of 2.5x.
2.3.2.3 Structural features: Faults, Folds, Fractures, and Dykes

Figure 2.9: Betio crater structural deformation features as observed in HiRISE (ESP_017583_1565 and ESP_027987_1565). (A) Large fold in NW section indicated with arrow (B) Two small folds (indicated with arrows) with dykes crosscutting the hinge, as well as faulting occurring within the folds supporting the idea that these
folds are actually radial transpression ridges. (C) A fairly undisturbed uplifted block with most layers intact, minor displacements are seen between the layers of a few metres (D). Series of small offsets in the layers near the scarp rim in the NE indicated with arrow. (E) A series of large offset dykes in the north indicated with arrows. (F) Close-in image of impact dykes with clasts noted by the arrow.

2.3.2.3.1 Faults and folds

The exposed bedrock within the floor pit contains numerous observed, approximated, and interpreted faults (Fig. 2.13). Towards the center of the crater, the faults are more densely distributed and radially oriented, displacing layers a few- to tens- of meters, offsetting dykes, displacing fold limbs (Figs. 2.10 & 2.11) and interpreted to rotate blocks up to 90 degrees. The faults bounding the blocks in this region are sometimes concentric (Fig. 2.10) possibly representing thrust faults.

Within the central pit, we observe multiple large (~500 m – 1300 km) radially striking fold axes (Fig. 2.11), within antiforms and synforms located in the periphery of the central floor pit that are relatively symmetrical with open (~70°–120°) interlimb angles (Fig. 2.9A). There are also multiple smaller (~45 m – 499 m) (Fig. 2.9B) folds, located towards the center of the floor pit. Some of these folds appear to be asymmetrical, with tight (0°-30°) to isoclinal (0°-10°) interlimb angles, with some chevron shapes. There is a general gradational transition in size, fold tightness, and symmetry from the outer regions to towards the blocks proximal to the flat centre, due to space issues. We also observe fold limbs that occasionally appear to be detached and offset into sheet-like blocks, bounded by faults.
Figure 2.10: Rose diagrams showing the orientation of faults within Betio crater. Concentric circles represent the distance in strike measurement frequency. Strikes are plotted on equal area projections. (A) The northern section has a strong ~NE-SW trend. (B) The eastern section has a ~NNE-SSW trend. (C) The southwestern section has three trends ~NE-SW, ~E-W, and ~NW-SE trend (D) Western section has a ~W-E trend.
2.3.2.3 Fractures and Dykes

We observe 81 tabular structures crosscutting the layered megablocks within blocks (Fig. 2.9E), fold hinges (Fig. 2.9B) and possibly at the boundary between two blocks. These structures are oriented almost perpendicular (~70–110° [average ~67%]) to the strike of the layers, as can be seen in (Fig. 2.12). The dykes range from ~2 to 51 m (average ~9 m, standard deviation of 7 m) in width and ~22 m up to 1.2 km (~135 m, standard deviation 164 m) in length (discontinuous); however, they can also vary in width with increasing distance to the crater center. Sometimes, the dykes disappear due to extensive coverage from other overlying units. As such, it can be challenging in some cases to determine their full extent. Fifty-five dykes show clear offset with two or more segments (~68% of
the total dykes). The offset ranges from ~40 m to 2 m with an average of 6 m (Figs. 2.9E,F). It is important to note that, although there are dykes with only one intact segment visible the dyke, may continue and be offset but may no longer be exposed due to coverage by other units. Some of the better-exposed structures appear to contain angular light-toned clasts in a smooth low albedo matrix situated in the northern quadrant (Fig. 2.9F). A higher concentration of these structures can be found in the north and eastern sections of the central uplift (Figs. 2.12, 2.14), with four times more dykes in the northern section (Figs. 2.9E, 2.12) than the west and southwest sections.

Figure 2.12: Rose diagrams showing the orientation of the dykes within Betio Crater. Concentric circles represent the distance in strike measurement frequency. Strikes are plotted on equal area projections. (A) Northern section with an ~NNE – SSW trend. (B) Eastern section with a ~NNE-SSW and ~E-W trends. (C) Southwestern section with a ~NE–SW and ~N-S trends. (D) Western section with an ~NE –SW trend.
Figure 2.13: Betio crater structural map, outlining the layers (purple), dykes (orange), faults (approximate [approx.], inferred [infer.], observed), and folds (antiforms and synforms).
Figure 2.14: Close-up view of southwestern structural map showing the intensely brecciated, and collapsed area with multiple, faults, folds, fractures and dykes.

2.4 Discussion

2.4.1 Interpretations of the morphologic units

In section 2.4, observations of each surface morphologic unit were made and a map was constructed for Betio craters floor pit (Fig. 2.4). Based on these observations in conjunction with observations made in terrestrial, lunar, and other Martian craters, we have provided interpretations as to what these units might represent.
2.4.1.1 Layered dark and light-toned outcrops –layered bedrock

The observations we previously noted on layered dark and light-toned outcrops are consistent with the early observations of Tornabene et al. (2010, 2012a) and Caudill et al. (2012). Caudill et al. (2012) defined LB as “a central uplift comprised of large megablocks (100s to 1000s of metres in diameter) of pre-existing relatively flat-lying layered bedrock that was exhumed from depth, tilted, and fractured during the impact process”. Moderately dust-free exposures of LB consist of alternating relatively dark-toned and light-toned, fractured, folded, and faulted bedrock (Tornabene et al., 2010, 2014; Caudill et al., 2012). The layers are not likely sedimentary in origin but are interpreted to be volcanic, based on their combined morphologic and spectral attributes, and the geologic setting and geographical distribution of LB central uplifts (Tornabene et al., 2010; 2014; Caudill et al., 2012; Wulf et al., 2012; Quantin et al., 2012). The concentration of LB central uplifts observed within the Tharsis Region was reported initially by Tornabene et al. (2010, 2012a) and it was subsequently estimated that ~90% of all known LB craters occur within the Hesperian-aged Ridged Plains Material unit defined by Scott and Tanaka (1986) (see Caudill et al., 2012). Caudill et al. (2012) further noted that the Tharsis Region is known for a prolonged history of effusive volcanic activity that produced voluminous layered, cyclic volcanic deposits and that they were likely flat lying and relatively undisturbed prior to impact excavation and uplift by more than 20 complex craters in this region. The Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) spectral analyses of these LB central uplifts by Quantin et al. (2012) indicate that they are dominated by unaltered mafic components (i.e., olivine and high-calcium pyroxene typical of basalts) and are relatively poor in aqueous alteration phases, which is consistent with a volcanic interpretation. We also note that an observable relationship between the bedrock exposure and the observable direction of the aeolian deposits and inferred predominant wind direction ENE, which suggests erosion and exposure of bedrock, likely from sand-blasting (Fig. 2.3A).

2.4.1.2 Impact melt rocks

On Earth, impact melt is typically occurs as melted rock, containing variable amounts of target rock clasts and can be either clast-rich, -poor or -free (Osinski et al., 2013). The
amount of impact melt produced during an impact event depends largely on velocity, size of impact, and rock porosities of the target and projectile (e.g., Cintala & Grieve, 1998; Wännemann et al., 2008). As the magnitude of the impact increases, the volume of melt generated grows disproportionately with respect to the size of the transient cavity (e.g., Grieve & Cintala, 1992; Cintala & Grieve, 1998). Where these impact melt products are concentrated depends on the final modification phase of crater formation (see Osinski et al., 2011). Melting during the excavation and modification flow might provide the necessary lubrication to lower the strength of block contacts during the later stages of movement (Dence et al., 1977; Spray, 2010). Strength degradation in the target blocks during impact will be discussed further in this paper.

2.4.1.2.1 Dark-toned smooth unit and dark-toned smooth unit exposing layers – clast-free or -poor impact melt

We interpret the dark-toned smooth material, as well as dark-toned smooth cap material coating the layers, to be a clast-free or -poor impact melt (at the 25 cm/pixel resolution of HiRISE) at different stages of erosion for the following reasons. These units display evidence for ponding, flowing, and formation of coherent sheets on low slopes (~2–15°), which mimic observations of melt deposits at other craters on the Moon (e.g., Hawke & Head, 1977), Earth (e.g., Grieve & Cintala, 1992) and Mars (e.g., Marzo et al., 2010; Tornabene et al., 2010; 2014). The THEMIS thermal inertia for both of the units is at the mid range with respect to the other units (~230–276 J m⁻² K⁻¹ s⁻½), suggesting a poorly sorted mixture of particle sizes (e.g., clasts within a fine-grained melt matrix). We suggest that the dark-toned smooth material is better-preserved and, thus, a more resistant unit that grades into the thinner dark-toned smooth unit, exposing layers as a result of erosion. We can assume this because of the slightly higher thermal inertia measurements for the layer-exposing unit, as a result of bedrock becoming partially exposed, and thus, increasing the grain size of that unit.

2.4.1.2.2 Light-toned clast-rich unit –clast-rich impact melt breccia and fault breccias

The light-toned clast-rich unit, which covers ~4.5% of Betio crater floor pit, is clearly a mixture of angular fragments from different types of rocks surrounded by a fine-grained
matrix that may be (comprised of melt or very fine fragmented material). It is not possible to ascertain whether the matrix is melt or clastic but based on the known stratigraphy from craters on Earth (i.e., that impact melt sheets become increasingly clast-rich towards their base (Osinski et al., 2013), a possible explanation is that this unit is clast-rich impact melt (e.g., Marzo et al., 2010; Tornabene et al., 2010; Osinski et al., 2011). The average thermal inertia for this unit is \(~278 \text{ J m}^{-2} \text{ K}^{-1} \text{ s}^{-\frac{1}{2}}\) suggesting a combination of particle sizes and higher than the dark-toned smooth material supporting the addition of fragmented clasts (Fergason et al., 2006). Alternatively, these breccias may represent fault breccias formed between the layered megablocks. Evidence supporting this comes from the larger clasts that contain layers, similar to the layering seen in the uplifted megablocks. These breccias may be evidence of friction between large blocks or slip between fault zones enabling crater collapse during the modification stage.

2.4.1.2.3 Quasi-circular pit bearing dark-toned unit – pitted material

We interpret quasi-circular pit bearings dark-toned unit to be eroded and degraded crater-related pitted material (see Tornabene et al., 2012b). Based on the observations made by Tornabene et al. (2012b) in over 200 impact craters ~1 to 150 km in diameter, the these pitted deposits are thought to have been emplaced as hot impact melt-bearing deposits that may have entrained volatiles from volatile-rich target materials. These pits also represent the best-preserved syn-impact morphological units and should represent the highest stratigraphic crater-related unit (Tornabene et al., 2012b).

Using the equation \(D_{ap} = 16.4D_c^{0.87}\) (where \(D_{ap}\) is the average of the 10 largest pit diameters in meters. \(D_c\) is the host crater’s diameter in kilometers) from Tornabene et al. (2012b), the average pit diameter of the ten largest pits for a crater ~29–32 km in diameter should be ~307–334 m – based on the best-preserved examples on Mars. Our average measurement was ~106 m from the 10 largest. This size difference may be the result of erosional processes not preserving the larger pits or aeolian infilling. This result also does not fit with expected diameter size as central floor pit craters generally fall above the trend line, which suggests that the size of crater-fill pits may be influenced by
central feature type (i.e., a larger volume of pitted deposit is accommodated by the lack of a central peak) or, if pit formation is related to the impact process, the target-types or conditions that are conducive to central floor pit formation (Tornabene et al., 2012b).

2.4.1.3 Post impact erosional units

Unconsolidated materials and Aeolian deposits are interpreted to be the youngest materials within the floor pit, as they overly the majority of the other units. Both of these units are formed by erosional processes, such as mass wasting (regolith) and wind saltation of grains (aeolian deposits), which affect exposure of impact generated units and bedrock.

2.4.1.3.1 Unconsolidated materials – regolith.

We interpret the unconsolidated materials to be regolith that forms as a result of weaker sediments eroding on steep slopes – up to ~46°, with an average of ~31° of the uplifted megablocks, as observed with the HiRISE DTM ArcGIS generated slope map. This interpretation supports the idea of cyclic episodes of lava and ash, where the weaker ash is eroded more easily than the strong basaltic bedrock. As this unit only makes up ~1.5% of the floor pit, we suspect that this is due to Betio crater having few steep slopes within its floor pit, thus, limiting mass wasting events from occurring in these limited high slope areas. These outcrops also have the second highest average thermal inertia of ~380 J m\(^{-2}\) K\(^{-1}\) s\(^{0.5}\), with respect to the other units, which is likely due to the eroded boulders containing light-toned bedrock and larger grain sizes.

2.4.1.3.2 Aeolian deposits– aeolian bedforms.

As we previously explained, this unit makes up ~10.4% of Betio crater central floor pit, forming linear Aeolian bed forms composed of accumulations of potentially sand-sized (~200 μm on Mars [Sullivan et al., 2005]) materials most likely transported through saltation, consistent with aeolian bed formations (Fig. 2.3G) observed elsewhere on Mars and on Earth. Confirmation of aeolian nature of these bedforms can be further established through a low THEMIS thermal inertia of ~205 J m\(^{-2}\) K\(^{-1}\) s\(^{0.5}\) that would suggest small sand size grains, similar to that of grain sizes of sand-sized particles, as noted above
(Fergason et al., 2006). As this unit covers all of the other morphologic units, we can interpret this unit as the youngest unit within Betio, which may still be actively forming. We also qualitatively view that orientations (~44° ENE) of these intra-crater aeolian bed formations seem to correspond with the direction of wind streaks in the immediate region, as observed with THEMIS IR of ~40–60°). These materials are important to note, as they likely contributed to the extent of the exposure of the bedrock within central uplifts on Mars (Tornabene et al., 2014).

2.4.2 Weakening mechanisms during cavity collapse and target uplift

The formation of a central uplift is a complex process that occurs during the end of the excavation stage into the crater modification stage (Melosh & Ivanov, 1999; Kenkmann et al., 2014), where two competing processes occur; a downward-directed gravitational collapse of the inner rim and an inward and upward movement of large (~100 m – 1 km scale) fault-bounded blocks from depth (Melosh, 1989; Melosh & Ivanov, 1999). To explain the magnitude of inward and upward motion of rocks, it has been proposed that a temporary strength degradation must occur (e.g., Melosh, 1977; McKinnon, 1978), which momentarily reduces friction both within a rock mass and along fault zones causing the rocks to deform. Geologists have benefited by using numerical simulations to better understand the impact process (e.g., Melosh & Ivanov, 1999; Collins et al., 2004; Senft & Stewart, 2009). However, numerical models do have limitations including being generally two-dimensional (assuming radial symmetry) and having limited spatial resolution, as deformation is a widespread process in nature (Kenkmann et al., 2005). Therefore, a direct correlation between model and field observations is difficult (Kenkmann et al., 2014). Despite these limitations, it is necessary to combine model and observation-based approaches to achieve better results.

One of the best developed models to explain this localized brittle faulting is the block oscillation model, whereby impact induced pressure vibrations in the target causes the target material to temporarily oscillate similar to waves in a fluid, although the material deforms as blocks of rock (see Ivanov & Kostuchenko, 1997). Important for the block oscillation model to be valid, “the sound speed of the matrix between blocks must be
much smaller than that of the intact rock” (Melosh & Ivanov, 1999), which translates into a megablock bounded by breccia zones ~10–20 % of the block’s thickness (i.e., ~100–200 m of breccia for a 1 km size block). Another earlier model known as acoustic fluidization has also been suggested by Melosh (1979) to explain this temporary weakening mechanism, whereby the strong shaking caused by the passage of impact-generated acoustic waves through fragmented rock debris reduces the overburden pressure, allowing fluidization. Other weakening mechanisms include thermal weakening (see O’Keefe & Ahrens, 1999) due to localized frictional heating, and strain rate weakening. The latter model predicts that the dominant weakening mechanism along faults at high slip velocities in target impact craters (100 km) in crystalline lithologies is frictional melting (Senft & Stewart, 2009). Our observations of Betio crater central uplift allow to make some comments on the weakening mechanisms, however calling on a single mechanism to dominate over crater collapse is not possible, as we suggest that multiple target-weakening processes are taking place as described below.

In contrast to terrestrial craters where exposure is typically <10% at best, in some areas of the studied craters, exposure is close to 100%. As previously described, the central floor pit contains an almost completely circumferential series of uplifted blocks (Fig. 2.13) with an average block size range of 100–300 m in diameter and an apparent increase in dip angle as a function of distance from the crater centre (Fig. 2.6). This observation is broadly consistent with observations of the terrestrial crater Jebel Waqf as Suwwan in Jordan (Kenkmann et al., 2010).

It can be difficult to determine the type of faulting that has taken place but we infer that there are some concentric thrust faults bounding the large blocks situated towards the outer portion of the pit, in addition to some radial extensional faulting, as evident in the collapsing and higher deformation in the centre (Fig. 2.8).

Within these layered blocks, a complex network of faults, folds, fractures, and dykes occur. As previously mentioned, folds appear larger and more symmetrical on the outer parts of the floor pit while radially striking smaller, tighter fold angles (0–30°; some isoclinical) antiforms and synforms with hinge lines often bent and plunging more steeply.
towards to centre. Folding is most likely accommodated along localized, small-scale brittle (micro-) faults (Fig. 2.9F), in agreement with studies from other terrestrial impact sites (e.g., Kenkmann, 2002; Lana et al., 2003). These folds are referred to as a radial transpression ridges (Kenkmann & Dalwigk, 2000) consistent with high strain rates and the non-ductile behaviour of rocks during an impact event. Some of the limbs of these folds are detached, which may be are the result of reverse faults in the core of these folds and their spread into the limbs to finally offset one of the fold limbs from the other. It has been noted in previous studies (e.g., Kenkmann et al., 2005) that the highest amount of brittle deformation is predicted to occur towards the centre. This is consistent with our observations.

Due to the exceptional exposure in Betio, we were able to map ~81 tabular, offset, crosscutting structures which we have interpreted to be impact generated breccia dykes within the uplifted blocks of the central floor pit. Terrestrial impact generated dykes are typically comprised of extensively cataclastic, deformed, and fluidized rock debris (with or without contribution of melted material) (e.g., Spray, 2001; Wittmann et al., 2004; Osinski et al., 2013). The macroscopically ductile appearance observed in these terrestrial dyke complexes is achieved through cataclastic flow that was initiated by the collapse of pore space, grain crushing, and consequential inter-granular shear (see Kenkmann et al., 2014). It is not possible to resolve with HiRISE whether the matrixes of these breccia dykes within Betio crater are comprised of melt or cataclastite.

The dykes within Betio are likely due to multistage emplacement and deformation during the impact process. Evidence supporting this is seen in the clast-rich nature, and multiple displacements within 67% of the dykes. We suggest that the majority of these dykes were emplaced prior to uplift (likely during excavation stage and growth of the transient cavity). As noted in the literature most dykes are radial to the centre of terrestrial (e.g., ~32 km Slate Islands [Dressler & Sharpton, 1997]) and some Martian craters (Kenkmann et al., 2014). However, not all of the dykes within Betio crater are radial (Fig. 2.12) which helps support the theory that the dykes were emplaced earlier in the impact process before the modification stage when the blocks were rotated and deformed along with observable displacements within the dyke networks. This supports the notion that the
deformation process in layered targets rather than in massive targets is more intense due to preferential weaknesses along the layers.

2.5 Conclusion

High-resolution observations of the Betio crater central uplift are similar to field observations of central uplifts on Earth – various types of impactites (breccias and melts [clast-poor, -free and -rich and pitted material], uplifted bedrock (paratochthonous LB), almost vertical blocks in central uplift center (75°–85°) and increased deformation towards the centre through impact-related structures, such as breccia dykes, radial transpression ridges and faults. Our observations support that the brittle failure was the prevailing deformation regime that occurred during the formation of the Betio crater central floor pit. Through the observations and measurements of brittle faulting, radial transpression ridges and dykes, we can determine that these structures likely aided in the collapse of the central uplift in the centre and supported target weakening. These structural deformations supported with a cyclically layered volcanic target could account for the formation of a central pit but this needs to be further explored through the analysis of additional floor pit craters on Mars.
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Chapter 3

3 A Morphologic and Structural Comparative of Three Mid-sized Floor Pit Craters in Thaumasia Planum, Mars

3.1 Introduction

Despite the ubiquitous nature of impact craters in the solar system, there are still some important aspects of the processes and products of their formation that remain relatively poorly understood. One such aspect is the formation of central uplifts in complex impact structures, particularly the unusual central uplift morphology known as a central pit found on Mars (as was introduced in Chapter 2). Central uplifts are of considerable scientific interest as they exhume deep-seated bedrock from the subsurface that is not otherwise exposed at the surface. As noted previously, very few of these complex impact structures are well preserved and/or well exposed on Earth (Kenkmann et al., 2005, 2014). Therefore, we must look to other bodies in our solar system. Now that accessibility to higher resolution orbital data is available for Mars and the Moon, it is now possible to conduct detailed mapping and structural studies of central uplifts on Mars (e.g., Ding et al., 2014; Chapter 2). Due to the various factors that influence the impact cratering process (e.g., impact angle, velocity, target properties), as on Earth (Osinski et al., 2013), the comparative study of central uplifts that have formed in almost identically sized craters and target lithologies may be particularly informative.

In this study, we carried out morphological and structural mapping and analysis of central uplifts that expose Layered Bedrock (LB) (see Tornabene et al., 2010 and Caudill et al., 2012) within three mid-sized Martian impact craters of similar size (~30 km diameter) – Betio (also see Chapter 2), Byala, and Halba. Comparison studies allow for more reliable assessments regarding the level of preservation, emplacement of various impactites and structural deformations. We provide a synthesis of our observations and interpretations of the craters in a comparative manner by utilizing our results and interpretations in Chapter 2 of the morphologic and structural observations, then outline similarities and differences within each of the structures to provide further insights into the weakening mechanism associated with central uplift formation and ultimately the formation of a
3.2 Overview of the Betio, Byala, and Halba impact craters

Figure 3.1: (Left) Context view the 3 craters as seen in MOLA elevation and geologic map. (Right) Geologic map of Thaumasia Planum (Scott & Tanaka, 1986). Both Byala and Halba boarder Noachian aged terrains and Betio is more central in the Hesperian ridged plains (Hr) unit. This ultimately will have some effect on the resulting morphology and exposed bedrock units, as will be discussed below.

Tornabene et al. (2010), in a Crater-Exposed Bedrock (CEB) database, initially characterized Betio, Byala, and Halba as all possessing layered bedrock (LB). These craters are all at the approximately same latitude at 25° S and are all separated by ~10° longitude (~600 km), situated ~830 km south of Valles Marineris (refer to Table 3.1 for latitude and longitude of each crater) (Fig. 3.1). These craters have many similar qualities that make them excellent candidates for a comparison study. First, they are all ~30 km in diameter (refer to Table 3.1 for a complete list of exact diameters) and they exhibit Layered Bedrock (LB) morphologies. Second, they all formed in Hesperian-aged Ridged Plains Material units (Scott & Tanaka, 1986). This unit is interpreted as a volcanic unit, containing a surface morphology known as wrinkle-ridges, which are commonly accepted to form when contractional deformation (Plescia & Golombek, 1985; Watters, 1988), that differentially affects the underlying and surface stratigraphy having differing
layer strengths (Golombek et al., 2001). We observe wrinkle ridges near or possibly intersecting each of the craters (indicated in Figs. 3.2A, 3.3A, 3.4A). This may have an effect of the formation of the crater and its structures. They display floor pit central uplift morphologies, defined as a depression into the crater floor (e.g., Wood et al., 1978; Barlow, 2010). They are common on Mars, Callisto, and Ganymede and were first identified on Mars using Mariner 9 (Smith, 1976) and Viking (Hale & Head, 1981; Hale, 1982, 1983; Hodges et al., 1980) images. However it is important to note that the diameter of these floor pits varies from smallest Betio, then Byala, then Halba, based on topographic maps (refer to Table 1 for measurements and Figs. 3.2, 3.3, 3.4, 3.5). More details on the central structure will be given in Section 3.4.3. Finally, a symmetrical Multiple Layer Ejecta (MLE) blanket is observed around Betio and Byala craters that reaches as far as ~54 km (Fig. 3.2A) and ~25 km (Fig. 3.3A), respectively. These are suspected to have formed as the result of two mechanisms either by impact heating and vaporization of subsurface volatiles during crater formation (Carr et al., 1977) or interactions of the ejecta with the thin Martian atmosphere (Barnouin Jha & Schultz, 1998). Halba crater’s ejecta blanket (Fig. 3.4A) is difficult to distinguish, as it is suspected have undergone a higher degree of erosion.

Each crater does display other minor differences. Firstly, regional minimum flood lava thickness estimated by Caudill et al. (2012) puts each crater in varying lava thicknesses. The thickest is at the borderer between Thaumasia Planum and Solis Planum, where Betio crater is located (~7.7–13.9 km), and become progressively thinner in Thaumasia Planum, where Byala is located (~7.7 km) and thinnest in Bospherus Planum where Halba is located (~4.6 km). This ultimately will have an affect on the units that are exposed in the central uplift, as the crater size increases and deeper units are exhumed. Halba crater’s rim is overlaying another ~29 km unnamed complex crater on the north-eastern side (Fig. 3.4A) that may have influenced its rim and possible central uplift during formation. Next, Byala and Halba are located near Noachian aged geologic units (Fig. 3.1). This likely will have some influence on the target sequence. While Betio crater is in the middle of Hr unit and may not have “touched” Noachian crust underneath. Each crater rim sits at a different altitude (refer to Table 3.1 for measurements), as indicated by MOLA elevation data. Byala (average altitude of ~3175 m) crater being the
highest, Betio (average altitude of ~2887 m) close behind, and Halba (~1575 m) at the lowest as it is on a slope. This may have some effect on the thickness of the underlying layered volcanics at the time of impact. Utilizing estimates for depth/diameter (d/D) shortly after the time of impact (d=0.302D^{0.52} [Stewart & Valiant, 2006]) and current measured d/D for each crater (using MOLA data), we were able to calculate the estimated amount of erosion within the crater since impact. Byala crater has the calculated least amount of erosion at 62 m, Betio crater at 328 m (note this measurement may be less if accounting potentially smaller diameter – discussed in Chapter 2), and Halba crater at 401 m (refer to Table 3.1 for the complete list of measurements). These assessments will be looked at further in the morphological and structural analysis. Finally, both Betio crater and Halba crater contain a fully rimmed floor pit (Garner & Barlow, 2012), with exposures of uplifted blocks surrounding the pit, while Byala crater is only partially rimmed with uplifted blocks observed in the west and north.
Figure 3.2: (A) Context view of Betio crater seen in THEMIS DAY IR. It is evident that two wrinkle ridges may have affected the formation of this crater (NW and S of the rim). Also it has a multiple layer ejecta blanket that is slightly asymmetrical. (B) Betio crater seen in CTX (B20_017583_1571XN_22S078W). Approximately 500 m to the east of the floor pit, there is a ~4.5 km diameter over-printing simple crater lacking a discernable ejecta blanket. (C) A cross-section of Betio crater using MOLA elevation data.
Figure 3.3: (A) Context view of Byala crater seen in THEMIS DAY IR. It is evident that two wrinkle ridges may have affected the formation of this crater (note the wrinkle ridge in the NE and SW). Also a MLE blanket surrounds the crater. (B) Byala crater seen in CTX (G09_021578_1541_XN_25S066W). The floor pit is not as discernable as Betio and Halba but becomes more apparent in elevation data as seen in Fig. 3.5. (C) A cross section of Byala crater using HRSC elevation data (h4204_0000_da4).
Figure 3.4: (A) Context view of Halba crater seen in THEMIS DAY IR. It is evident that Halba crater has impacted ontop a pre-existing complex crater. It is also quite difficult to see the ejecta blanket. There are also very prominent large light toned fractured bedrock in a unit on the western side of Halba and a wrinkle ridge in the south-east portion of Halba appearing to terminate in the at the rim. (B) Halba crater seen in CTX (B08_012901_1546_XN_25S056W). It is evident that the crater floor is quite variable and uneven. (C) Cross section of Halba crater using HRSC elevation data.
Figure 3.5: (A) Betio crater floor pit topography illustrated using a HiRISE DTM, with an applied colourized stretch for the eastern portion of the uplift (DTEEC_016805_1565_017583_1565) Min: 1106 m, Max: 1495 m and contours. (B) Byala crater floor pit topography illustrated, using an HRSC DTM (h2486_0001_da4), with a rainbow colourized elevation of Min: 1400 m, Max: 2100 m and contours. (C) Halba crater floor pit topography illustrated, using an HRSC DTM (h4204_0000_da4), with a rainbow colourized elevation of Min: -534 m, Max: 265 m, and contours. Please note the differences in scale.

Table 3.1: Summary of the main attributes of the Betio, Byala, and Halba craters

<table>
<thead>
<tr>
<th>Lat/Long</th>
<th>Betio</th>
<th>Byala</th>
<th>Halba</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geographic Area</td>
<td>281.38°E, -23.15°N</td>
<td>293.54°E, -25.75°N</td>
<td>303.91°E, -26.03°N</td>
</tr>
<tr>
<td>Crater Diameter (km)</td>
<td>31.7</td>
<td>27.5</td>
<td>32.7</td>
</tr>
<tr>
<td>Average Rim Altitude (m)</td>
<td>2887</td>
<td>3175</td>
<td>1578</td>
</tr>
<tr>
<td>Max Rim Altitude (m)</td>
<td>3086</td>
<td>3277</td>
<td>1721</td>
</tr>
<tr>
<td>Average Crater floor Altitude (m)</td>
<td>1494</td>
<td>1545</td>
<td>128</td>
</tr>
<tr>
<td>Min-Max Crater floor Altitude (m)</td>
<td>1540-1440</td>
<td>1453-1637</td>
<td>(-) 54-270</td>
</tr>
<tr>
<td>Estimated d/D</td>
<td>(Stewart and Valiant, 2006)</td>
<td>1822</td>
<td>1692</td>
</tr>
<tr>
<td>Actual rim to floor depth (m)</td>
<td>1293</td>
<td>1630</td>
<td>1450</td>
</tr>
<tr>
<td>Difference of estimated d/D and measured d/D (m)</td>
<td>328</td>
<td>62</td>
<td>401</td>
</tr>
<tr>
<td>Floor-pit diameter NW-SE (km)</td>
<td>10.8</td>
<td>12.1</td>
<td>13</td>
</tr>
<tr>
<td>Floor pit diameter NE-SW (km)</td>
<td>8.8</td>
<td>10</td>
<td>12</td>
</tr>
<tr>
<td>Area of floor-pit (km²)</td>
<td>67</td>
<td>107</td>
<td>97</td>
</tr>
</tbody>
</table>

3.3 Data and methods

This study followed the same methodology and used the same data products as in Chapter 2. Mapping was conducted primarily on a combination of orthorectified (where available) and non-orthorectified HiRISE images (Table 3.2). The HiRISE images were augmented
by various other data products (Table 1) including, elevation data (Digital Terrain Models [DTMs] derived from the Mars Orbiter Laser Altimeter [MOLA~462 m/pixel]; [Smith et al., 2001], High Resolution Stereo Colour Imager [HRSC~150-175 m/pixel] [Neukum & Jaumann, 2004]), visible (~18 m/pixel) and thermal infrared (~100 m/pixel) images from the Thermal Emission Imaging System (THEMIS) (Christensen et al., 2004), and context camera (CTX) images (~6 m/pixel) (Malin et al., 2007). In addition, we utilized THEMIS-derived thermal inertia measurements (Fergason et al., 2006; Putzig & Mellon, 2007).

Table 3.2: List of data used for each crater

<table>
<thead>
<tr>
<th>Camera</th>
<th>Betio</th>
<th>Byala</th>
<th>Halda</th>
</tr>
</thead>
<tbody>
<tr>
<td>HiRISE</td>
<td>ESP_027987_1565</td>
<td>ESP_021578_1540</td>
<td>ESP_017002_1535</td>
</tr>
<tr>
<td></td>
<td>ESP_016805_1565</td>
<td>ESP_020734_1540</td>
<td>ESP_017147_1535</td>
</tr>
<tr>
<td></td>
<td>ESP_017583_1565</td>
<td>PSP_008761_1540</td>
<td>ESP_030808_1535</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ESP_030782_1540</td>
<td>ESP_031230_1535</td>
</tr>
<tr>
<td>HiRISE DTM</td>
<td>DTEEC_016805_1565</td>
<td>No</td>
<td>DTEEC_017002-031230_1535_50cm</td>
</tr>
<tr>
<td>HRSC DTM</td>
<td>No</td>
<td>h2486_0001_da4</td>
<td>h1204_0000_da4</td>
</tr>
<tr>
<td>CTX</td>
<td>P07_003872_1547_XN_25S078W</td>
<td>P20_008761_1544_XN_25S066W</td>
<td></td>
</tr>
<tr>
<td></td>
<td>D02_027967_1567_XN_23S078W</td>
<td>G09_021578_1541_XN_25S066W</td>
<td></td>
</tr>
<tr>
<td></td>
<td>B20_017583_1571_XN_22S078W</td>
<td>G06_020734_1540_XN_26S066W</td>
<td></td>
</tr>
<tr>
<td></td>
<td>B18_016805_1567_XN_23S078W</td>
<td>B08_012901_1546_XN_25S056W</td>
<td></td>
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<tr>
<td></td>
<td>B03_010647_1546_XN_25S078W</td>
<td>B19_017002_1538_XN_26S056W</td>
<td></td>
</tr>
</tbody>
</table>

Orientation (i.e., strike) of the faults, fractures, and dykes were measured using the coordinate geometry (COGO) report tool, in the editor extension in ArcMap using the right hand rule. Strike and dip values for the layers within uplifted megablocks were geometrically calculated using the Layertools extension for ArcGIS (Kneissl et al., 2010) on a 2m/pixel HiRISE DTM (Mattson et al., 2011). It should be noted that dip measurements using the layer tool in ArcMap were only conducted where HiRISE DTMs were available (i.e., the eastern half of Betio crater floor pit and almost the full extent of Halba crater floor pit [refer to table 3.2 for complete Id’s]). Where high-resolution DTMs were not available, dip measurements were estimated based on stereo-images and 3D viewing in ArcScene. Each block was measured ~10–20 times, depending on the size (i.e., more measurements for larger blocks), and an average was then calculated from the multiple measurements. These values measured for each linear feature were plotted on
an equal area stereonet projection, using either planes or rose diagram in OSX Stereonet v. 9.0.1. (Allmendinger et al., 2013).

3.4 Observations and interpretations of mapped units

Detailed morphological maps of the central pits of Betio, Byala, and Halba crater are provided in Figures 3.14, 3.15, 3.16, respectively. Units were previously defined and interpreted based on surface appearance (tonality, surface textures, structures, features (HiRISE image [~25 cm/pixel]), thermophysical properties of the surface, average THEMIS (~18–100 m/pixel) thermal inertia (Table 3.3) value (size of the particles), elevation maps outlining the units height (Fig. 3.5) and their average slope (HiRISE DTM [~2 m/pixel]) (Table 3.3) (steepness of a unit in degrees). Each unit was also qualitatively observed in various stages of erosion and preservation, which is noted therein and will be discussed in section 3.5.2. Below, we outline the different morphologic units observed in the three craters, using Betio as a foundation. Observations for each unit are summarized in Table 3.3.

3.4.1 Morphologic units present in all three craters

Within Betio, Byala, and Halba craters there were many similar unit identifications made based on previous interpretations made in Chapter 2 for Betio crater.

3.4.1.1 Layered bedrock

This unit is characterized by outcrops of alternating layers of generally thinner and higher-standing dark-toned layers and thicker lower-standing, light-toned, fractured, folded, and faulted bedrock layers (Fig. 3.6), consistent with the observations from Tornabene et al. (2010, 2012a) and Caudill et al. (2012). The layers are not sedimentary in origin but are interpreted to be volcanic, based on their combined morphologic and spectral attributes and the geologic setting and geographical distribution of LMB central uplifts (Caudill et al., 2012; Quantin et al., 2012; Tornabene et al., 2010; Wulf et al., 2012). This unit exposure varies, with respect to each crater and location within the floor pit. Similar percentages of this unit are observed and mapped in Betio (Figs. 3.6A, 3.7) and Byala (Fig. 3.6B) (~15% and 13%, respectively); however, the level of exposure is
best within Betio and less so in Byala. The amount of layered bedrock almost doubles for Halba (Fig. 3.6C) at ~26% of the total central uplift area. The layered bedrock in Halba also appears to be more densely fractured than the other two craters with less preserved layers, possibly suggesting a deeper level of erosion (discussed further in section 3.5.2).

Figure 3.6: Exposed layered bedrock (A) Betio crater floor pit east central outcrop, showing excellent exposure (HiRISE image ID: ESP_016805_1565) (please refer to Appendix A.2 for context view). (B) Byala crater floor pit west central outcrop, showing less exposure of light-toned outcrop but still intact layering (HiRISE image ID: ESP_021578_1540_RED) (please refer to Appendix A.3 for context view). (C) Halba crater floor pit north central outcrop, showing well exposed but degraded layers being offset by a fault (HiRISE image ID: ESP_031230_1535) (please refer to Appendix A.4 for context view).

3.4.1.2 Impact melt units

On Earth, impact melt typically occurs as a rock containing variable amounts of target rock clasts and can be clast-rich, -poor or -free (Osinski et al., 2013). On Mars, we have classified three types of impact melt units (based on ~25 cm/pixel resolution of HiRISE images), these include: clast-free -poor, clast-rich, and pitted materials.

3.4.1.2.1 Clast-free -poor impact melt

In Chapter 2, we interpreted the dark-toned smooth unit as clast-free -poor impact melt (based on the ~25 cm/pixel resolution of HiRISE). It appears to preserve overprinting craters, suggesting it is coherent in nature (also see Table 3.3). This unit displays evidence for ponding, flowing, and formation of coherent sheets on low slopes (~2–15°).
which mimic observations of melt deposits at other craters on the Moon (e.g., Hawke & Head, 1977), Earth (e.g., Grieve et al., 1977; Grieve & Cintala, 1992) and Mars (e.g., Marzo et al., 2010; Osinski et al., 2011; Tornabene et al., 2010, 2014; Ding et al., 2013). This unit is observed in all three craters, with varying levels of exposure and coverage. This unit is most prominent within Betio crater, where it is present throughout much of the central uplift covering ~62 % (Figs. 3.7, 3.8A) of the total area, as well as the crater floor (Fig. 3.7). It is much less prominent in Byala (Fig. 3.8B) and Halba (Figs. 3.8C, 3.11), covering only ~6 % and 7 %, respectively. It is important to note that these estimates are conservative, as there are clearly some aeolian deposits on top of this unit (Fig 3.7).

Figure 3.7: Orthorectified red mosaic image ESP_016805_1565 draped over a stereo-derived DTM (DTEEC_01805_017583_1565) with the vertical exaggeration (VE) set to 1.5x. (B) Close-up showing exposed layered bedrock with faults, folds, and an off-set dyke going through the hinge of a fold.
Figure 3.8: Dark-toned smooth unit interpreted as clast-free impact melt. (A) Betio crater: north-eastern ponded material (HiRISE image ID ESP_016805_1565) For context refer to appendix A.2. (B) Byala crater: south-central region of floor pit (HiRISE image ID: ESP_021578_1540_RED). For context refer to appendix A.3. (C) Halba crater: north-central region of the floor pit exhibiting flow banding (larger image in Fig. 3.11) (HiRISE image ID: ESP_031230_1535). For context refer to appendix A.4.

3.4.1.2.2 Clast-rich impact melt units

A breccia is a rock that is a mixture of angular fragments from different types of rocks surrounded by a fine-grained matrix that may be similar to or different from the fragmented material, as previously described in Chapter 2. Impact breccias and suevites both contain melt derived from the melting of target rocks, however, not all breccias contain melt. The breccia that contains no melt is referred to as impact breccia. It is not possible to ascertain whether the matrix of these units is melt or clastic but, based on some surface features such as evidence of flow textures (flow, ponding, channel, etc.), we can suggest if the matrix of some of these outcrops may be melt. Typically this unit is observed to be relatively clast-poor and smaller clast size <1–5 m, when observed at higher elevations, and becomes more clast-rich, where it borders or embays portions of the layered bedrock unit or uplifted megablocks. An alternative explanation is that some of the narrower elongated zones between uplifted blocks (parallel to the fault plane) of this unit, (seen in Halba crater [Fig.3.9AB]), may represent fault breccias (or breccia
dykes).

In Betio and Byala this unit represents ~4 and 4.5 %, respectively of the total floor pit area, typically in areas bounding blocks and at the “scarp” (Figs. 3.14, 3.15). Halba crater has a much more pronounced occurrence of the unit. Outcrops are narrower elongated zones between uplifted blocks, which may represent fault breccias and comprise 9 km² and ~8.5 %. Outcrops of what are interpreted as crater fill deposits cover ~31 km² and ~29.5 % (together ~38 %) are found in a smooth and/or rough textured dark-toned matrix surrounding the outer portion of the floor pit that may represent a clast-rich impact melt (Fig. 3.9C).

Figure 3.9: (A) Halba Crater mosaicked DTM (2m resolution, height 261 to -535 m), with mosaicked HiRISE images (ESP_017002_031230_1535_50cm) with a VE of 1.5x. View of the south-eastern section, showing the different location and types of breccias in the uplift. (B) A clast-rich material (possibly dykes) bounding two uplifted blocks (ESP_017002_1535). (C) A clast-rich melt unit interpreted to be part of the crater-fill, that surrounds the crater in a moat, and may have been exposed by erosional forces (ESP_017002_1535).
3.4.1.2.3 Pitted material

This unit is characterized as a dark-toned unit, containing quasi-circular (honey-comb-like) clusters of subdued pits (Fig. 3.10) and varying in preservation from crater to crater. Some pits contain aeolian deposits within the pits, along with what appears to be light-toned eroded small boulders or rocks around the “rims” of the pits (Tornabene et al., 2012b). This unit is located in relatively low-lying topographic areas that are straigraphically higher than the dark-toned smooth unit. Crater-related pitted materials likely represent the best-preserved impact-generated morphological units that likely represent the stratigraphically highest portion of the crater fill deposits (Tornabene et al., 2012b). They are thought to have been emplaced as hot impact melt-bearing deposits that entrained volatiles from volatile-rich target materials. This unit is the most prominent within Byala, where it comprises the majority of the floor pit ~65 km² (~60.5 %) surrounding the pit and uplifted megablocks (Fig. 3.10C). In Betio crater, this unit is located between the floor pit boundary and the scarp, with the best-preserved clusters in the south and north-east, covering ~3 km² and making up 4.5 % of the total floor pit area (Fig. 3.10B). Lastly, in Halba, this unit is almost non-existent within the floor pit boundary, comprising only 0.5% of the total area in the NE on a higher elevation portion of crater floor.
Figure 3.10: Interpreted pitted material (A) An example of nearly pristine crater-related pitted materials in Tooting crater (northern fragment of the crater floor) (HiRISE image ID: PSP_001538_2035). (B) Betio crater best preserved pitted material on the south-eastern side, near scarp rim (HiRISE image ID: ESP_016805_1565). For context of image please refer to appendix A.2. (C) Byala crater, north-western section of crater floor just outside the floor pit (HiRISE image ID: ESP_021578_1540_RED). For context of image please refer to Appendix A.3.

3.4.1.3 Undifferentiated material

This unit contains a mixture of dust, aeolian and dark-toned relatively smooth textured material. It is distributed on over most of the central uplift in Halba crater, making it 15 km$^2$ and 14.5% of the total area. In Byala crater, it is located on the eastern side, where there are almost no uplifted blocks. This unit is difficult to determine a distinct interpretation, as it is most likely a combination of units that have been obscured by dust coverage.

3.4.1.4 Post impact erosional units

Regolith (brown unit) and aeolian deposits (beige unit) are interpreted to be the youngest materials within the floor pit, as they overly the majority of the other units. Both of these units are formed by erosional processes, such as mass wasting (regolith) and wind saltation of grains (aeolian deposits). In Betio crater, regolith is a very minimal unit, covering ~1 km$^2$ and ~1.5 % of the total central floor pit and sometimes deposited in areas of lower slope. In Byala crater, this unit has a similar occurrence as Betio crater, comprising ~2 km$^2$ and ~2 % of the total area and is typically located on east facing
slopes in the west. In Halba crater, it is the most prominent out of the three craters, covering 7.5 km² and 7% of the total floor pit (Fig. 3.11) and is situated on east-facing slopes in the west, similar to Byala crater.

Aeolian deposits are typically situated in topographic depressions throughout the central structures, covering the lowest portions of the pits, and within the depressions between the uplifted blocks and overlaying the majority of the other units. These materials are important to note, as they likely contributed to the extent of the exposure of the bedrock within central uplifts on Mars (Tornabene et al., 2014). In Betio, this unit covers ~7 km² and ~10.4% of the total central uplift area normally trending in a NE-SW direction. We also qualitatively view that orientations (~44° ENE) of these intra-crater aeolian bed formations, which correspond with the direction of wind streaks in the immediate region as observed with THEMIS Night IR of ~40–60°). We also include, herein, a facies of this unit that contains aeolian deposits, with some exposure of layered bedrock unit just underneath. These areas are found on pit floors, with slightly higher thermal inertias of ~255 J m⁻² K⁻¹ s⁻¹/₂. In Byala crater, this unit covers ~6 km² and ~5.5% of the total central uplift area, typically trending in a NE-SW direction. In Halba crater, this unit covers ~6.5 km² and ~6% of the total central uplift area typically trending in an EW–SE direction. We should note that these estimates are likely higher, as aeolian deposits overlie the majority of units but were mapped based on dominance within a polygon.
Table 3.3: Overview of similar morphologic units observed in all three craters

<table>
<thead>
<tr>
<th>Observation Name</th>
<th>Interpretation</th>
<th>Description</th>
<th>Possible Formation Mechanism</th>
<th>Map Colour</th>
</tr>
</thead>
<tbody>
<tr>
<td>Layered dark and light toned outcrops</td>
<td>Layered Bedrock</td>
<td>Outcrops of alternating layers of generally thinner, and higher-standing dark-toned layers and lower standing, light-toned layers (Tornabene et al., 2010, 2012a; Caudill et al., 2012) that have been fractured, folded, and uplifted in megablocks. CRISM spectral analysis found olivine and high-calcium pyroxene spectral signatures — consistent with basalt (Quantin et al., 2012) and high thermal inertia — averages for each crater Betio (~422 J m(^{-2}) K(^{-1}) s(^{-1})), Byala (~395 J m(^{-2}) K(^{-1}) s(^{-1})), Halba (~455 J m(^{-2}) K(^{-1}) s(^{-1}))</td>
<td>Cyclic volcanism producing alternating flood basalts and ash deposits emplaced largely during the Hesperian Era (Caudill et al., 2012).</td>
<td>Light-green</td>
</tr>
<tr>
<td>Dark-toned smooth unit</td>
<td>Clast-free – poor impact Melt</td>
<td>Dark-toned with a relatively smooth texture that appears to preferentially preserve several overprinting craters suggesting it is rigid and coherent in nature. Also observed embaying, coating, pooling, and flowing typically on low slopes 3-13(^{\circ}). Mid range thermal inertia — averages for each crater Betio (~230 J m(^{-2}) K(^{-1}) s(^{-1})), Byala (~225 J m(^{-2}) K(^{-1}) s(^{-1})), Halba (~240 J m(^{-2}) K(^{-1}) s(^{-1}))</td>
<td>Melted target rock, emplaced during the excavation stage (Tornabene et al., 2010; Marzo et al., 2010; Osinski et al., 2011; Skok et al., 2012)</td>
<td>Light-orange</td>
</tr>
<tr>
<td>Light-toned clast-rich unit</td>
<td>Impact melt breccia/Impact breccia/ Brecciated Dykes</td>
<td>Contains light-toned (~4-60 m), angular and subrounded rock fragments (i.e., clasts) embedded in a dark toned matrix (melt or smaller brecciated material). Can be relatively clast-poor and smaller &lt;1-5 m at higher elevations and becomes more clast-rich where it borders or engulfs portions of the layered bedrock unit or uplifted megablocks. Often appears to be “channelized” within long and narrow topographic lows that are between high-standing ridges of bedrock (megablocks). An average thermal inertia ~278 J m(^{-2}) K(^{-1}) s(^{-1}) is measured at all three craters.</td>
<td>Allochthonous breccias that covered the crater floor and subsequently crowded together in the crater moat and in local depressions in the topography of the central uplift (in a melt of finer brecciated material matrix). The breccia formation may be the result of shear movements of the large blocks moving past one another during crater formation. Alternatively they may be the product of breccia injections into tensile gaps that periodically open during block movements and uplift of the crater floor (Kernkamm et al., 2014).</td>
<td>Red and coral (Halba)</td>
</tr>
<tr>
<td>Quasi-circular pit-bearing dark-toned unit</td>
<td>Pitted Material</td>
<td>Dark-toned quasi-circular (honey-comb-like) clusters of eroded subdued pits. Varying in preservation-some containing aeolian deposits within the pits along with what appears to be light-toned eroded small boulders or rocks around the “rim” of the pit. Located in relatively flat areas that are stratigraphically higher than the clast-poor impact melt. Average slope of ~4.5(^{\circ}); Average thermal inertia of ~285 J m(^{-2}) K(^{-1}) s(^{-1}) in both Betio and Byala.</td>
<td>Interpreted to be a eroded/modified version of the crater-related pitted materials described by Tornabene et al. (2012b), which is described as “Emplaced as hot impact melt-bearing deposits that may have entrained volatiles from volatile-rich target materials” (Tornabene et al., 2012b)</td>
<td>Aqua</td>
</tr>
</tbody>
</table>
Table 3.3: Overview of similar morphologic units observed in all three craters, continued.

<table>
<thead>
<tr>
<th>Observation Name</th>
<th>Interpretation</th>
<th>Description</th>
<th>Possible Formation Mechanism</th>
<th>Map Colour</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unconsolidated Materials</td>
<td>Regolith</td>
<td>Various sized light- and dark-toned boulders up to a few metres in diameter observed on steep slopes (up to 46º with an average of 31º) in sections of the layered exposed bedrock. Second highest thermal inertia of (consist average is all three craters ~380 J m⁻² K⁻¹ s⁻¹)</td>
<td>Mass-wasting on steep slopes past the angle of repose (~30º)</td>
<td>Brown</td>
</tr>
<tr>
<td>Aeolian Deposits</td>
<td>Aeolian Bedforms</td>
<td>Linear beds composed of accumulations of approximately unconsolidated materials exhibiting various bedforms. Typically in topographic depressions throughout the central structures, covering the pits in the centre, and in the depressions between the uplifted blocks. Overlying the majority of the other units and typically in areas of low slope ~4.3º. Lowest thermal inertia (average for all three craters ~205 J m⁻² K⁻¹ s⁻¹)</td>
<td>Saltation of fine grains Beige (Sullivan et al., 2005; Ferguson et al., 2006)</td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.11: (A) Halba crater mosaicked DTM (2m resolution, height 261 to -535 m), with mosaicked HiRISE images (ESP_017002_031230_1535_50cm) with a VE of 1.5x. View of the eastern side. Exposed bedrock, aeolian beds and talus. (B) Subset of clast-free impact melt showing flow banding (indicated with yellow arrow).
3.4.2 Other morphologic units

3.4.2.1 Light-toned polygonal unit

This unit is exclusive to Byala and is comprised of light-toned polygonal-shaped fractures of bedrock found within the centre of the central depression of the floor pit (Fig. 3.12A, 3.12B). This unit is stratigraphically the lowest unit observed, comprising of 1.2 km$^2$ and 1% of the total area and with an average thermal inertia of $\sim$377 J m$^{-2}$ K$^{-1}$ s$^{1/2}$. These polygons could be the result of pre-existing or post-impact related processes. Polygons in deposits and within bedrock can form in numerous ways, including cooling and contraction of a melt-sheet (e.g., Tornabene et al., 2012a), ice related freeze-thaw cycles (e.g., Mellon et al., 2008), and alteration of bedrock into phyllosilicates, sulfates or carbonates (e.g., Bishop et al., 2008; Weitz et al., 2012; Ehlmann et al., 2011). However, the scale of these polygons is not consistent with cooling and contraction patterns in some lunar and Martian craters (e.g., Tornabene et al., 2012a). Freeze-thaw mechanisms are not consistent with the elevation and latitude of these craters – even during high-obliquity events. Furthermore, after some preliminary analysis of CRISM spectral parameter images, we did not observe any alteration-related minerals within this unit, including phyllosilicates, mono- or poly-hydrated sulfates or carbonates. Further, detailed spectral analysis is suggested to better understand the composition of these polygons and their possible origin.

3.4.2.2 Light-toned eroded outcrops of polygonal material

This unit is exclusively found in Byala and appears as an eroded light-toned bedrock, with occurrences of linear features that intersect into polygonal fractures, as well as relatively small 1 to 10 m dark-toned holes giving it a “swiss cheese” appearance. Surrounding these outcrops, there are dark-toned unconsolidated materials. This unit is found in five irregularly shaped outcrops in the centre of the floor pit, ranging from 600 to 1100 m in diameter and covering $\sim$3.3 km$^2$ of the total floor pit (Figs. 3.12A,C & 3.13 Table 3.3). This unit is observed stratigraphically higher and/or at equal elevation as the interpreted clast-free-poor impact melt and at similar levels as the pitted material. In the merged IRB HiRISE product, these outcrops appear to be dark blue, possibly indicating a...
basaltic nature. The thermal inertia for this unit ranges from \(~400-449\) J m\(^{-2}\) K\(^{-1}\) s\(^{\frac{1}{2}}\) for the light-toned outcrops, then drops to \(345\) to \(399\) J m\(^{-2}\) K\(^{-1}\) s\(^{\frac{1}{2}}\) for the darker-toned unconsolidated materials surrounding the units (Fig. 3.12 & Table 3.3). The only features that are seen overprinting this unit are aeolian beds. Due to this, we suggest that this may be eroded and/or post-impact altered pitted material, possibly composed of weaker material that has since eroded away; however, further spectral analysis is required.

Figure 3.12: (A) An eroded and altered outcrop of what is interpreted as pitted material seen in the centre of the central depression (Images from HiRISE: ESP_021578_1540_MIRB). (B) Zoom in northwest of the outcrop - a light toned polygonal outcrop is observed. (C) A zoom-in of the eroded pitted material unit. (Please note the differences in scales).
Figure 3.13: Graph showing a comparison of percentage of each floor pit covered in each morphologic unit.

Table 3.4: Morphologic Units within each crater indicate with an X.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Betio</th>
<th>Byala</th>
<th>Halda</th>
</tr>
</thead>
<tbody>
<tr>
<td>Layered Bedrock</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Polygons</td>
<td></td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>Clast-poor impact melt</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Clast-rich impact melt/breccia</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Pitted material</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aeolian deposits</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Regolith</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
</tbody>
</table>
Figure 3.14: Betio crater geomorphologic map with 35% transparency, overlain onto a mosaicked HiRISE images (ESP_016805_1565_RED_A_01_ORTHO with ESP_027987_1565_RED).
Figure 3.15: Byala crater geomorphologic map with 35% transparency, overlain onto a HiRISE images (ESP_021578_1540 and ESP_030782_1540).
Figure 3.16: Geomorphologic map of Halba crater with 35% transparency, overlain on HiRISE images (ESP_017002_1535 and ESP_031230_1535).
3.4.3 Structural assessment

3.4.3.1 General floor pit structure

Betio, Byala, and Halba craters’ central floor pits each have a distinctly different quasi-circular shape, extent, and overall morphologies (Table 3.5). However, there are some consistent features that have been previously defined in Chapter 2 for Betio crater. These consistent features include: relatively smooth floors (at the lowest elevations in the crater – observed in Betio and Halba [slightly larger in diameter and area]) (Fig. 3.5 & Table 3.5), the apparent floor pit rim (Fig. 3.5), and uplifted exposed blocks surrounding (Betio and Halba) or semi-surrounding (Byala) the central depression.

3.4.3.2 Uplifted blocks

Table 3.5: Betio, Byala, and Halba craters’ central floor pit details

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Betio</th>
<th>Byala</th>
<th>Halda</th>
</tr>
</thead>
<tbody>
<tr>
<td>Floor pit diameter based on topographic data (NW-SE, NE-SW) km</td>
<td>10.8 x 8.8</td>
<td>12.1 x 10</td>
<td>13 x 12</td>
</tr>
<tr>
<td>Area of floor-pit (km²)</td>
<td>67</td>
<td>107</td>
<td>97</td>
</tr>
<tr>
<td>Max Depth of pit from crater floor (m)</td>
<td>1110</td>
<td>1453</td>
<td>-406</td>
</tr>
<tr>
<td>Difference between crater floor altitude and max depth (m)</td>
<td>484</td>
<td>92</td>
<td>534</td>
</tr>
<tr>
<td>Morphologic Map Floor pit diameter (NW-SE, NE-SW) km</td>
<td>6.8 x 7.6</td>
<td>6.7 x 7.2</td>
<td>8.5 x 8.8</td>
</tr>
</tbody>
</table>

As previously defined, the megablocks in these craters are exposed, layered, coherent blocks, ranging from tens of metres to a few kilometres across that have been uplifted, rotated and deformed during the modification stage during crater formation. In Betio and Halba, the blocks are arranged around an hourglass shaped relatively flat central depression that sits at the lowest elevation within the crater floor (Table. 3.5). Byala crater floor pit also has uplifted blocks; however, they partially surround a “wish-bone” shaped central depression (on the north and west side). The blocks within Betio and Byala do not extend in elevation beyond the floor pit and past the average elevation of the crater floor. In contrast, in Halba some of the blocks are elevated from what is thought to
be the floor and floor pit. In all three craters, the uplifted blocks generally display dips anywhere ranging from ~5 to 90°.

In all three craters, the average block sizes range from ~100-300 m in diameter, with a few >800 m blocks situated closer to the edge of the floor pit. In general, there was no observable trend is for and increase in block size with respect to the linear radial distance from the crater centre. However, there was a trend for blocks to become more steeply dipping (~45–90°) and highly deformed, situated proximal to the centre of the central depression (Fig. 3.17). Blocks, with shallower dips (~5–15°), occur within the outer more peripheral zone (Fig. 2.9 [Chapter 2]). In Betio, there is a noticeable trend in the blocks that appear to deform with the scarp, from ~ESE-WNW or E-W (i.e., concentric orientation in the north) to ~NE-SW or NNE-SSW (i.e., radial orientation in the west and east) (Fig. 3.17A; refer to Chapter 2 for more figures); however, more frequent intra-block strike changes occur towards the centre of the structure and the SW portion. In Byala crater, the blocks are larger and more coherent around the pit; however, the data density decreases in the southern and eastern part of central pit due to absence of exposed layered bedrock, where the blocks become more disturbed. There is a noticeable trend of blocks striking radial (~NEN-SWS). Finally, in Halba Crater, there is a general radial trend of the blocks striking NE-SW in the east and ~NW-SE in the west. We should also note that Halba crater is highly deformed and possible eroded (as discussed in Section 3.5.2.), which may have some affect on measurements taken, i.e., where they are permitted due to exposure and a more complex target.
3.4.3.3 Faults, folds, fractures, and dykes

3.4.3.3.1 Faults

Within the floor pits of Betio, Byala, and Halba craters there are numerous observed, inferred, and approximated faults (as previously discussed in Chapter 2 for Betio crater). Consistently, in each crater, they are more densely distributed closer to the centre. They displace and offset layers, dykes, and fold limbs a few- to tens- of metres and are interpreted to rotate megablocks up to 90°. We interpret the large faults bounding the blocks and striking perpendicular to the bedding to be thrust faults. Within the blocks there are numerous smaller faults with observable displacements (offsetting the layer a few- to ten- of meters), which we interpret to be normal faults.
3.4.3.3.2 Folds (radial transpression ridges)

Within Betio and Byala crater’s central pits, we observe multiple large (~500 m – 1300 km) radially striking antiforms and synforms located in the periphery of the central floor pit that are relatively symmetrical and open (~120°–70° interlimb angles). There are also multiple smaller (~45 m – 500 m) folds located towards the center of the floor pit. Some of these folds appear to be asymmetrical, with tight (0°–30°) to isoclinal (0°–10°) interlimb angles and some chevron shapes. There is an apparent gradational transition in fold tightness, size, and symmetry from the outer regions to towards the blocks proximal to the flat centre due to space issues. We also observe fold limbs that occasionally appear to be detached and offset into sheet-like blocks, bounded by faults. There are very few observable intact folds in Halba. This may be due to degradation of the layers or additional deformation towards the end of uplift formation.

3.4.3.3.3 Fractures and dykes

We define a dyke as a fracture that has been infilled by impact melt and they are observed in all three central uplifts; however, these structures are much more abundant and better preserved in Betio crater. As we previously described in chapter 2, we observed ~81 dykes (ranging from ~2–51 m in width and ~22-1.2 km in length) that are typically oriented perpendicular to the bedrock. Sixty-eight percent of the dykes show clear offsets (ranging from ~2–40 m), with two or more segments, suggesting that they have been offset due to faulting during crater formation. It is important to note that, although there are dykes with only one intact segment visible, the dyke may continue and be offset by faults but may no longer be exposed due to coverage by other units. Some of the better-exposed structures contain angular light-toned clasts in a smooth low albedo matrix situated in the northern quadrant (refer to Chapter 2 for figures). A higher concentration of these structures can be found in the north and eastern sections of the central uplift. Breccia dykes are much scarcer in Byala and Halba (interpreted to be bounding the blocks), with minimal exposures. This is likely the result of either extensive erosion (Halba) or lack of exposure (Byala), as discussed further in section 3.5.2. Where there are exposures, we note that the dykes have similar measurements and observations as Betio crater; however, they tend to be more deformed within Halba.
3.5 Discussion

3.5.1 Stratigraphy

In Chapter 2, we briefly outlined a potential post-impact stratigraphic order for the units within Betio crater, based on morphological and structural mapping and using qualitative and quantitative measurements in conjunction with knowledge from craters on Earth (e.g., impact melt sheets become increasingly clast-rich towards their base [Grieve et al., 2013]) and general principles of geology, along with other remote morphologic studies conducted on central uplifts on Mars (e.g., Tornabene et al., 2010, 2012b; Marzo et al., 2010; Caudill et al., 2012; Ding et al., 2014). Based on this, we suggest that units were emplaced in the following order:

1. Layered Bedrock – (Pre-existing)
2. Polygonal Bedrock – exclusive to Byala (Possibly pre-existing)
3. Fault Breccias – (Dykes)
4. Crater Fill in the crater floor – Breccia Lens (Excavation and Modification)
5. Impact Melt Clast-rich – (Excavation and Modification)
6. Impact Melt Clast-free-poor – (Excavation and Modification)
7. Pitted Material – (Excavation and Modification)
8. Highly Altered and Eroded pitted Material – exclusive to Byala – (Excavation and Modification; eroded Post-impact)
9. Regolith/Talus Deposits – (Post-impact)
10. Aeolian Deposits – (Post-impact)
Figure 3.18: Post-impact stratigraphic column in the order of emplacement for each observed and mapped morphological unit. Please note the question marks on the eroded pitted material unit and the polygonal unit (both found within Betio crater), as their stratigraphy is difficult to determine, due to lack of information regarding their formation and composition.

3.5.2 Erosion, degradation, and preservation

The Betio, Byala, and Halba central floor pits have some consistent morphologic and structural attributes (e.g., uplifted faulted, folded, and rotated LB blocks, impact melt deposits, and post-impact erosional deposits); however, they are also quite morphologically diverse. This diversity may reflect the effect of the target properties on the impact process (e.g., Melosh, 1989) and the result of post-impact active geologic processes. Craters within the plains region are interpreted to be quite “well-preserved”, based on their morphologic features remaining intact (e.g., intact rim, terraces, internal features, minimal deposits, and ejecta) (Golombek & Bridges, 2000; Carr, 1992). Once craters have formed, they are exposed to conditions on the Martian surface and become vulnerable to modification from various processes including: mass wasting, erosion by overprinting impacts, wind, and burial by aeolian deposits and, in some cases, exhumation. One implication of this study is that it allows us to compare the level of erosion and preservation that has taken place within each crater after its formation. This is important, as studying central uplifts at various stages of preservation can be used as a standard to understand target material properties and the effects and rates of subsequent modifying processes on Mars.
The state of preservation of an impact crater is critical to determine in order to draw comparisons. We interpret Byala crater to be the best preserved, because it is dominated by what is interpreted to be impact melt deposits covering ~76% of the total central floor pit area (Fig. 3.15). Pitted material (interpreted to be the topmost portion of the crater-fill deposits – see Tornabene et al. [2012b], comprising ~64% of the total central floor pit area), is the dominant melt unit that surrounds the uplifted blocks, as well as persists within the central depression in what appears to be highly eroded and altered outcrops. In addition to this, bedrock exposure is quite low (~13% of the total floor pit area), compared to both Betio and Halba crater’s. Betio crater has less pitted materials (~4.5% of its total floor pit area) and is dominated by clast-poor impact melt (Fig. 3.14), interpreted to be the next second highest crater-fill deposit (Fig. 3.18). There is some increased exposure of layered bedrock especially in the eastern sector. We propose that this is likely a result of wind erosion, as confirmed via wind streaks in THEMIS Day IR. In Halba crater, there is no clear indications of pitted material and very little clast-poor impact melt, which we interpreted to be the topmost units of the crater fill. In contrast, there is more exposure of clast-rich material, interpreted to be crater-fill deposits surrounding the uplift, exposure of potential fault breccias between the blocks, along with densely fractured layered bedrock (possibly suggesting a deeper level of exposure), and talus coming down all the uplifted blocks. As noted above, based on terrestrial observations, the base of impact melt sheets are typically clast-rich (Grieve et al., 1977). We suspect that erosion played a larger role in the exposure and, therefore, state of these three craters. Of interest is that the floor pit of Byala craters – which we propose is the best-preserved crater – is much shallower (~90 m) compared to both Betio and Halba, whose floor pits reside at maximum depths of ~500 m below the crater floor (it is important to note that this may also be a result of pit formation). This suggests that Halba crater may be what Betio crater would look like after prolonged exposure and erosion due to their similar structural (hour-glass shaped floors) and morphological aspect of there floor pits.

Finally, we measured the extent of the layered bedrock exposures from the inner portion of the pit outwards towards the crater rim. We found that Halba crater has the most extensive exposures (i.e., widest) of LB megablocks, with an average distance of ~3 km,
and ~2.4 km and ~2.1 km for Betio and Byala, respectively. This is also consistent with an increase in the extent of erosion from Halba to Betio to Byala and also that the extent of exposure of the underlying bedrock of the uplift will become wider with respect to the crater diameter, as seen in terrestrial examples, such as Upheavel Dome (Kenkmann, 2005). We suggest that variations in target lithology may have played a role (e.g., proximity to a region of Noachian rocks and Byala occurring within a unit that is interpreted as older Hesperian-aged materials) in the preservation of these units; thus, affecting the amount of exposure within each crater. Therefore, differences in morphologies may possibly indicate a subsurface variation in the geology; however, it is important to not rule out any post-impact effects on the exposed units.

3.5.3 Regional survey of central pits vs. central peaks

To strengthen the argument that target effects play an essential role in the formation of floor pits, we conducted a regional survey of ~395 craters located south of Valles Marineris in a ~1400 x 2500 km area (~3,500,000 km²) (Fig. 3.19). Classification of the central structures involved a combination of HiRISE, CTX, and THEMIS Day IR images, in conjunction with HRSC and HiRISE DTMs, when available. Central structures were classified as either floor pits (red), summit-pits (green) or central peaks (blue). Most craters (~145) had no clearly exposed uplift, due to crater infilling and degradation (these were not included on the map). We found that within the 111 craters (that had observable central features) located in the Hr geological unit defined by Scott and Tanaka (1986) that ~78% (86 craters) were identified to have floor pit morphologies. Summit-pits were 16% (18 craters) and central peaks were 6% (7 craters). The majority of the central peak craters (excluding two) are located very close to boarders of other geologic units, particularly Noachian terrains.
Figure 3.19: Regional analysis of central uplift morphologies near Betio, Byala, and Halba craters. Red squares represent floor pit morphologies, green squares represent summit-pit morphologies and blue squares are central peaks. Within the Hr geologic unit, ~94% of the craters that have an exposed central structure have a pit morphology.

3.5.4 Mechanics of floor pit formation

The formation of a central pit is still controversial. Some authors assume that central pit craters are the result of uplift of the transient crater followed by collapse of a central peak during the modification stage (Passey & Shoemaker, 1982; Hodges, 1978; Greeley et al., 1982; Whitehead et al., 2010), while other studies suggest that late-stage melt drainage (Croft, 1981; Bray, 2009; Elder et al., 2010, 2012; Senft & Stewart, 2009, 2011), vaporization of ice in the target during the modification stage (Wood et al., 1978; similar models by Kieffer, 1977; Smith & Hartnell, 1977; Senft & Stewart, 2011; Alzate & Barlow, 2011) or even that pit formation is largely post-impact phenomenon, due to differential erosion of uplifted materials. These various models are summarized in Table 3.6. Below, we evaluate these various models based on our study of Betio, Byala, and Halba impact craters in combination with observations made in previous studies.
Table 3.6: Summary of floor pit formation models.

<table>
<thead>
<tr>
<th>Description</th>
<th>Assumptions</th>
<th>Evidence Supporting</th>
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</table>
| Ice Vaporization | Central pits are the result of subsurface volatile degassing | 1. Numerical modeling proving that the region underlying the center of the transient cavity experiences temperatures high enough for volatile vaporization of impacts into pure ice (Stewart & Selnf, 2008) and mixed ice-soil (20% ice, 80% basalt) (Pierazzo et al., 2005) targets 30 km in diameter, suggesting collapse occurs either during or shortly after crater formation.  
2. Lack of updomed floors (seen in more ice-rich targets such as Ganymede) in Martian floor pit craters (Kagy & Barlow, 2008) supports suggestions of Martian regolith ice content values of ~20% which is the value used in the simulations conducted by Pierazzo et al. (2005).  
3. Most Martian central pit craters occur in the 7- to 30-km-diameter range (Barlow, 2010; 2012) and by utilizing estimated structural uplift equations (Pilkington & Grieve, 1986), these craters are excavating to depths between 0.7 and 2.5 km, within the region where subsurface ice is expected to exist on Mars (Barlow, 2010). |
| Layered Target/Collapse | Central pits are the result of pre-existing weaknesses in the target layering (i.e., a stronger layer overlaying a weaker one) thus initiating collapse | 1. Are the thermal conditions required to produce the central pits are reached for craters as small as 5km diameter (simulations run, thus far, are typically for larger craters 30-km-diameter craters modeled).  
2. Modeling so that the vapour does not escape early in crater formation.  
3. Pit craters also identified in terrestrial, lunar, and Mercurian craters (Storm et al., 1992) bodies are unlikely to possess any significant crustal volatile layers (Whitehead et al., 2010). |
| Description | Assumptions                                      | Evidence Supporting                                                                 |
| Layered Target/Collapse | Layering is present in the target at the time of impact | 1. Greeley et al. (1982) performed a series of gas-gun experiments into differently layered targets (e.g., water, clay, sand and ice in a variety of different layering combinations), which produced some crater forms deemed analogous to central pit craters supporting that layering within a target has a direct effect on crater morphology at laboratory scales.  
2. Global studies completed by Barlow (2010) and Whitehead et al. (2010) noted a concentration of floor pit craters in units mapped as volcanics rather than sedimentary targets, supported by Caudill et al. (2012) study showing ~75% of the LB craters reported possess central pit morphologies. |
| Evidence Against | Larger scale numerical modeling of impact into layered rock, ice, and water targets has not yielded any large final craters that have pitted centers (e.g., Bray, 2009; Selnf & Stewart, 2011). | 1. |


Table 3.6 (continued): Summary of floor pit formation models.

<table>
<thead>
<tr>
<th>Melt-drainage</th>
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<tr>
<td>Description</td>
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<td>Assumptions</td>
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</table>
| Evidence Supporting    | 1. The highest density of impact-induced fractures occur at crater centers (e.g., Kenkmann, 2002).  
2. Largest pockets of impact melt are concentrated in the central regions (e.g., Grieve et al., 1977), leading to a maximum potential for melt drainage at the center of craters (Bray et al., 2009).  
3. Computer simulations by Bray (2009) and Senft and Stewart (2011) have both produced modeled craters with melt pools of widths approximately equal to the diameter of central pits on Ganymede (needs to be done for Mars). |
| Evidence Against       | 1. Neither model above, notes a central depression produced purely from central uplift collapse.  
2. Thermal models suggest that a subsurface liquid water layer on Mars seems unlikely given that Mars should have undergone cooling with time and, thus, any liquid subsurface reservoirs would be expected to have migrated to greater depths (Barlow, 2010).  
3. Most form pits in craters in ice or ice-rock mixes, but not those in volatile-poor targets (Elder et al., 2012). |

<table>
<thead>
<tr>
<th>Differential Erosion of Uplifted Materials</th>
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<tr>
<td>Description</td>
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<td>Assumptions</td>
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</table>
| Evidence Supporting                       | 1. Central structures within craters observed on Earth, e.g., Gosses Bluff, Australia (Milton et al., 1972; Hodges, 1978).  
2. Wulf et al. (2012) demonstrate that the uplifted strata in the surrounding of the central pit in Arima crater are not affected in strike or dip by the crater pit nor is a concentric downward faulting or downward flexure of strata observable. This study confirms with Wulf et al (2012) that crater pit formation most-likely postdates crater modification and formation of the central uplift. |
| Evidence Against                          | 1. Previous studies of Martian pit crater ranging in diameter from 5 to 157 km and Robbins and Hynek (2012) reported that 92% of central pit craters are qualitatively fresh, consistent with the interpretation that they are mostly relatively young and not likely the result of central peak erosion. |

3.5.4.1 Ice vaporization model

As previously stated, ice vaporization model predicts that central pits are the result of subsurface volatile degassing (Wood et al., 1978). Large volumes of ground ice are not predicted to occur at the latitudes of the craters studied here, even during higher obliquities. Evidence supporting the presence of limited amounts of volatiles in the target at the time of impact of Betio, Byala and Halba craters include includes carter-related pitted materials (e.g., Tornabene et al., 2012) found on the crater floor of Byala and Betio, and layered ejecta blanket morphologies around Betio and Byala (e.g., Barlow & Brady, 1990). The question is whether there would have been sufficient volatiles to form
a massive pit in the centre of the crater. Also as observed and shown in Figure 3.19, the preferential development of pits in lava-dominated targets, appears to be in contradiction to their proposed origin by explosive devolatization (since lavas are likely to contain less volatiles than sedimentary and aqueously altered Noachian targets) and may, instead, support the models in which target layering is paramount, as discussed below.

The other question brought up by Barlow (2010) is why craters of similar size and age but without central pits exist in the same regions as central peak craters. However, we do not know how the underlying surface behaves, therefore, we suggest that craters existing in similar targets and regions but containing different central uplift morphologies is due to inhomogenites in the subsurface areas. We do not know how the underlying Noachian surface is behaving underneath the Hr plains (i.e., judging from other Noachian terrains, it is no doubt heavily cratered and very irregular in topography [possibly including buried large basins]). An additional explanation in support of this ice vaporization model, given by Elder et al. (2010), is that possible reasons for irregularities in subsurface volatile distribution stem from the emplacement of the volcanic plains unit. The lava may have been added a ~1–2 km thick additional cover over pre-existing volatiles, or, in other cases, lava may have destroyed ground ice deposits.

Other contradictions to this model include the observation of no visible ejecta surrounding the pits, despite the work presented by Williams et al. (2013) explaining that the high thermal inertia associated with these pits may be ejecta. However, once inspected in HiRISE and CTX data, there is clearly a correlation with bedrock exposures and a high thermal inertia, not radial ejecta deposits emanating from the central pits. It is important to note that this ejecta could have been eroded away, particularly in Betio and Halba, as discussed below.

### 3.5.4.2 Melt-drainage model

The melt-drainage model predicts that central pits are the result of impact-melt drainage, through fractures on the crater floor resulting in a hot plug (Croft, 1981). This model is quite difficult to prove or disprove based on the fact that we are unable to see the subsurface in the three craters. Furthermore, the pit floor of each crater is covered in
aeolian deposits; as well as, what is interpreted to be a impact melt deposits, therefore, no visible fractures are observed there.

3.5.4.3 Differential erosion of uplifted materials

This model for floor pit formation suggests that they may be a consequence of differential erosion of weaker materials in the core of an uplift, exposing a pit with blocks within it. As previously mentioned in Chapter 1 and Table 3.6, crater central structures observed on Earth for example Gosses Bluff, Australia (Milton et al., 1972; Hodges, 1978) have been through some degree of erosion and sometimes preferential erosion can occur in the centre of the central uplift. Morphologic evidence found in the craters, in this study, that would support that some degree of erosion has occurred include: exposure of what is interpreted to be crater-fill deposits (e.g., surrounding the uplifted blocks in Halba crater), exposures of potential clast-rich fault breccias in Betio and Halba, degraded and altered pitted materials in Byala, along with the lack of observable ejecta from a overprinting craters within each floor pit; in particular the, ~4.5 km simple crater to the east in Betio crater. Topographic evidence supporting this comes from the observation and measurement of deeper floor pits in both Betio and Halba, which are interpreted to be less preserved than Byala crater, as well as deeper rim to crater floor ratios in Betio and Halba than in Byala, indicating a few hundred metres of erosion has taken place since the time of impact based on equation from Stewart and Valent (2006).

However, the observation of pitted materials in Betio and Byala (thought to represent the top of the crater-fill) suggests that not as much erosion has taken place as initially suspected. For comparison, in Corinto crater, a Martian rayed crater (i.e., a nearly pristine crater), posses floor pit morphology filled with crater-related pitted materials (Tornabene et al., 2012b). Thus, we suggest that erosion is not a serious contributing factor with respect to the floor pit form.
Figure 3.20: Left image: Gosses Bluff, Australia (~22 km in diameter) (image credit: http://aboriginalastronomy.blogspot.ca/2011/03/impact-craters-in-aboriginal-dreamings_28.html). The central part of the uplift has been eroded away. Right image: The floor pit within Halba crater seen in ArcScene, CTX image (G09_021578_1541_XN_25S066W) draped over the HRSC DTM (h2486_0001_da4), with a 1.5x VE. The central depression within this Martian crater has a disputed formation mechanism, as discussed in Section 3.5.4.

3.5.4.4 Layered target model

The layered target model predicts that central pits are the result of pre-existing weaknesses in the target layering for units (e.g., water, clay, sand and ice in a variety of different layering combinations – i.e., a stronger layer overlaying a weaker one); thus, resulting in the collapse of the “weaker” central section (Greeley et al., 1982; Passey & Shoemaker, 1982). Although the original model does not account for the fine-scale layering of a stronger thinner volcanic layer overlying a weaker, thicker, sediment/ash layer, we suggest that this would be an ideal case for pre-existing weaknesses (i.e., stronger cyclically layered flood-basalts, with interfaces of weaker sediment/ash layers) based on previous studies of there target stratigraphy. These target surfaces may also contain lateral heterogeneities, such as topography, pre-existing faults, fractures, and joints or tilted stratigraphy (seen in Halba). We suggest that this model should be referred to as differing layer strength model, as to not confuse its original meaning.
As noted in chapter 1 and Table 3.6, previous studies (e.g., Barlow, 2010; Whitehead et al., 2010; Caudill et al., 2012) have noted a concentration of floor pit craters in units mapped as volcanics. This was confirmed in our regional study showing ~94% of the craters in the Thaumasia Planum Hesperian Ridged Plains unit contain central pit morphologies (summit or floor). Central peaks are dominantly found in regions of either Noachian geologic units or on the boarder of a Noachian unit and Hr unit. Thus, implying that regional variations in a wide range of crater morphologies suggest that Martian crater forms reflect subsurface heterogeneities (Barlow & Bradley 1990).

We suggest that such a layered target may provide a more easily deformed target material that possibly generates a more extensively brecciated uplift core that then collapses. In the craters involved in this study (Betio, Byala, Halba), this is observed through increased deformation in outcrops more proximal to the central depressions and lack of uplifted blocks in the pit centres. However, it should be noted in the centre is where minimal exposure of layers is observed, due to melt/aeolian coverage and prevents us from observing the amount of deformation taking place. Fortunately, we can assume that the highest amount of deformation has occurred in the centre due to observations made at terrestrial craters (e.g., Kenkmann et al 2005; Osinski & Spray, 2005). If this interpretation is correct, a peak may have formed early on during the modification stage, but later collapsed during the final stages of crater formation (Grieve & Therriault, 2004). This collapse may have been aided by relatively lower bulk strength of the collapsing material, due to the presence of layering in the target rocks. However it is important to note that other impact characteristics (e.g., projectile density, impact velocity) may also be involved (Grieve & Therriault, 2004). This work also suggests that the rims surrounding central pits might be formed due to the basal collapse and spreading of a central uplift. We suggest that further work be done on modeling impacts into layered targets, with cyclic layering that alternate in strength, to see if pits occur.

3.6 Summary and Conclusions

To summarize, the floor pit formation models described above, all contain plausible formation explanations; although, some are more likely than others based on our morphological and structural maps of Betio, Byala and Halba craters, along with our
regional study of central pit craters versus central peak craters in Hesperian Ridged Plains. To date, the preferred models used to explain central pit formation are explosion and vaporization of target volatiles and melt-drainage (e.g., Wood et al., 1978; Passey & Shoemaker, 1982; Barlow, 2010; Senft & Stewart, 2011; Elder et al., 2012); however, these models are difficult to prove, especially in the scope of our study (i.e., lack of visual evidence of the subsurface stratigraphy). Although the temptation exists to call on a single mechanism to dominate crater collapse, it is possible that a combination of models form these central floor pits. We suggest that observations made in Betio, Byala, and Halba crater, along with our regional study of central structure morphology, provide significant evidence that if a target material has varying strengths (e.g., layering), the central peak cannot be supported (especially in the centre, where the most amount of deformation typically occurs– as noted in terrestrial observations [e.g., Kenkmann et al., 2005] and Martian observations [e.g., Wulf et al., 2012]) and, therefore it will collapse, perhaps forming a central pit in its place, as described in Section 3.5.4.. This argument is strengthened by observations noted on the Moon by Hawke and Head (1977), who observed differences in central uplift characteristics – greater abundance of central peaks in mare versus coherent nature of the mare substrate, in contrast to mega-regolith of the highlands, as well as observations made in terrestrial craters, such as the lack of an emergent central uplift at Haughton (layered sedimentary target) versus with a well-formed central peak at Boltysh (coherent crystalline target) (e.g., Mastasis, 1999; Grieve & Therriault, 2004).

In this study, we have also clearly illustrated that although there are many similar morphologic units found within the floor pits, the extent and state of preservation varies from crater to crater. We attribute variations in crater impactite morphologies and preservation to be the result of subsurface variations in the underlying stratigraphy. It is clear that each crater has experienced various levels of erosion and we have determined that based on the observations made is Section 3.5.4 that the order from best to least preserved crater is Byala, Betio, Halda. These results provide further information regarding target material properties, and the effects and rates of subsequent modifying processes on Mars (e.g., Craddock et al., 1997; Bleacher et al., 2003; Forsberg-Taylor et al., 2004; Grant et al., 2008). Next steps for this study would include spectral analysis of
the various morphologies within each crater to provide more evidence of their similar classifications, as well as analyze more craters in this area. Information on deeper lithologies acquired through future missions may help resolve the true effect of subsurface competence on intracrater structure.
3.7 References


Chapter 4

4 Final discussions and conclusions

It has been well established, throughout the literature, that the formation of central uplifts is poorly understood and the formation of floor pit morphologies even less so. The main goal of this thesis is to provide some additional observations, interpretations and constraints on current models by utilizing high-resolution images and digital terrain models of relatively well-preserved central uplifts on Mars. The central uplifts selected for this thesis are within Betio, Byala, and Halba craters – all of which are ~30 km in diameter, located with Hesperian Ridged Planes geologic unit (Scott & Tanaka, 1986) in the Thaumasia Planum region this unit is interpreted to be composed of stronger cyclic layers comprised of alternating basaltic lavas and weaker sediment or ash layers, based on the geologic context, morphology, morphometry and spectroscopy of the layered bedrock exposed in central uplifts across this particular region of Mars (Tornabene et al., 2010, 2012; Caudill et al. 2012; Quantin et al., 2012). The exposure of this bedrock can be found within the uplift megablocks within the central uplifts with alternating lighter-toned and thicker fractured basalts and thinner, darker-tone sediments and ash layers. These layers are key to this study, as they provide a frame of reference for structural deformation, which can be readily measured and interpreted.

Chapter 2 focused on what is interpreted as the best-exposed floor pit within Betio crater, morphologically and structurally mapping the uplift. This map allowed measurements and observations to be made; thus, leading to interpretations regarding not well-constrained issues, such as weakening mechanisms within uplifts. Many observations have been made within craters on Earth but these are quite eroded, so a comparison to a better-preserved structure is essential. Many structural similarities were noted between Betio craters’ floor pit and terrestrial observations, including: increased brittle deformations of faults (i.e., higher concentration), folds (i.e., become overturned and tighter), and, blocks (i.e., higher dip angles and rotated up to 90°) proximal to the centre suggesting increased deformation. An in-depth analysis of dykes was provided and, although dykes have been observed previously in Martian central uplifts (e.g., Tornabene et al., 2012), this was one of the first studies to look at dykes in detail making structural
measurements and in depth observations and analysis. These observations helped to support central uplift formation models, such as block oscillation model (Melosh & Ivanov, 1999). Observations and interpretations of the pre-, syn-, and post-impact morphologic units, including, impact melt and/or breccias, exposed bedrock, and erosional units was also provided. These concepts were then used in Chapter 3, in a comparison study.

Chapter 3 is a continuation of Chapter 2, as it focuses on morphologic and structural mapping and measurements of two new craters (Byala and Halba) and we, then, compare these results to that of Betio crater discussed in Chapter 2. However, this chapter has a stronger focus on the formation of a central floor pit and the comparison of extent and preservation of the morphologic units within the craters. We compared three ~30 km, layered bedrock craters with floor pit central uplift morphometries to assess the morphologic and structural similarities and differences

The major results form this chapter suggest that a layered target of weaker and stronger units may have a role in the formation of the floor pit. This was suggested based on the observations of the three craters, a regional study of central uplift morphologies and observations made on Earth and the Moon. Other results involved the categorizing of each of the crater from the best preserved to the least preserved, Byala, Betio, Halba, respectively.

4.1 Challenges and limitations of a remote based study on central uplifts

When conducting a photogeologic study utilizing remotely obtained images of another planets surface features, there are multiple challenges and limitations that one needs to accept and recognize.

• Many assumptions need to be made, i.e., the composition, strength, and depth of the subsurface geology

• Unable to ground truth (make measurements, samples, etc.), our interpretations
• Data acquisition limitations: lack of high-resolution DTMs and/or full HiRISE coverage of the central uplifts and limited to an up to 25 cm/pixel resolution. The closest thing we have are using terrestrial analogs to make comparisons to our observations

4.2 Implications

The results from this study provide a new understanding on the distribution and classification and various morphologic units interpreted to be melt deposits, post-impact erosional deposits, and highly deformed layered bedrock. The improved spatial resolution allow for a more detailed look at:

• Weakening mechanisms (structure)
• Level of preservation within each crater
• Stratigraphy of the morphologic units
• Differences in morphologies, which may possibly indicate a subsurface variation in the geology. However, it is important to not rule out any post-impact effects on the exposed units
• Similarities to terrestrial observations of central uplifts
• Further analysis of floor pit models

4.3 Future work

• The inclusion of spectral analysis utilizing CRISM data of each crater and morphologic units to determining the compositional characteristics of the deposits, providing further support for their interpretations

• Applying these methods and mapping on three craters of similar target features but with different diameters (e.g., one small ~15 km diameter, medium ~30 km diameter, large 100 km diameter)
4.4 References


Appendices

Appendix A: Context images for subsets in Chapters 2 and 3 for Betio, Byala, and Halba craters.

Figure A.0.1: Betio Crater floor pit with subset images from Chapter 2 marked and labeled. CTX image B20_017583_1571_XN_22S078W(left) noting the extent of image right. HiRISE images ESP_017583_1565 and ESP_027987_1565.
Figure A.0.2: Betio Crater floor pit with subset images from Chapter 3 marked and labeled. CTX image B20_017583_1571_XN_22S078W (left) noting the extent of image right. HiRISE images ESP_017583_1565 and ESP_027987_1565.
Figure A.0.3: Byala crater floor pit with subset images from Chapter 3 marked and labeled. CTX image G06_020734_1540_XN_26S066W (left) noting the extent of image right. HiRISE image ESP_021578_1540_RED.
Figure A.0.4: Halba crater floor pit with subset images from Chapter 3 marked and labeled. CTX image B19_017002_1538_XN_26S056W (left) noting the extent of image right. HiRISE image ESP_031230_1535(left), ESP_017147_1525(right).
Appendix B: High-resolution geomorphologic map of Betio crater.
Curriculum Vitae

Name: Anna Nuhn

Post-secondary Education and Degrees:
University of Western Ontario
London, Ontario, Canada
2008-2012 B.Sc.

Honours and Awards:
NSERC CREATE
2012-2014

Related Work Experience
CPSX Outreach Provider
The University of Western Ontario
2012-2014

Teaching Assistant
The University of Western Ontario
2012-2013

Publications: