Investigating the Complexity of Impact Crater Ejecta

Michael Raymond Zanetti
Washington University in St. Louis

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Investigating the Complexity of Impact Crater Ejecta

by

Michael Raymond Zanetti

A dissertation presented to the
Graduate School of Arts & Sciences
of Washington University in
partial fulfillment of the
requirements for the degree
of Doctor of Philosophy

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# Table of Contents

List of Figures ........................................................................................................................................................................ vi

List of Tables ............................................................................................................................................................................... ix

Acknowledgements ......................................................................................................................................................................... x

Abstract of the Dissertation .......................................................................................................................................................... xiv

## Chapter 1: Introduction to the Dissertation ................................................................. 1

1.1 Impact Crater Ejecta Emplacement ........................................................................ 3

1.2 Goals and Objectives ............................................................................................. 7

1.3 Major Results ........................................................................................................ 8

1.3.1 Chapter 2 ........................................................................................................... 8

1.3.2 Chapter 3 .......................................................................................................... 10

1.3.3 Chapter 4 .......................................................................................................... 12

1.4 Statement of Labor ............................................................................................... 15

1.5 Copyright and Permissions ................................................................................... 16

Chapter 1 References .................................................................................................. 17

Chapter 1 Figures ......................................................................................................... 18

## Chapter 2: The Geology and Geomorphology of Aristarchus Crater .......................... 21

Abstract ....................................................................................................................................................................................... 21

2.1 Introduction ............................................................................................................ 22

2.1.1 Geologic Background of the Aristarchus Region ............................................. 22

2.1.2 General Geology of the Crater ejecta ............................................................... 27

2.1.3 Age Estimates .................................................................................................. 30

2.1.4 Objectives ....................................................................................................... 31

2.2 Data and Methods ................................................................................................. 31

2.2.1 Datasets .......................................................................................................... 31

2.2.2 Methods........................................................................................................... 35

2.3 Results: .................................................................................................................... 37

2.3.1 Morphometry ................................................................................................. 38

2.3.2 Geomorphologic Map ...................................................................................... 40

2.3.3 Geologic Map from Image Classification ....................................................... 57

2.4 Discussion .............................................................................................................. 59
4.4 Discussion ........................................................................................................................................ 187
  4.4.1 Petrogenesis of Decomposed Zircons ...................................................................................... 187
  4.4.2 Provenance of the Impact Melt Glass ...................................................................................... 190
  4.4.3 Rheology of Impact Melt Glass .............................................................................................. 191
  4.4.4 Material exchange within Decomposition Rim ........................................................................ 192
  4.4.5 Cathodoluminescence Dependence ....................................................................................... 193
4.5 Conclusions .................................................................................................................................... 195
Acknowledgements .............................................................................................................................. 197
Chapter 4 References: ........................................................................................................................ 198
Chapter 4 Figures: ............................................................................................................................... 201
Chapter 4 Tables: ............................................................................................................................... 217
List of Figures:
FIGURE 1-1: EXCAVATION FLOW AND ZONE OF EXCAVATION 18
FIGURE 1-1: BALLISTIC SEDIMENTATION 19
FIGURE 1-1: CENTRAL PEAK FORMATION SCHEMATIC 20
FIGURE 2-1: ARISTARCHUS PLATEAU CONTEXT 97
FIGURE 2-2: ARISTARCHUS CRATER CONTEXT 98
FIGURE 2-3: WAC-COLOR IMAGE 99
FIGURE 2-4: MORPHOLOGY AND ALBEDO NAC CONTROLLED MOSAICS 100
FIGURE 2-5: TOPOGRAPHIC PROFILES ACROSS ARISTARCHUS CRATER 101
FIGURE 2-6: PREVIOUS GEOMORPHOLOGIC MAPS 102
FIGURE 2-7: M3 AND CLEMENTINE UVVIS SPECTRAL PARAMETER MAPS 103
FIGURE 2-8: WAC COLOR RATIO IMAGES 104
FIGURE 2-9: WAC COLOR PRINCIPAL COMPONENT ANALYSIS (PCA) IMAGE 105
FIGURE 2-10: CRATER IN WAC COLOR RATIO AND PCA IMAGES 106
FIGURE 2-11: GRAIL BOUGUER GRAVITY MAP 107
FIGURE 2-12: ARISTARCHUS CRATER GEOMORPHOLOGIC MAP 108
FIGURE 2-13: NAC CONTROLLED-MOSAIC MAP CONTEXT IMAGE 109
FIGURE 2-14: SMOOTH FLOOR AND HUMMOCKY FLOOR 110
FIGURE 2-15: ARISTARCHUS CENTRAL PEAK 111
FIGURE 2-16: CHANNELED AND VENEERED WALL 112
FIGURE 2-17: CONCENTRIC RIDGED EJECTA 113
FIGURE 2-18: MELT-RICH CHANNELS AND LOBES 114
FIGURE 2-19: COLLAPSED TERRACE MELT POND 115
FIGURE 2-20: GLOBULAR MELT TEXTURE 116
<table>
<thead>
<tr>
<th>FIGURE 2-21: LARGEST MELT FLOW</th>
<th>117</th>
</tr>
</thead>
<tbody>
<tr>
<td>FIGURE 2-22: MELT FLOW CROSSING EJECTA RAY.</td>
<td>118</td>
</tr>
<tr>
<td>FIGURE 2-23: SPLATTER FLOWS</td>
<td>119</td>
</tr>
<tr>
<td>FIGURE 2-24: MELT VENEER AND SMALL IRREGULAR CRATERS</td>
<td>120</td>
</tr>
<tr>
<td>FIGURE 2-25: STRATIFIED BLOCK FIELD</td>
<td>121</td>
</tr>
<tr>
<td>FIGURE 2-26: LARGEST STRATIFIED BLOCK</td>
<td>122</td>
</tr>
<tr>
<td>FIGURE 2-27: STRATIFIED BLOCK LOCATIONS ON NORTHEAST WALL</td>
<td>123</td>
</tr>
<tr>
<td>FIGURE 2-28: IMPACT MELT AND BOULDER DISTRIBUTION</td>
<td>124</td>
</tr>
<tr>
<td>FIGURE 2-29: BONITO LAVA FLOW LAYERING</td>
<td>125</td>
</tr>
<tr>
<td>FIGURE 2-30: ARISTARCHUS CRATER GEOLOGIC MAP</td>
<td>126</td>
</tr>
<tr>
<td>FIGURE 2-31: SCHEMATIC CLASSIFICATION OF ROCK TYPES</td>
<td>127</td>
</tr>
<tr>
<td>FIGURE 2-32: REGIONAL IRON CONTENT AND DIVINER CF POSITION</td>
<td>128</td>
</tr>
<tr>
<td>FIGURE 2-33: IRON AND CF POSITIONS AT ARISTARCHUS CRATER</td>
<td>129</td>
</tr>
<tr>
<td>FIGURE 2-34: LUNAR PROSPECTOR GRS – THORIUM ABUNDANCE</td>
<td>130</td>
</tr>
<tr>
<td>FIGURE 2-35: LABORATORY CF POSITIONS FROM GLOTCH ET AL., 2010</td>
<td>131</td>
</tr>
<tr>
<td>FIGURE 2-36: CF POSITION VS IRON CONTENT AND DISTRIBUTION</td>
<td>132</td>
</tr>
<tr>
<td>FIGURE 2-37: IRON CONTENT VS THORIUM ABUNDANCE DISTRIBUTION</td>
<td>133</td>
</tr>
<tr>
<td>FIGURE 2-38: IRON CONTENT VS THORIUM ABUNDANCE - SAMPLES VS MAP</td>
<td>134</td>
</tr>
<tr>
<td>FIGURE 3-1: CRATER POPULATION DENSITY MAPS – TYCHO, ARISTARCHUS</td>
<td>167</td>
</tr>
<tr>
<td>FIGURE 3-2: CSFD - WHOLE-AREA AND MELT PONDS</td>
<td>168</td>
</tr>
<tr>
<td>FIGURE 3-3: CSFD - DENSITY SUBDIVISION COUNT AREAS</td>
<td>169</td>
</tr>
<tr>
<td>FIGURE 3-4: COUNT AREAS ON IMPACT MELT AND EJECTA AT TYCHO</td>
<td>170</td>
</tr>
<tr>
<td>FIGURE 3-5: GHOST CRATERS IN IMPACT MELT AT TYCHO</td>
<td>171</td>
</tr>
</tbody>
</table>
FIGURE 4-1: MISTASTIN GEOLOGIC MAP 201
FIGURE 4-2: MISTASTIN IMPACT MELT GLASS 202
FIGURE 4-3: CROSS-POLARIZED PHOTOMICROGRAPHS 203
FIGURE 4-4: BACKSCATTERED ELECTRON IMAGES 204
FIGURE 4-5: DETAIL OF THE DECOMPOSITION RIM AND CORE-RIM INTERFACE. 205
FIGURE 4-6: ZRO₂ – SIO₂ PHASE DIAGRAM 206
FIGURE 4-7: LRS RAMAN SPECTRA – ZIRCON, BADDELEYITE, TETRAG-ZRO₂ 207
FIGURE 4-8: LRS RAMAN PHASE MAPS OF MZRN-1 208
FIGURE 4-9: ALUMINUM AND SODIUM X-RAY INTENSITY MAPS 209
FIGURE 4-10: CONCORDIA PLOTS OF SIMS SPOT ANALYSES 210
FIGURE 4-11: MZRN-2 PMT-CL IMAGE 211
FIGURE 4-12: TRACE ELEMENT SPOT ANALYSIS TRAVERSE OF MZRN-2 212
FIGURE 4-13: HYPERSPECTRAL CL SPOT ANALYSES 213
FIGURE 4-14: FLOW BANDING IN GLASS AND ZR HALO 214
FIGURE 4-15: SOUTH-TO-NORTH ZR CONCENTRATION TRAVERSE 215
FIGURE 4-16: VISCOSITY VERSUS TEMPERATURE 216
## List of Tables:

<table>
<thead>
<tr>
<th>Table</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>TABLE 2-1</td>
<td>ARISTARCHUS CRATER MORPHOMETRY</td>
<td>135</td>
</tr>
<tr>
<td>TABLE 2-2</td>
<td>PRINCIPAL COMPONENT ANALYSIS - WAC COLOR</td>
<td>136</td>
</tr>
<tr>
<td>TABLE 2-3</td>
<td>GEOLOGIC MAP LEGEND</td>
<td>137</td>
</tr>
<tr>
<td>TABLE 3-1</td>
<td>CSFD MEASUREMENTS FOR TYCHO AND ARISTARCHUS</td>
<td>172</td>
</tr>
<tr>
<td>TABLE 4-1</td>
<td>IMPACT MELT GLASS COMPOSITION</td>
<td>217</td>
</tr>
<tr>
<td>TABLE 4-2</td>
<td>SELECTED MAJOR AND TRACE ELEMENT CONCENTRATIONS</td>
<td>218</td>
</tr>
</tbody>
</table>
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Michael Raymond Zanetti

Washington University in St. Louis

October, 2015
For the two I love most:

Nora Weitz
&
Thomas Oliver Zanetti
ABSTRACT OF THE DISSERTATION

Investigating the Complexity of Impact Crater Ejecta

by

Michael Raymond Zanetti

Doctor of Philosophy in Earth and Planetary Sciences

Washington University in St. Louis, 2015

Scott Rudolph Professor Bradley L. Jolliff, Chair

The formation of an impact crater ejecta blanket can be viewed as a form of organized chaos. Material that is ejected from a crater is heavily brecciated, but falls back to the surface along ballistic trajectories, generally preserving an inverted sense of the original stratigraphy. As the ejecta re-impacts the area surrounding the crater it forms a thick blanket of ejected material and reworked target surface that gradually thins away from the crater rim. Within the crater, crater modification processes, such as wall terrace formation and impact melt drainage, transform the crater in expectable ways.

The approach adopted in this research is to use what is known about impact cratering and ejecta emplacement processes to geologically map craters on the Moon using remote-sensing data, determine the timing of individual impacts on the Moon, and investigate terrestrial impact melt glass. Research has been divided into three parts: 1) a detailed geomorphologic and geologic map of the lunar crater Aristarchus; 2) detailed crater size-frequency distribution measurements on the ejecta blankets of the lunar craters Aristarchus and Tycho; and 3) characterization of zircon decomposition in impact melt glass from the Mistastin Lake impact structure, Labrador, Canada.
Mapping the geomorphology and geology of Aristarchus has shown that there are differences in the distribution of morphologic and compositional units related to pre-existing topography. I use the basic principles of inverted stratigraphy and remote-sensing data to investigate the geology of the subsurface material excavated by the crater and determine that Aristarchus likely excavated a buried pluton, or hypabyssal intrusive body, related to the large, possibly bi-modal, Cobra Head volcanic complex on the southern Aristarchus Plateau.

Measuring crater size-frequency distributions on the ejecta blankets of Aristarchus and Tycho were done to determine the timing of these impacts; however, my measurements revealed that there is a significant difference in crater density, irrespective of crater diameter, between impact melt and ejecta blanket units. I show that the difference in crater density between these units can most likely be explained by a mechanism of self-secondary cratering, where late-arriving fragments of ejecta crater the surface of the ejecta blanket after it forms, but prior to the arrival of impact melt flows. These measurements call into question the long-held notion that ejecta blankets represent completely resurfaced units through ballistic sedimentation, free of impact craters immediately after formation, and these measurements suggest that cratering flux over the last billion years of the Solar System may be considerably lower. Lastly, I use field observations and a number of state-of-the-art laboratory analyses of a sample of impact melt glass from the Mistastin Lake impact structure to study the decomposition of zircon grains and the provenance of the impact melt. From my measurements, I show that zircon grains from a mangerite target rock were entrained in a superheated melt of very low viscosity and quenched, preserving high temperature mineral phases, and revealing how zircon grains undergo decomposition in a natural sample.
Chapter 1: Introduction to the Dissertation

The formation of impact craters on planetary surfaces is a fundamental geologic process. Craters are observed on the surface of every Solar System body, from the smallest asteroids to the largest rocky and icy surfaces of planets and moons. The formation of impact craters is the dominant geologic process on all asteroids, most planetary moons, and the majority of terrestrial planets. Terrestrial bodies, like the Moon and Mercury, are pock-marked by innumerable craters from thousand-kilometer-diameter impact basins, to microscopic craters on pyroclastic glass beads. On Mars, heavy impact cratering in the early history of the planet is still evident, but aeolian and climate processes have over-taken cratering as the dominant geologic process. On Earth, plate-tectonics and high erosion-rates have efficiently and effectively removed most of the evidence of the geologic importance of impact cratering. As such, impact cratering is often a forgotten and overlooked process, one relegated to a page or two of introductory Earth Science textbooks. However, the significance of impact cratering in the Solar System cannot be overstated, and it is through the study of impact craters on the Moon, Mars, and other bodies, that we can learn about the history of the Solar System, planetary evolution through time, and the geology of the subsurface on other worlds.

The study of impact craters dates to around the turn of the 20th century when asteroidal impacts became increasingly credible as a geologic process. Early adopters of the suggestion that impacts could be an agent of geologic change were G. K. Gilbert, D. Barringer, and even A. Wegener. Impact structures, such as Meteor Crater in Arizona, the Kaali Craters in Estonia, and the Henbury Craters in Australia were all identified in the 1920’s and 1930’s as probable meteorite impacts, predominantly based on association with meteoritic iron fragments, the lack

1 With the notable exception of Jupiter’s moon Io, the most volcanically active body in the Solar System.
of volcanic rocks and structures, and their crater shapes. Definitive proof still lacked until the identification of shock-metamorphosed quartz grains identified in Meteor Crater, and the Ries Crater, Germany by Shoemaker and Chou (1961) in the 1960’s. Having a diagnostic indicator of incredible shock pressures resulting from an impact was the observation necessary to define an impact crater, and from then, our modern understanding of the impact cratering process began. The developments in the field of impact cratering since the 1960’s have been enormous and now at least 188 impact craters have been confirmed on Earth. Hydrocode modeling of impact craters and crater scaling laws derived from experiments and atomic bomb testing now allow us to extrapolate cratering dimensions and energies.

The past 50 years of impact crater investigations from field studies on Earth and studies of photogeologic and spectral datasets, predominantly from the Moon and Mars, have provided a great number of tools for investigating impact craters on the Moon. Using basic morphometric measurements, such as crater diameter and crater depth, it is possible to infer a wide range of geometrical relationships, such as the crater rim height, the depth of excavation, the amount of stratigraphic uplift, and the thickness of the ejecta blanket (e.g. Pike, 1977; 1988; Melosh, 1989; Grieve et al., 1998). Although these relationships only result in estimates of the given parameter, they are useful as tools to aid geologic interpretation.

In the following chapter I briefly introduce the ejecta emplacement process to provide context for the interpretations I make from geomorphologic and geologic mapping, and auto-secondary cratering on ejecta blankets. I then present the goals and objectives of the dissertation in general, and how each subsequent chapter addresses them. I conclude by briefly outlining the major results from the subsequent chapters.
1.1 Impact Crater Ejecta Emplacement

Although the material in this dissertation is primarily concerned with the distribution of materials ejected by the cratering process, rather than the physical formation of ejecta by fracturing and mobilizing target material, here I provide the basic background for the crater excavation process. I am disregarding a considerable amount of detail, and those interested can find excellent reviews on the entire cratering process in Melosh (1989), and Osinski and Pierazzo, (2012), among others.

Impact craters form when a projectile, moving at great speed, impacts a target. In order to provide general background information, here I am only considering the process of hypervelocity impacts, in which the incoming projectile is travelling at > 3 km/sec, and impacts the target with essentially the speed at which it was traveling through space. The cratering process is typically described in 3 stages, known as contact and compression, excavation, and modification (Gault et al., 1968). The early stages of the crater formation process is generally the same for small, bowl-shaped craters as it is for larger, complex craters (those with a prominent central peak), with the major differences occurring during the modification stage.

Contact and compression, which last at most a few seconds, are the beginning stages of cratering during which the projectile contacts the target, and a tremendous amount of kinetic energy is transferred to the target (and into the incoming projectile) in the form of shock waves. As the initial shock wave, transmitted directly into the target, moves hemispherically outward away from the impact site, a second wave is transmitted into the projectile and is reflected off the back of the bolide and into the target as a rarefaction wave. The combination of these two major shock waves, the initial shock wave that compresses the target material, and the rarefaction wave that “releases” the compressed material, act together to compress both the projectile and target to
high pressures (and temperatures) that are then released, resulting in melting and vaporization of the highly compressed material (Melosh, 1989; French et al., 1998).

The excavation stage immediately follows the compression stages and is the time when the opening of the transient crater begins. The passage of the shock wave imparts a radial velocity to fractured target materials which begin to move, and upward-directed pressure gradients form through the interaction of the shock wave and the target surface (Melosh, 1989). The result of the upward pressure gradient and moving particles is the excavation flow field, through which material is ejected from the crater. Excavation flow velocities are highest at the point of impact, and grade away with distance from the impact site as an approximate inverse power function (Melosh, 1989). The excavation flow follows curved streamlines (Fig. 1) that also define the excavated volume of material that is ejected. Areas close to the impact site are driven downward, and are not ejected. Areas that are very distal to the impact site do not travel fast enough to be excavated from the crater. The result is a relatively thin excavation zone that is roughly 1/10\(^{th}\) of the transient crater diameter, and only represents about 1/3\(^{rd}\) the transient crater depth (Fig. 1-1) (Melosh, 1989).

The fractured and dislocated blocks being ejected by the excavation flow are moving along independent ballistic trajectories, but remain as part of a coherent wall of debris in the shape of an inverted cone known as the ejecta curtain (Melosh, 1989). The ejecta curtain remains a more or less contiguous wall throughout the deposition process and effectively drapes down over the area surrounding the crater rim forming the ejecta blanket. The emplacement of ejecta as a curtain of material along ballistic paths has the effect of preserving the original stratigraphy of the target, only in an inverted form known as the overturned flap (Shoemaker, 1963). The concept of the overturned flap is important, as it means that one can effectively reconstruct the
relative subsurface stratigraphy of the excavated zone. An often cited example of the utility of this principle is observed at Meteor Crater, AZ, where the three major stratigraphic units (the uppermost Moenkopi sandstone, the middle Kaibab dolostone, and basement Coconino sandstone) are inverted in the ejecta blanket, with the shallowest unit (the Moenkopi) overlain by the middle unit (the Kaibab), which is in turn overlain by the basement rocks (the Coconino) (Shoemaker, 1963). As the ejection velocity also varies with depth, the stratigraphy can also be estimated as a function of distance from the crater, with shallowest units found most widespread and distal, and the deepest units closer to the rim. Although not precisely quantifiable, this relationship is useful when considering the distribution of ejecta.

The materials in the ejecta curtain moving on ballistic trajectories travel at their ejection velocity when they re-impact the surface, and as they land they impart radial momentum to the surface and entrain material in a process known as ballistic sedimentation (Fig. 2) (Oberbeck, 1975). Although the material nearest the crater rim is emplaced at low velocity, because it has only traveled a short distance, and the ejecta landing with greater velocity can have a great effect on the morphology of the ejecta blanket (Oberbeck, 1975; Melosh, 1989). Radial flow features in the distal parts of the ejecta blanket are the result of material landing and flowing away from the crater, mixing with the surface as it does so. A consequence of ballistic sedimentation is that for large craters the ejecta blanket may include a significant proportion of admixed material (Oberbeck, 1975). Additionally, the process of ejecta blanket emplacement and ballistic sedimentation should have the effect of being a completely resurfaced unit, as pre-existing small craters are mantled by ejecta and sedimentation effectively erodes the surface smooth.

The final stage in the cratering process, the modification stage, is when major structural changes take place and the crater assumes its final form. The modification process begins when
the crater has reached its maximum size and while ejecta is still lofted on ballistic trajectories. The crater walls, which also experienced some sense of motion radially away from the crater (but did not fragment and become lofted), begin to collapse back into the crater center, usually occurring along long curvilinear, concentric, listric faults.

With respect to the formation of large impact structures, the modification stage is complicated. Craters that form above a certain size threshold (~17 km diameter on the Moon) have complex morphologies consisting of numerous circumferential terraces and prominent central peaks (French, 1998). The detailed formation of these so-called complex craters is relatively uncertain, but it is generally considered to follow the development shown in the schematic diagram in Fig. 3.

The formation of the transient cavity in simple, bowl shaped crater and complex craters is essentially identical (Fig. 3a). The transient cavity is lined with impact melt, and the ejecta is lofted along ballistic trajectories. However, at the much larger energies involved in forming complex craters, the target rocks respond by elastically rebounding after the transient cavity reaches its maximum depth (Fig. 3b). An often-used analogy is the rebound seen in a pool of water after a single droplet of water hits the surface, and a large central mound forms, which then oscillates forming ripples. The target rocks of complex craters respond in a similar manner, rebounding once to form a central peak (Fig. 3c). The rebounding material that forms the central peak uplifts deeply seated material that would otherwise not be exposed through the excavation process. The amount of stratigraphic uplift that occurs and the depth of material exposed in central peaks is estimated to follow a power law \( SU = 0.086D^{1.03} \); \( D \) = final crater diameter) (Grieve and Pilkington, 1996). At Aristarchus Crater this equates to a sampling depth of ~4 km.
A final note about the morphology of complex craters worth mentioning is the relationship between the ejecta blanket and the final crater diameter. In simple craters the rim crest is effectively the edge of the transient crater, and ejected material piles up around the rim as the ejecta blanket. However, in complex craters, owing to the development of wide terraces that slump back into the crater, material in the low velocity zone that becomes the overturn flap drops down into the crater (Fig. 3c). The significance of this activity with respect to the morphology of the crater is that impact melt and ejected debris effectively mantle the interior walls of the crater. Molten material can therefore flow along the walls, pond on terraces, and entrain ejecta boulders. However, as a consequence of terrace development, assumptions about the original stratigraphic position that could be inferred with the principle of the overturned flap are blurred. Materials on the near rim ejecta of craters such as Aristarchus are thus sampled from a greater depth than would otherwise be assumed.

1.2 Goals and Objectives

The overall goal of this dissertation is to use new, high-resolution, remote-sensing datasets to better understand the ejecta blanket emplacement process. This goal is addressed through the following questions:

1) How well can we apply our current understanding of crater formation to detailed investigations using high resolution observations?
2) Can the distribution of various morphologic and lithologic units excavated by a crater be used to reconstruct detailed pre-impact geologic history?
3) Can the timing of impact for individual impact craters be well-constrained?
4) Do longstanding tenets of ejecta blanket emplacement, for example ballistic sedimentation, hold up against new, higher resolution investigations of ejecta blankets?
5) What can we infer from detailed observations of terrestrial impact melts that can inform our knowledge of impact melt temperatures, viscosities, and lifetimes?
The following chapters address these questions through the detailed exploration and photogeologic mapping of the lunar crater Aristarchus, large-area crater size-frequency distributions on the ejecta blankets of Aristarchus and Tycho on the Moon, and laboratory analyses of zircon grains from a terrestrial impact melt glass from the Mistastin Lake impact structure, in Labrador Canada.

1.3 Major Results

1.3.1 Chapter 2

Aristarchus Crater is found in one of the most geologically complex regions of the Moon. It is one of the most prominent small features on the lunar nearside, and shines brightly to the naked eye within the dark mare regions of Oceanus Procellarum. At only 42 km in diameter, it is not a particularly large structure, but is probably one of the most significant impact craters on the Moon. Aristarchus is among the youngest complex lunar craters, with an extensive bright ray system, and has an exceptionally well-preserved morphology. The crater formed on the southeastern corner of the Aristarchus Plateau, a prominent 220 km x 170 km topographic platform embayed by mare basalts, and excavated material from both the high plateau and surrounding mare basalts. The excavated materials have been studied by remote-sensing techniques for decades, and are host to a number of different, and rare, lunar lithologies, including dark mantling pyroclastic glasses, mare basalts, olivine-rich materials, and highly silicic and felsic materials. Some of these materials are petrogenetically incompatible (i.e., olivine-rich units and silicic units), making the geologic history of the region very complex and worthy of detailed study.

In chapter 2, I use high-resolution imagery from the LROC instrument, including the narrow-angle camera (NAC) and wide-angle camera (WAC), to map the geomorphology of
Aristarchus Crater at 1:24,000 scale. The results of geomorphologic mapping provide information about the distribution if materials within the crater and on the ejecta blanket, including the distribution of impact melt products (melt ponds, flows, channels and veneers), and ejecta block fields (boulders). I further use compositional information from satellite remote sensing in conjunction with the geomorphologic map, in order to constrain excavation depths of various lithologies, and infer the pre-impact stratigraphy of the target.

On the basis of the study, I develop a new geologic history of the southern part of the Aristarchus Plateau. Compositional data indicate that Aristarchus excavated materials that are rich in thorium (e.g. Jolliff et al., 2004; Hagerty et al., 2006; 2009); have very low iron concentration (e.g. Jolliff et al., 2004; Lucey et al., 2001), and are likely highly silicic (e.g. Glotch et al., 2010), and that these same materials are present in other areas of the southern plateau. In particular, the Cobra Head volcano has a large deposit of compositionally similar material in the source region of Vallis Schröteri, the most prominent sinuous rille on the Moon. I suggest that the high-silica materials in Cobra Head, and the morphology of the volcano as a large topographic construct, are evidence that it was built through silicic volcanic processes, similar to those observed in other parts of the Moon (e.g. Compton Belkovich Volcanic Province; Jolliff et al., 2010). If this is the case, then the construction of Cobra Head was also accompanied by voluminous basaltic eruptions, and possibly indicates bi-modal volcanism on the Moon.

**Major results and new observations from this project are:**

- A new detailed geomorphologic map of Aristarchus Crater is produced; with 1:25000 scale mapping of the distribution of impact melt features and ejecta boulder fields.
  - Impact melt ponds and boulder fields are asymmetrically emplaced around the crater, likely due to pre-existing topography affecting ejecta emplacement.
• A globular melt morphology is observed on shallow slopes above steep cliffs, likely reflecting adhesion of melt to surfaces following rapid flow downslope.

• Crater walls and terrace widths are steeper and wider, respectively, on the elevated plateau side compared to the mare side, likely reflecting differences in bedrock properties.

• In general, morphologic units do not correlate with specific lithology. However, in some places impact melt features and boulders can be correlated with specific lithologies.

• Stratified ejecta boulders, likely broken-up packets of lava flows, were first observed and reported (Zanetti et al., 2011).

A synthesis of geologic remote sensing data to investigate the pre-impact stratigraphy of the Aristarchus Crater region, including the first use of WAC-Color products to describe the geology in detail.

• Ejecta facies show evidence for olivine-rich units, highly felsic and/or granitic units, and plateau and mare units.

• Pre-impact stratigraphy may include a shallow, layered pluton to account for highly felsic materials

• The southern portion of the Aristarchus Plateau may be better described as a large, silicic volcanic complex, and the Cobra Head volcano may represent large scale bimodal volcanism on the Moon.

1.3.2 Chapter 3

The high resolution imaging capabilities of the LRO-NAC were intended, in part, to allow for the dating of small geologic units by means of crater size-frequency measurements (CSFDs). However, soon after researchers tried to use CSFD measurements to date individual craters, by counting craters on various ejecta blanket units, it was discovered that impact melt ponds and the surrounding ejecta give disparate absolute model ages (AMA). At issue is, how can the surfaces of what are essentially the same aged surfaces show large discrepancies in AMA? The leading candidates for why a discrepancy exists are differences in target material properties and auto-
secondary cratering (van der Bogert et al., 2010; Plescia et al., 2011; Zanetti et al., 2012; . Target material properties can affect the final crater diameter for a given impactor size and velocity, whereby impacts into a hard, crystalline target (such as impact melt) would produce smaller craters than impacts into a less competent material (such as ejecta or regolith). The effect of target properties on diameter would therefore affect the CSFD measurement, and produce artificially younger AMAs. Auto-secondary cratering, the formation of craters on the ejecta blanket by late-arriving ejecta fragments from the parent crater, can explain the CSFD discrepancy as a result of more impacts occurring on the ejecta prior to impact melt emplacement (the last emplaced and longest-lived ejecta unit).

In order to investigate this question, I counted tens of thousands of craters on the continuous ejecta blankets of Aristarchus and Tycho Craters, with the intention of determining if the crater density and spatial distribution of small impacts on the ejecta was morphology dependent. I determined that impact melt ponds have nearly 30% fewer impact craters than surrounding ejecta units, and that based on their spatial distribution, the likely cause was related to the formation of the parent impact crater. My results show that auto-secondary cratering is the most likely explanation for the CSFD discrepancy between melt and ejecta units, and additionally suggest that the impact cratering flux in the inner solar system over the last billion years may be overestimated, possibly by a factor of 4. The formation mechanism of putative auto-secondary craters is not understood, and theoretical and numerical models do not account for these small features. A Lunar Data Analysis Program (LDAP) proposal is being submitted with Natasha Artemieva to evaluate possible modes of origin of the fragments and their emplacement through modelling and crater counting.
Major results and new observations from this project are:

- Crater populations on impact melt ponds are a factor of 4 less than on the ejecta and crater density increases with distance from the parent crater rim.
- Although target material properties may affect crater diameters, they alone cannot completely reconcile crater density and population differences observed within the ejecta blanket.
- Auto-secondary cratering contributes to the population of small craters (<300 m diameter) on ejecta blankets, and must be taken into account if small craters and small count areas are to be used for relative and absolute model age determinations on the Moon.
- Observations of impact craters on the ejecta blanket embayed by melt and “ghost” craters in impact melt support the existence of auto-secondary craters.
- Using the cratering flux recorded on Tycho impact melt deposits calibrated to accepted exposure age (109 ± 1.5 Ma) as ground truth, and using similar crater distribution analyses on impact melt at Aristarchus crater, we infer the age of Aristarchus Crater to be ~250 Ma.

1.3.3 Chapter 4

The previous two chapters made use of remote-sensing data from the Moon to characterize the Aristarchus Crater in detail, and use photogeology and crater counting techniques to constrain ejecta processes and absolute model ages of young craters on the Moon. The final chapter uses field observations and laboratory analyses to investigate terrestrial impact melts and the thermal decomposition of zircon.

In 2011, I was awarded the Barringer family fund for meteorite impact research which provided funding to join an expedition to the Mistastin Lake impact structure (Osinski et al., 2012). The Mistastin impact structure is a ~28 km diameter, ~36 Myr, complex crater in Northern Labrador, Canada. The structure is one of the few remaining large impact structures with abundant impact melt outcrops. The target rocks of the structure are primarily granodiorite, mangerite, and anorthosite (mostly labradorite), which make it a good analog crater for lunar highlands craters. The month long expedition, from Aug. 19 to Sept. 20, 2011 was organized by
Dr. Gordon Osinski at the University of Western Ontario. The goal of the expedition was to investigate the Mistastin impact structure as a geological analog for lunar highland craters and to provide a means of testing analog lunar sample-return mission scenarios. My role in the expedition was to assist in activities related to the analog mission scenarios, including robot traverse planning, sampling strategies, and astronaut training.

As part of this expedition, I collected a suite of impactite rock samples from various locations around the crater for subsequent analysis in thin sections. I was fortunate to find a very rare sample of impact melt glass that appears very similar to obsidian. The impact melt glass is completely glassy, with no crystallization textures, and represents a rapidly quenched melt. In another fortunate find, the first thin-section made of the glass contained a well-preserved zircon grain that had begun thermal decomposition. An interesting characteristic of the mineral zircon is that it doesn’t melt in the traditional sense, but undergoes an initial stage of solid-state dissociation into constituent mineral phases ZrO₂ (zirconia, with monoclinic form known as baddeleyite) and SiO₂ (silica). Subsequent thin-sectioning of the hand sample revealed a second decomposed zircon grain for analysis.

The excellent preservation state of the grains provided the opportunity to use state-of-art laboratory analytical techniques to study the decomposition of zircon in a natural sample. In 2013, I wrote successful proposals to the Mineralogical Society of America (MSA) grant for student research in mineralogy and petrology and the Geological Society of America (GSA) E. Shoemaker Impact Cratering Award to independently fund electron microprobe (EPMA), laser Raman spectroscopy (LRS), and secondary ion mass spectrometry (SIMS) analyses of the zircon grains, with the intent to characterize the thermal decomposition of zircon induced by impact melt, and attempt to independently determine the timing of the Mistastin impact structure by U-
Pb dating of the baddeleyite decomposition rim. As the samples represent a rare opportunity to study the decomposition of zircon, Dr. Edward Vicenzi of the Smithsonian Conservation Institute agreed to provide hyperspectral cathodoluminescence (CL) analyses, and Dr. Nicholas. Timms of Curtin University in Australia agreed to provide electron backscatter diffraction (EBSD) analyses to complement my study for subsequent publications.

I conclude that the zircon grains identified in thin-sections of the Mistastin Lake impact melt glass underwent partial thermal decomposition as a result of being entrained in high temperature impact melt. Impact melt infiltrated the decomposition zone prior to being quickly quenched, possibly as a result of rapid cooling adjacent to a mega-breccia block.

**Major Results from this investigation:**

- Decomposed zircon grains were discovered in a holohyaline impact melt from the Mistastin impact structure
- Major phases identified by LRS are zircon, baddeleyite, and tetragonal-ZrO$_2$, with no crystalline SiO$_2$ phases identified.
- Observation of tetragonal-ZrO$_2$ indicates melt temperature in excess of 1687$^\circ$C and melt viscosity of ~0.5 Pa*s (similar to motor oil).
- The zircon cores show no signs of shock metamorphism (apart from fracturing), and no high-pressure polymorphs (e.g. reidite) are observed.
- The ages of the zircon cores are constrained to ~1400 Ma, which is slightly below the ages of the target may have been marginally reset.
- Glass composition provided an estimate of provenance of the target rocks.
1.4 Statement of Labor

All work, except where indicated here and in later sections, was done by Michael Zanetti. Work in Chapter 2, regarding the Geomorphology and Geology of Aristarchus Crater, was done entirely by M. Zanetti using datasets publicly available in the Planetary Data System (PDS) and data from the Lunar Reconnaissance Orbiter Camera (LROC) system operations center (SOC) at Arizona State University. Image processing, including calibration, map projection, and georeferencing was done at Washington University in St Louis using the Integrated Software for Imagers and Spectrometers (ISIS). Digital elevation models (DEMs) were produced by LROC team members and supplied with consent for inclusion in the thesis and subsequent publications. Preliminary geomorphologic mapping was done my M. Zanetti at the Westfälisches Wilhelms – Universität Münster, Germany with Dr. H. Hiesinger. Geologic interpretations build on previous work by advisor B. Jolliff. For the work presented in Chapter 3, crater counting on the ejecta blanket of Aristarchus Crater was done entirely by M. Zanetti. Crater counting at Tycho Crater was done by graduate research assistant Amanda Stadermann and M. Zanetti. All analyses were done by M. Zanetti. In Chapter 4, electron microprobe (EPMA) analyses were done by M. Zanetti and assisted by Paul Carpenter. Laser Raman Spectroscopy (LRS) was done by M. Zanetti and assisted by Axel Wittmann and Jei Wei. Secondary Ion Mass Spectrometry (SIMS) was done by Alex Nemchin at the Naturhistoriska riksmuseet in Stockholm, Sweden and assisted M. Zanetti. Initial cathodoluminescence (CL) observations were made using EPMA by M. Zanetti and P. Carpenter. Hyperspectral-CL data, collected by Ed Vicenzi at the Smithsonian Museum Conservation Institute, Maryland, is only briefly reported on and is in preparation for publication with M. Zanetti as third co-author. Electron Backscatter Diffraction (EBSD) data, collected by Nick Timms at Curtin University, Australia, is only briefly reported on and is in
preparation for publication with M. Zanetti as second co-author. B. Jolliff provided melt-mixing calculations of the Mistastin Lake impact melt glass composition.

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References

Figures

Figure 1-1: Cross section of a theoretical transient crater, showing excavation flow lines, and zone of excavation (modified from Osinski and Pierazzo, 2012).
Figure 1-2: Conceptual diagram of ballistic sedimentation after the ideas of Oberbeck (1975), and modified from Melosh (1989), showing the position of the ejecta curtain at four time intervals after crater formation (T). Fragments follow ballistic trajectories from a launch point within the crater. Fragments near the crater rim are launched with low velocity and impact close to the crater rim. Bottom cartoons show the representative fraction of target material that is entrained based on ejecta velocity and distance from the crater rim.
Figure 1-3: Schematic diagram of the development of a central peak crater. A) Transient cavity prior to rebound; B) rebound begins, ejecta still lofted and moving radially outward; C) rebound continues, ejecta blanket emplaced, terrace development begins; D) final theoretical cross-section.
Chapter 2: The Geology and Geomorphology of Aristarchus Crater

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Abstract

Combining detailed morphologic mapping using Lunar Reconnaissance Orbiter Camera imagery, with compositional datasets including WAC – color, Diviner, Clementine spectral reflectance (CSR), Lunar Prospector, and interpretations from the Chandryaan Moon Mineralogical Mapper (M3) and Clementine ultraviolet and visible light spectroscopy (UVVIS) research, we characterize the geomorphology and geology of Aristarchus Crater in detail. Geomorphologic mapping shows that the distribution of impact melt and boulder fields are controlled by pre-existing topography, and provides evidence for long-lived impact melt mobility. Using principal component analysis of WAC color imagery, combined with compositional data, we created a new map of the geology of Aristarchus Crater and its ejecta. Comparisons between the morphologic mapping and geologic mapping show no major dependencies of composition with morphology, with the notable exception of compositionally distinct boulder fields. Impact melts show no unique compositional characteristics, and are not homogenized by the impact process. Silicic regions of the crater floor and ejecta are very well correlated with low FeO content and high Th abundance, possibly indicating the excavation of an evolved hypabyssal intrusion or pluton. Compositional anomalies on the Plateau near Herodotus Crater, Vaisala Crater, and the Cobra Head volcanic vent share the same composition as evolved excavated units in Aristarchus ejecta. Silicic material in Cobra Head suggests it may have formed as a silicic dome, despite being the source region of Vallis Schröteri, a prominent basaltic
The southern portion of the Aristarchus Plateau may have formed as a consequence of silica-rich intrusions, and bi-modal volcanism.

### 2.1 Introduction

#### 2.1.1 Geologic Background of the Aristarchus Region

The 42 km diameter, lunar crater Aristarchus is one of the brightest features on the nearside of the Moon and has been observed and commented on for more than a century (e.g. Herschel, 1787; Whitley, 1870; Birt, 1890). The crater (centered at 23.7°N, 47.5°W) impacted into the southeastern edge of the Aristarchus Plateau and is arguably one of the most important, unsampled craters on the Moon owing to the complex geology of the pre-impact target, which was part volcanic plateau, part mare basalt, and partly underlying substrate (likely Imbrium ejecta deposits and possibly other volcanic or intrusive rocks) (Fig. 1). Materials excavated by the crater are spatially highly variable and represent a number of different lithologies.

The western part of the crater lies at the edge of the Aristarchus Plateau, a topographic anomaly in Oceanus Procellarum that rises 1 – 2 km above the surrounding flood basalt plains and consists of uplifted crustal rocks (Guest, 1973) (Fig. 1). The composition of plateau materials is commonly interpreted to be mostly noritic in composition (Lucey et al., 1986; Chevrel et al., 2009), presumably a result of kilometers-thick deposits of Imbrium ejecta that likely covered the plateau and surrounding pre-Imbrium topography (e.g., Zisk et al., 1977; McEwen et al., 1994). The eastern part of Aristarchus impacted into mare basalts that surround and embay the plateau, concealing the Imbrium ejecta not deposited on the plateau. Overlying the plateau are the largest dark mantle deposit (DMD) on the Moon (Gaddis et al., 2003), interpreted to be fine-grained glass beads formed by gas-rich phases of basaltic volcanism (Wilson and Head, 1981). As a result of the pre-existing topography, and complex nature of the
target stratigraphy, the distribution of materials excavated by the impact is asymmetric, and reflects the different lithologies impacted (Guest and Spudis, 1985; McEwen et al., 1994; Mustard et al., 2011). The crater has a maximum relief of 4 km, measured from the deepest part of the crater floor (-3372 m) to the highest elevation of the western rim (+476 m), on the plateau (Fig. 2). The difference in height from the western (plateau) rim’s highest point to the rim’s lowest point on the eastern (mare) side is ~1.2 km.

In this study, we present and use a newly created geomorphologic map to compare the morphology of units such as impact melt and boulder streams with the geology of the crater interior and ejecta blanket as interpreted from remote-sensing data. The paper first provides a general sequence for the geologic history of the region prior to the Aristarchus Crater impact. We then present the results of our geomorphologic mapping, and compare these results with Lunar Reconnaissance Orbiter - Wide Angle Camera (Fig. 3. Robinson et al., 2010) color imagery and previous remote-sensing observations. We discuss a generalized geologic map of Aristarchus crater and ejecta facies created using image classification techniques from principal component analyses of WAC color data, and conclude with a discussion of the possible rock types excavated by the crater.

2.1.1.1 General Geologic History

The sequence of major geological events that has been proposed and is generally accepted, prior to the formation of the Aristarchus crater, is presented in chronological order below. This geologic history is a summary of the research by a number of previous workers, following mostly the interpretations of Zisk et al. (1977), Chevrel et al. (2009), and Mustard et al. (2011), and is presented here to provide context for our interpretations and major revision of these ideas presented in the discussion section of this paper.
A. Pre-Imbrium Terrain: Nearly all evidence for the pre-Imbrium topography and composition has been lost to subsequent deposition of Imbrium ejecta and mare flood basalt emplacement. It is possible that a few ghost craters on the plateau represent this surface, but this is unclear. It is also possible that Aristarchus excavated pre-Imbrium basement materials, particularly on the plateau side, but these ghost craters are not identified in any remote sensing data to this point (Zisk et al., 1977).

B. South AP crater: A 110 km crater was proposed by Mustard et al. (2011) to be responsible for the plateau scarp, and was suggested to be pre-Imbrian in age. Mustard et al. suggest that there is no remaining surface expression of this feature because it has been subsequently buried by mare basalts. This putative crater is discussed in more detail in section 5.6.1.

C. Imbrium Basin: The Imbrium Basin is a massive (1160 km diameter) impact structure on the Moon that formed ~3.9 Ga, and has played an important role in modifying much of the lunar nearside (Spudis et al., 1986; Wilhelms et al., 1989). The edges of the Aristarchus plateau are roughly linear, and it is suggested that the plateau is a fault-bounded, tilted block related to the formation of the ring structures of the Apennine-Carpathian Imbrium basin ring (e.g. Zisk et al., 1977; Guest and Spudis, 1985). Thick deposits of Imbrium ejecta have been emplaced on the plateau, with thickness estimates ranging from 1.5 to 4 km (Zisk et al., 1977; Haskin et al., 2003; ~1000 km from the center of Imbrium). Specific Imbrium ejecta facies that are suggested to be found on the plateau are Apenninus (concentric) facies and Fra Mauro (radial) facies (Zisk et al., 1977), but the bulk of the plateau is usually mapped as Alpes (unoriented) facies (Guest and Spudis, 1985; Spudis et al., 1988).

D. Orientale Basin: Although the ~3.7 Ga Orientale Basin is nearly 2000 km distant from the plateau; materials from this basin are likely present in relatively small quantities. Model thickness estimates from Haskin et al. (2003) suggest a minor (<100 m thick) covering; however, secondary cratering from Orientale likely contributed to the crater population on the plateau (e.g., Zisk et al., 1977).

E. Herodotus and Prinz: The large craters to the west and east, respectively, of Aristarchus post-date the emplacement of Imbrium ejecta and have subsequently been flooded by later-arriving mare flood basalts.

F. Pyroclastic glass deposits: Dark mantling deposits (DMDs) blanket much of the plateau and are interpreted to be pyroclastic glass beads formed from fire-fountain style eruptions (Zisk
et al., 1977; Gaddis and Pieters, 1985; Lucey et al., 1986; Ryder and Coombs, 1995; McEwen et al., 1994). The pyroclastic glass covers the plateau with 10 to 30 m thick (McEwen et al. 1994) or up to 50-100 m thick deposits (Zisk et al. 1977), based on estimated excavation depths of crater in the 100 – 1000 m diameter range. Since mare materials onlap the pyroclastic materials, leaving sharp boundaries around the plateau margins, the glassy materials likely pre-date, or are contemporaneous, with much of the mare flood basalt emplacement surrounding the plateau.

G. Mare flood basalts: Mare flood basalt units of Oceanus Procellarum embay the plateau on all sides. Hiesinger et al. (2003) determined the absolute model ages (AMAs) of these basalt units, and found that generally the units to the east and north are Eratosthenian in age (AMAs 2.76-2.0 Ga). Mare to the south of the plateau have an AMA of 1.2 Ga, indicating they are possibly the youngest mare units on the Moon (Hiesinger et al., 2003). Recent work by Stadermann et al. (2015) has confirmed this young AMA with a detailed, complete count of craters >400 m diameter in the large P60 count area of Hiesinger et al. (2003). Despite extensive contamination of the mare surface by secondary craters from Aristarchus, Stadermann et al. (2015) find AMAs derived for areas between rays correspond to similarly young ages (1.2-1.8 Ga), and suggest the unit cannot be more than ~1.8 Ga. These authors also note a number of small, previously unmapped vents in the region south of Aristarchus, and postulate that, although numerous sinuous rilles from the plateau empty into the surrounding mare, they are distant, and local sources likely contributed the majority of material (Stadermann et al., 2015). The basalts in the region south of Aristarchus Crater are ~300-500 m thick, based on rim height estimates of buried ghost craters and the methods of DeHon et al. (1978).

H. Aristarchus Crater: Aristarchus crater impacted the southeastern corner of the Aristarchus Plateau, and likely excavated plateau and mare materials, along with Imbrium ejecta and other distinctive lithologies (Guest, 1973; Zisk et al., 1977; McEwen et al., 1994; Zhang and Jolliff, 2002; Chevrel et al., 2009; Mustard et al., 2011).
2.1.1.2 General Morphometry of the crater

The interior of Aristarchus Crater is typical of a lunar complex impact crater, containing a central peak, a relatively flat floor, and terraced walls created during the modification stage of crater development (Fig. 4). Previous studies have characterized the structure and morphology of Aristarchus Crater through photogeologic mapping. Detailed mapping of flows on the ejecta blanket and units on the crater floor was first done by Strom and Fielder (1970). These researchers noted the lobate flow units in the ejecta and surmised that they were the result of multiphase volcanic eruptions based on their likeness to terrestrial lava flows and crater size-frequency distribution (CSFDs) between flow, floor, and ejecta units. Mapping by Guest (1973) focused on the nature and distribution of the ejected materials using 1:30,000 photogeologic mapping of areas of the ejecta blanket, and avoided remapping the interior (Fig. 6a). Guest (1973) noted that hummocky rim and blocky lobe material was only observed on the plateau side of the crater and that areas of ejecta containing concentric ridges were only observed on the mare side of the crater ejecta, and interpreted this as the result of differences in the emplacement of plateau and mare lithologies. Guest (1973) also extensively mapped a number of radial flow features and large, dark, lobate flows, which he inferred were related to impact processes and not subsequent volcanic activity (e.g., Strom and Fielder, 1970; Greeley and Gault, 1970). Guest and Spudis (1985) revised the map of Guest (1973) and added detailed mapping of the crater interior at 1:250,000 scale (Fig. 6b). These previous maps are the foundation of structural and geomorphologic studies of Aristarchus Crater.

Numerous observations of features on the floor and walls, which contain striking examples of channelized impact melt, flow lobes, ponding of material, and fracturing, made using Lunar Orbiter V imagery are contained in these previous works (Strom and Fielder, 1970; Guest, 1973;
Guest and Spudis, 1985) and in the Moon Morphology Book (Schultz, 1976). More recent observations of features at Aristarchus Crater using LRO-NAC imagery were presented by Mustard et al. (2011). Due to shadows obscuring much of the eastern crater interior in Lunar Orbiter V images, the opportunity was available to complete the map using Lunar Reconnaissance Orbiter Camera high-resolution imagery in this study. Our new mapping was conducted independently of these previous efforts, but we reached many of the same conclusions as previous workers. The high-resolution of the LRO-NAC images has allowed us to identify a number of previously unreported features at Aristarchus, including stratified boulders, impact melt pond collapses, and systematic asymmetries in ejecta units. In addition to previous work, new spectral datasets allow us to compare composition to morphology in greater detail. We discuss the similarities and differences in our observations in later sections.

2.1.2 General Geology of the Crater ejecta

Previous remote-sensing studies have noted the crater’s bright appearance and focused on the composition of materials in the central peak and ejecta facies, particularly ejecta southeast and southwest of the crater (e.g., Zisk et al., 1977; Lucey et al., 1986; McEwen et al., 1994; Le Mouelic et al., 1999; Chevrel et al., 2009). Telescopic spectral data suggested early-on that the Aristarchus Plateau, beneath the pyroclastic cover, was dominated by noritic compositions (orthopyroxene-plagioclase dominated rocks), and that the Aristarchus Crater region contained pyroxene-rich, olivine-rich, and feldspathic compositions (e.g., Zisk et al., 1977; Lucey et al., 1986), but did not have high enough resolution to determine the specific morphologic units to which these materials corresponded. Observations using Clementine data (Nozette et al., 1994; McEwen et al., 1994, LeMouelic et al., 1999; Chevrel et al., 2009) had the spatial resolution (~100 mpp) to allow relatively detailed analysis of the central peak, walls and impact melt units,
but lacked high spectral resolution. McEwen et al. (1994) noted olivine-rich materials in the southeastern ejecta and surmised that the central peak was predominantly feldspathic based on anomalous brightness and similarity to laboratory spectra for anorthite (Fig. 1b). Using integrated Clementine UVVIS and NIR spectra, LeMouelic et al. (1999) identified spatially extensive regions of olivine-rich materials in the southeastern crater ejecta. Radiative transfer mixing models presented by Lucey et al. (2008) suggested that much of the walls and floor of Aristarchus are mineralogically gabbronorites, but with low Mg/(Mg+Fe). Iterative linear mixing models of Clementine UVVIS spectra presented by Chevrel et al. (2009) were used to infer that regions of the central peak are likely anorthositic with some mafic component (probably clinopyroxene and olivine), that much of the northeastern crater wall (and plateau beneath DMD deposits) is dominated by anorthosite with Cpx and Opx contributions, and that the southwestern ejecta is mainly similar to the central peak materials but with a large glassy component (Fig. 7b).

Mustard et al. (2011) presented detailed observations of Aristarchus Crater using the hyperspectral Moon Mineralogical Mapper (M3) spectral measurements and RGB parameter mapping (R=integrated band depth (IBD) of 1 μm; G=IBD of 2 μm; B= apparent reflectance at 1.58 μm) and made comparisons with LRO-NAC morphology observations (Fig. 7a). Their mapping prominently highlighted the olivine-rich areas in the southwestern ejecta, and they found that the distribution of impact melt is strongly correlated with areas of olivine, feldspathic crust, and glass. The M3 results were interpreted to suggest that the central peak is strongly enriched in feldspar and is likely sourced from the upper plagioclase-rich crust of the Moon (Mustard et al. 2011), an interpretation similar to that of Ohtake et al. (2009) who interpreted the spectral evidence as indicating the occurrence of pure anorthosite (PAN) within Aristarchus using the Multiband Imager spectral dataset.
Additional compositional inferences can be made on the basis of other global datasets. Clementine-derived compositional data (e.g., Lucey et al., 2000; Wilcox et al., 2005) show low FeO content (~<10 wt%) in materials excavated in the southwestern areas of the ejecta blanket, and FeO content of the central peak ~<5 wt%. Shkuratov et al. (2005) modeled the global surface distribution of clinopyroxene (Cpx) using Clementine UVVIS, and noted Aristarchus as an anomalous crater with > 20% Cpx content. Lunar prospector Thorium (Th) abundances (Lawrence et al., 2003) show that the Aristarchus Crater is a circular “hot-spot” centered broadly over the crater, with estimated Th abundances of >10 ppm and modeled abundances >15 ppm (Hagerty et al., 2009); these results are similar to those measured using the Apollo 15 gamma ray spectrometer (18.2-21.7 ppm; Etchegaray-Ramirez et al., 1983). Glotch et al. (2010), using Diviner radiometer Christiansen Feature positions, inferred that regions of the crater interior, including the central peak, and large portions of the southwestern ejecta blanket are likely silica-rich, and consist of quartz, Si-rich glass, or alkali feldspars (or mixtures thereof).

The complicated geology as seen in remote sensing data has led to a number of hypotheses about the origin of the excavated materials. General agreement exists that the central peak materials are high albedo and rich in feldspar, but whether this feldspar is similar to typical lunar highlands or represents a more alkalic species is debated. If the material is simply similar to highlands materials, then the central peak region likely represents exposure of a global layer of plagioclase/anorthosite lunar magma ocean flotation crust (e.g. McEwen et al., 1994; Ohtake et al., 2009; Mustard et al., 2011). Jolliff (2004) suggested the association of materials with low FeO contents and high Th abundances could be related to the excavation of highly evolved lithologies, such as monzogabbro, or possibly K-feldspar-rich granite, or alkali anorthosite. Similar suggestions of the presence of evolved materials have been made by Hagerty et al.
(2009) on the basis of Th abundance, and Glotch et al. (2010) on the basis of Diviner CF signatures, and Zhang and Jolliff (2008) and Zanetti et al. (2012) based on multiple datasets. The distribution and origin of geologic units is discussed in more detail in later sections of this paper.

2.1.3 Age Estimates

On the basis of Aristarchus Crater’s fresh appearing morphology, few superposing craters, and extensive bright ray system, it is clearly Copernican in age. However, an absolute age for the formation of Aristarchus is lacking because the crater was not sampled during the Apollo program, and no returned samples are suggested to have come from it (Heiken et al., 1991). Based on stratigraphic superposition of bright crater rays, the formation of Aristarchus Crater post-dates the formation of Copernicus Crater (~850 Ma – 1 Ga; Stöffler and Ryder, 2001) and Kepler Crater (625-950 Ma; Koenig et al., 1977), but formed prior to Tycho Crater (~109 Ma; Arvidson et al., 1976; Hiesinger et al., 2012). Crater size-frequency distribution measurements have been used by a number of researchers to determine relative and absolute surface ages using remote sensing images (e.g., Hartmann, 1968; Strom and Fielder, 1970; Guest, 1973; Neukum and Koenig, 1976; Koenig and Neukum, 1977; Zisk et al., 1977).

Early estimates for the age of Aristarchus Crater, based on poorly calibrated or uncalibrated crater production functions, range considerably depending on where around the crater the counts were done and by which researcher. Counts of the ejecta blanket made by Strom and Fielder (1970) and Hartmann (1968) suggested an age of ~1.0-1.1 Gyr. Strom and Fielder (1970) also made detailed counts of the crater floor and northern rim units, and suggested and age of 400-460 Mya for ejecta flow units, and ~230 Ma for floor units, which bolstered their view that the ejecta flow units and floor were volcanic in nature. Based on a crater density at Aristarchus Crater that is about half of that of Copernicus Crater, Zisk et al. (1977) suggested an
age of ~ 450 Ma, assuming a constant impact flux. Extensive refinement of the lunar chronology function was conducted by Neukum and Hartmann, and many others throughout the 1970’s and 1980’s (e.g., Neukum et al., 1994; Neukum et al., 2001; Ivanov et al., 2001). Ages for Aristarchus Crater using the revised production functions given by Neukum and Koenig (1976) and Koenig and Neukum (1977) are 130-180 Ma. Zanetti et al. (2011, 2012, and 2013) suggested formation ages of 160-195 Ma using very large area counts and the production functions of Neukum et al. (2001). Zanetti et al. (2015, in revision; Chapter 3 of this thesis) present evidence for self-secondary contamination of the ejecta blankets of Aristarchus and Tycho, and suggest that impact melt ponds best record the true flux of impacts. They postulate that if the cratering rate has indeed been lower in the last Ga, then Aristarchus is on the order of 250 Ma (Zanetti et al., 2015, in revision; Chapter 3 of this thesis).

2.1.4 Objectives

The objectives of this study were to characterize the geomorphology of Aristarchus Crater and compare spatial relationships of different geomorphologic and geologic units. We use these observations to investigate excavated materials in the ejecta blanket and crater walls and bring these materials into context to determine possible pre-impact stratigraphic relationships.

2.2 Data and Methods

2.2.1 Datasets

Lunar Reconnaissance Orbiter Camera (LROC) Narrow Angle Camera (NAC) and Wide Angle Camera (WAC) images (Robinson et al., 2010) were used extensively for photogeologic mapping and geologic context interpretation. The Narrow Angle Camera (NAC) system consists of a pair of identical 700 mm telescopic CCD line-array imagers. Combined, the left and right cameras (designated NAC-L and NAC-R) provide 0.5 m/pixel spatial resolution panchromatic
images of the surface at 50 km altitude. The WAC is a 7-color push-frame camera which uses a 1000 x 1000 pixel charged coupled device (CCD). The camera’s 7 narrow-band interference filters, two in the ultraviolet (321 and 360 nm) and 5 in the visible spectrum (415, 565, 605, 645, and 690 nm), are used to characterize spectral features associated with regolith composition and maturity. The visible spectrum imaging has a ground spatial resolution of 75 m/pixel, and the ultraviolet system has a spatial resolution of 384 m/pixel at 50 km altitude, in the nadir position (Robinson et al., 2010). The 605 nm filter is also used for monochrome imaging of the surface. The color bands were specifically chosen to complement Clementine UVVIS and NIR datasets (415, 750, 900, 950, 1000 nm and 1100, 1250, 1500, 2000, 2600, 2780 nm respectively) (Robinson et al., 2010; Nozette et al., 1994).

Geomorphologic mapping of the crater interior was mostly completed using the 1.2 m/pixel Aristarchus Crater Controlled Mosaic, with more distal areas of the ejecta blanket mapped by individual NAC Frames. Highest resolution NAC imagery (<.5 m/pixel) were used went available to characterize features as small as 2 m in diameter (e.g., Stratified Boulders, see Section 3.XX). LROC-WAC monochromatic imagery (<100 m/pixel) was used to provide geologic context and WAC color products (~400 m/pixel; Sato et al., 2014) were used for comparison with existing multi- and hyper- spectral datasets. Kaguya Terrain Camera (K-TC; Chin et al., 2007) imagery was used to provide additional context and fill data gaps.

Topographic measurements were made using the 100 m/pixel LRO-WAC stereo Global Lunar Digital Terrain Model (GLD100; Scholten et al., 2012), and 2-5 m/pixel LROC-NAC Digital Elevation Models (DEMs) when available (created by M. Henriksen at ASU). Slope maps and topographic cross-sections from the GLD100 were made using native tools in ArcGIS.
2.2.1.1 Compositional Data

Clementine UVVIS / NIR multispectral data (Nozette et al., 1995) and Clementine derived iron abundance maps (presented as wt% FeO) (e.g., Lawrence et al., 2002; Lucey et al., 2002; Gillis et al., 2004) are used to determine the spatial variability of FeO in the ejected materials. Spectral unit maps derived from PCA and iterative linear mixture modelling (ILMM) of Clementine UVVIS/NIR data made by Chevrel et al. (2009) were integrated and georeferenced into an ArcGIS map project and used for comparison to geomorphologic features. Chevrel et al. (2009) mapped 5 distinct units in the area within and around Aristarchus Crater (Fig. 7b).

The LRO-Diviner radiometer is a multispectral push broom sensor collecting information in 9 spectral channels at high resolution of 200 m/pixel at lunar equator (Paige et al., 2009). Out of the nine channels, three channels, b3, b4 and b5 are “mineralogical channels” located near 8 μm. These channels centered around 7.8, 8.2 and 8.6 μm are designed to characterize the Christiansen feature (CF), an infrared emissivity maximum feature that is sensitive to the silicate polymerization (Greenhagen et al., 2010; Glotch et al., 2010). Christiansen Feature maps (Paige et al., 2009) were obtained from the PDS and the Diviner team (K. Shirley, personal communication, Sept 2015), and are corrected for local noon temperatures. Laboratory emissivity experiments have shown that the CF is related to the degree of silica polymerization (Greenhagen et al., 2010). In felsic minerals the CF occurs at shorter wavelengths, and for mafic minerals the feature occurs at longer wavelengths. Minerals with “concave-up” signatures in laboratory measurements of CF include quartz (7.2 μm emissivity maximum), albite (~7.7 μm), and labradorite (~7.75 μm); and anorthite has a nearly flat, but positive value c-index and CF value of 8 μm (Glotch et al., 2010; their Fig 2). For reference, the pure plagioclase regions on the Orientale Inner Rook Ring observed in M3 data by Cheek et al. (2013) have a modeled CF value.
of 7.9 µm, with a positive c-index value (Glotch et al., 2010), and the mare basalts surrounding the plateau have average CF positions >8.25 µm (our measurements).

Lunar Prospector Thorium (Th) abundance maps (e.g., Lawrence et al., 2002; Hagerty et al., 2009) show the concentration of Th-rich materials related to areas of the Aristarchus crater floor and ejecta. The M3 parameter maps presented by Mustard et al. (2011) were integrated and georeferenced into our ArcGIS map project for comparison to the geomorphologic map (Fig. 7a). The M3 parameter map is an RGB composite image calculated from M3 data, where red is the integrated band depth (IBD) at 1 µm, green is the IBD at 2 µm, and blue is the apparent reflectance at 1.58 µm. In general, red areas in the map are interpreted to be olivine-rich areas, green regions are dominated by pyroxene absorptions, and blue areas are spectrally indistinct, with no mafic absorptions and show a melt texture (Mustard et al., 2011; 2012). The Gravity Recovery and Interior Laboratory (GRAIL) Bouguer anomaly maps were used to establish plateau margins and provide context for sub-surface interpretations (Zuber et al., 2013; Lemoine et al., 2014).

2.2.1.2 WAC-Color Ratio Images

Although not used in this work to derive quantitative compositional and mineralogical information, we used WAC color ratio maps to infer relationships about unit maturity and mineralogic similarity. Color ratio maps, created from WAC color bands (Fig. 8), serve to cancel out the dominant brightness variations of the image, which are controlled by albedo differences and topographic shading effects, in a manner similar to color ratio maps from Clementine UVVIS (McEwen et al., 1994). Fig. 8a shows an RGB composite image with band ratios of 689/321 nm in the red channel, 415 nm in the green channel, and 321/689 nm in the blue channel (a close-up of the crater viewed in this ratio is shown in Fig. 10a). Fig. 8b shows an RGB
composite ratio image focused on the ultraviolet end of the spectrum measured by WAC; with 415/321 nm in red, 360 nm in green and 415/321 nm in blue. Both datasets reveal differences in composition in the makeup of materials in the crater floor, walls, and ejecta, and show important correlations between units in Aristarchus ejecta and exposed units on the plateau.

2.2.2 Methods

2.2.2.1 Geologic Mapping

Photogeologic mapping was done using EsriTM ArcGIS 10.1 software. Units were drawn by hand as shape files and feature classes. Datasets were obtained from multiple sources, including the Planetary Data System (PDS) Lunar Orbital Data Explorer and Map-a-Planet resources, LROC-Science Operations Center (SOC), and the Kaguya SELENE Data Archive; and processed with the United States Geological Survey’s Integrated Software for Imagers and Spectrometers (ISIS) software. All datasets were radiometrically corrected, map projected, and georeferenced to various points within and around the crater when necessary.

2.2.2.2 Principal Component Analysis of WAC-color bands

Principal component analysis (PCA) was done on the 7 band WAC color spectral dataset using the principal component and spatial analysis tools in ArcMap 10.1. PCA is a statistical tool that identifies the variables that can explain the most variation within a multivariate dataset by transforming the input band (in our case 7 WAC bands) through rotating the multivariate attribute space orthogonally with respect to the original space. As a tool in ArcMap 10.1, PCA is an automated process, identifying eigenvectors (direction in multivariate space of the principal component of interest) and eigenvalues (magnitude or spread of the data) of the first 3 principal components, representing more than 95% of the total variance of the dataset. The covariance
matrix, correlation matrix, and Eigenvectors and Eigenvalues derived from the WAC-color PCA are presented in table 2.

An RGB composite image of the first three principal components is shown in Fig. 9 (close-up view of the crater interior in Fig. 10b). In this image, the red channel is assigned to areas with the greatest variance (i.e., the first principal component), the green channel is assigned to areas that show the second most variance not described by the first principle component, and the blue channel is assigned to the smallest variance areas between the bands of the dataset. In short, areas with a large red color component show the most variance compared to other values in the dataset, and areas in blue have similar reflectance across all 7 bands of the dataset. The PCA is used in this work to identify regions that share spectral similarities in terms of spectrum shape and reflectance, and that are likely genetically related.

2.2.2.3 Image Classification

Supervised image classification tools in ArcMap 10.1 were used to identify and extrapolate regions of the PCA map with similar characteristics. This classification was done by identifying representative groups of pixels with similar color, spatial distribution, and underlying morphology. For example, the red pixels of the central peak were grouped together because of their color and location on the central peak. A polygon was drawn around them and the image classification tool searched identified areas in the scene with similar characteristics, identifying only the area on the central peak (not unexpected due to the high degree of variance of the central peak in the PCA). In more chaotic areas, for example the yellow pixels on the wall of the crater in the PCA, multiple locations that shared the same morphology (in this case boulder fields) were mapped by polygons can combined into a larger sample size. We identified 10 regions of the PCA that could accurately represent a schematic overview of the spectral variation
in the PCA, and created a classified image map. From this classified image, a tool developed to streamline color-based geologic mapping and generalize and smooth raster datasets was used to create shapefile polygons of the different classified units (Baum and Zanetti, 2015). This tool, called Raster to Generalized Polygons (R2GP), works by taking any raster image and converts it to generalized polygon shapefiles after identifying and removing outlier pixels (based on user-defined inputs) and smoothing unit boundaries (Baum and Zanetti, 2015). The output of this tool when applied to the PCA image is used as our schematic geologic map (Fig. 30). The value of this tool is that illustrative mapping can be done in a fraction of the time it may take to map by hand (~15 minutes), when starting with a sufficiently well-classified raster image. The resulting geologic map is representative of the surface geology, but, like most schematic geologic maps, may suffer from over-simplification in some areas. For example, detailed distribution of pink-colored areas in the PCA, particularly on the western wall of the crater, is lost through the generalized mapping procedure. However, a high enough level of detail is preserved to build a cohesive geologic history of the region.

2.3 Results:

In the following sections we present the general morphometry of Aristarchus Crater based on measurements made with the LROC-WAC GLD100, followed by the results of our geomorphologic and WAC color geologic mapping. Our geomorphologic map contains 11 morphology units organized by Crater Floor Units: central peak, hummocky floor, and smooth floor deposits; Crater Wall Units: Terrace Scarps, Chaotic Deposits, Channeled and Veneered Walls, and Interior Walls, and Ejecta Units: Concentric-Ridged Ejecta and areas of Melt-Rich Channels and Lobes. We focus on the mapped distribution of impactites, including impact melt features and boulder streams, both of which are found both interior and exterior to the crater rim.
Our map is roughly 1: 24,000, with some units, such as stratified block units mapped at higher resolution. Many of the units and features described follow observations and mapping of Guest (1973), Schultz (1976), and Guest and Spudis (1985).

2.3.1 Morphometry

2.3.1.1 Topography

The WAC stereo DEM draped over a shaded relief map of the region provides an good context for the regional geology (Fig. 2b). Figure 5 a and b are topographic profiles oriented West to East that follow the 23.7 °N latitude line, and North – South along 47.5 W longitude, respectively. Both of these profiles cross the central peak of Aristarchus Crater.

Regionally, the plateau near the Aristarchus Crater ranges between 800 m to 1000 m higher than the surrounding mare. The prominent scarp on the southern side of the plateau has as much as 2 km of relief above the mare in the area south of the Herodotus crater rim. The plateau scarp is partially, but not completely buried by Aristarchus Crater ejecta, and ridge line can be observed to extend beneath the ejecta to the crater rim. The margins of the northeastern side of the plateau are more difficult to discern. We have drawn the general trace of the plateau based on the apparent topography and GRAIL Bouguer gravity signature (Fig 11). The topographic rise south of the Cobra Head feature, at 1453 m elevation, dominates the plateau. The total relief in the topographic map of figure 2b is almost 5 km, and the distance between the highest and lowest points is only 50 km. The 100 m/pixel resolution of the DEM is high enough to resolve the secondary channel which is seen in Valles Schröteri, and many other small volcanic rilles and mounds on the plateau.

The Aristarchus Crater rim is a well-defined circular feature with an average diameter of ~42 km. Using the WAC GLD100 stereo DEM we measure the highest point on the crater rim, found
on the western plateau side, to be 476 m above the global reference datum. The lowest point on the rim is found in the eastern mare side, and is 700 m below the datum. The maximum difference in relief between the western and eastern rim is ~1.2 km. The deepest points in the crater lie 5 km east of the central peak and are ~3372 m below the global reference datum. The crater floor is approximately 20 km in diameter, generally flat, and is populated by a few domes of slumped material and a small, but prominent central peak. The central peak rises ~300 m above the surrounding crater, and is ~3 km long and ~1.75 km wide. The central peak lies approximately 1.5 km closer to the southwestern rim than the northeastern rim, and is approximately equidistant to the northwestern and southeastern rims (Fig 5).

The topography of the continuous ejecta blanket, within 1 crater radius of the rim, differs greatly between the plateau and mare hemispheres of the crater. The plateau portion has considerably less relief, and is relatively flat west of the crater rim, where no discernable ejecta blanket is seen (Fig 5). The eastern side of the crater the ejecta blanket consists of a distinct raised rim and extended concave-down slope toward the flat plains of the mare. In the topographic profile (Fig 5), a sharp flat terrace on the eastern wall of the crater can be seen at ~1600 m. The North-South profile is typical of a complex impact crater, and a raised rim and proximal ejecta blanket are observed on both sides of the crater. There is a slight elevation difference of ~400 m, where the northern side is elevated. The walls of the crater on the western (plateau) side are steeper (avg. ~17º) compared to the eastern (mare) slopes (avg. ~14º).

We have tested a number of our morphometric measured values against predicted values from crater scaling laws from Pike (1977) and Melosh (1989) (Table 1). The values of properties such as crater depth and rim height can be determined from simple calculations based on rim to rim diameter. Using the following relationships we created Table 1, which is a comparison of our
values against predicted values from Pike (1977) and Melosh (1989). This comparison is done primarily to give context to Aristarchus Crater in relationship with other impact craters on the Moon, and to test if there is a possible target material relationship between the hemispheres of the crater.

2.3.2 Geomorphologic Map

The following sections explain the units of our geologic map (Fig. 12). A NAC controlled mosaic context image of the crater with figure locations is shown in Fig. 13.

2.3.2.1 Crater Floor Units

Smooth Floor

Smooth floor deposits are areas of the crater floor that are largely free from hills and hummocks that make up the hummocky floor unit. The smooth floor unit has a ropey and wavy texture, and contains many rough appearing ridges and large open fractures; it likely represents the impact melt sheet. Figure 14a shows a typical area of the crater floor containing these smooth deposits. Large open fractures are meters to tens of meters wide and vary from a few tens of meters in length to more than 1 km. Small hills and hillocks are observed to be embayed by the smooth floor deposits, and in many cases the small hills appear draped by impact melt. The impact melt likely veneered many of the hills which now show boulders and signs of erosion. Much of the edge of the crater floor is ringed by curvilinear fractures parallel to the crater wall (floor fractures structural unit). These fractures appear along the edge of the floor, where there is a slight change in topography. These curvilinear fractures are concentric to the crater walls and are similar to “bathtub ring” deposits formed when the melt sheet was molten and at its highest point. The top of the melt sheet solidified and the remaining melt either drained into subsurface
cracks, or the sheet contracted while cooling, leaving the high-stand of material. Erratic blocks of ejecta are seen on the smooth floor unit, indicating they arrived after the melt sheet had crystallized. Numerous craters < 100 m in diameter are seen; however, the wavy, ridged, and fractured nature of the deposits make crater-counting difficult. Craters on the smooth deposits are typically very small and shallow.

**Hummocky Floor**

Large areas of the floor of Aristarchus are covered in large hills and hummocks (Fig. 14b). Most of this unit occurs along the northern floor of the crater and in areas surrounding the central peak. The hills and hummocks have a mammillary or bulbous appearance, and are typically rounded mounds with no ridges. The mounds are often overlapping, appearing as stacked appearing hills. Smooth floor material and some smaller fractures are often found within the mounds, and all areas of the hummocky deposits appear to have been embayed by the smooth floor deposits. Fresh-appearing blocky rubble occurs on the tops and sides of the hills, although recent boulder tracks are not observed. Hummocky Floor Units cover an area of roughly 93 km².

**Central Peak**

The central peak of Aristarchus crater is centered at 23.71°N and 47.52°W. It measures approximately 3.2 km in length from north to south, and is 1.8 km wide. It is part of an arcuate band of small hills and hummocks in the middle of the crater. The height of the peak, as measured using NAC DEMs is ~472 m (measured using the WAC GLD100 it is very close to 300 m), and smooth floor units embay the west side of the peak approximately 100 m in elevation higher than the east side.

The smooth floor melt unit sharply embays the central peak, which shows three zones with distinct differences in reflectance (e.g., Robinson et al., 2010; Mustard et al., 2011). Seen in
Figure 15, the northern section (a) appears dark gray, the central section (b) appears light gray, and has a high albedo, and the southernmost section (c) has a transitional albedo. Dark albedo material is seen between sections A and B, where numerous large boulders and blocks (10 m diameter) mantle the peak. The northern-most section of the peak appears to be covered by a smooth mantling, likely melt-sheet material. At the base of the western flank of (b) it appears that bright material has moved down slope and forms a small debris fan on the floor. The eastern flank of the bright albedo unit (b) appears to be mottled with dark plates of material, which we interpret as thin crusts of impact melt, or dark underlying units that are mantled with high albedo debris eroding from the peak ridge. The southern portion (d) also contains many small boulders observed in chains, starting at the peak of the mound and extending to the floor. The southern portion appears to have a similar platy texture to the east flank of the bright central region. A small hill near the central peak (e) has a reflectance similar to area (d).

2.3.2.2 Crater Wall Units

The walls of Aristarchus Crater show signs of intense deformation related to massive blocks of materials that have slumped along curvilinear, inward-dipping, listric faults. Terraces of varying widths (<100 m to ~2 km) have developed around the crater and define large fault blocks. These terraces often serve as catchment regions for impact melt that has flowed down the surface of the walls (Fig. 16). Typically, the surface of the fault blocks is covered in a melt veneer, and flow features indicating varying viscosities of melt are observed. Large lobes of apparently thick, viscous melt are observed, as well as thinner, less viscous flow veneers. The uppermost fault surfaces are relatively melt-veneer free, and have a fresh appearance with only a very thin melt veneer covering. The fault surfaces along the rim are very well exposed, and the scarps are in some cases more than 500 m high. The fault scarps and terraces in the middle
portion of the wall are shallower and narrower, and mantled in thicker melt veneer, resulting in a mottled and hummocky appearance. Blocky rubble from the rim crest has fallen over the walls in some places, and boulder tracks and minor debris flows are observed, mostly on the uppermost fault surfaces.

**Terrace Scarps**

The most prominent terraces in the crater are on the east and west walls, and areas in the northern and southern terrace walls generally lack well-defined terraces like those seen in the east and west. The western (plateau) portion of the crater has 5-6 well developed terraces, compared to 3-4 terraces in the eastern (mare) portion. Terraces in the west are also, on average, a few hundred meters wider than those in the east. These areas occupy an area of roughly 96 km².

**Channeled and Veneered Walls**

Channeled and veneered walls are characterized by numerous channels, commonly displaying levees, flowing down the crater walls. This unit has a relatively smooth appearance owing to thick veneers of melt that occur between the channelized flows. Areas of channeled and veneered walls have poorly developed terraces, but melt channels do terminate into perched melt ponds. Some channels end in lobate, debris-filled lobes on the walls of the crater, indicating some flows were of higher viscosity than others. Single channels can be seen to extend nearly the > 10 km length of the crater wall. The channels also commonly begin at melt ponds on the terraces, and we observe areas where these melt ponds have drained subsequent to the melt beginning to solidify (e.g., Fig. 16). These areas occupy roughly an area of 504 km².

**Interior Walls**
The interior crater walls are only thinly veneered by impact melt features, and little channelized melt is observed. This unit is predominantly found in the steep western wall and along the steepest slopes of the interior crater rim, topographically above the channeled and veneered walls. Melt has likely completely drained from these areas, leaving only a thin veneer. These areas occupy roughly an area of 317 km².

**Chaotic Walls**

Areas of chaotic hills and hummocks are found in the northeastern and eastern areas of the crater walls. These areas lack clearly defined terraces and resemble areas of hummocky floor. Channeled melt flows often feed into the chaotic terrain, leaving small perched melt ponds, but this unit lacks the smooth appearance of the channeled and veneered walls. These areas occupy roughly an area of 58 km².

**2.3.2.3 Crater Ejecta Units**

The distal ejecta blanket is mapped according to the definition by Melosh (1989) and represents the area beyond the continuous zone of ejecta in the proximal ejecta blanket. The location of the farthest extent of the distal ejecta is somewhat subjective, and we marked the edge as the approximate location of a noticeable change in crater density and albedo variation. The furthest point coincides with the location of the large vertical scarp on Aristarchus Plateau, and is seen as a long bright albedo deposit. It is characterized by a discontinuous zone of hillocks and dunes closest to the proximal ejecta that grade into smoother terrain containing numerous clusters and chains of secondary impact craters. The albedo of the distal ejecta is brighter than the surrounding mare material in the east and southeast. The albedo is brightest to the southwest (also where it extends the farthest). Secondary craters and chains occur closest to the crater in the
northwest and west, which is most likely due to interference of the ejecta with the topography of the Aristarchus Plateau.

The proximal ejecta blanket is located within approximately one crater radius from the crater rim (Melosh, 1989) and is probably the region of the ejecta blanket that contains the bulk of the excavated material created by the excavation process (after Melosh, 1989, but with the significant caveat that volume estimates are fraught with uncertainties, including uplifting of the crater rim and the incorporation of substrate material during ballistc sedimentation). We have mapped the proximal (also known as continuous) ejecta blanket as the region of the crater exterior between the rim and beginning of herringbone and dune-like forms at approximately one crater radius.

**Concentric Ridged Ejecta**

The eastern hemisphere of the crater ejecta, nearest the crater rim, is characterized by a series of meters-high ridges that parallel the crater rim (Fig. 17). Impact melt commonly ponds in these concentric ridges, creating linear ponds before the melt continues to flow down the ejecta slope, and much of this surface is covered with a thin melt veneer (marked with the letter p, Fig. 17). Some concentric ridges are folded and sub-parallel to the crater rim. This unit was mapped in great detail in 2 large sections by Guest (1973), who noted that the ridges have the appearance of small roches moutonnées, and suggested that they are an erosion feature caused by high velocity surges of debris during crater excavation. Schultz (1976) interpreted the features as either concentric fractures or pressure ridges associated with interior bulk ejecta movement. Because these features were emplaced as part of the ejecta curtain they did not experience the passage of the main shock wave or later release of stress (as suggested by Schultz, 1976), but may be related to downward slumping of terraces during the modification stage. Melt channels
and flows commonly cut across and incise the concentric ridges, thus the concentric ridges are probably the “bedrock” unit of the ejecta blanket.

**Melt Rich Channels and Lobes**

A large number of melt morphology features characterizes the ejecta nearest the rim in the western part of the crater and the distal parts of the eastern ejecta. This unit contains channels and flow lobe termini indicating melt-rich ejecta played a large role in its formation (Fig. 18). In the eastern ejecta, channels that extend from the crater rim, cross-cutting the concentric ridges, end in a large number of melt pools and flow lobes. The channels and lobes end entirely within the continuous ejecta, before the onset of hummocky dunes and hilly ridges that mark the beginning of the discontinuous ejecta. In the western part of the crater, the channels and lobes begin at the crater rim and extend radially away from the crater. Many of the channels in the ejecta are leveed, and are likely to have only been active once. The channels end in flow lobes that are boulder- and clast-rich, indicating a change in viscosity, probably due to the incorporation of colder clasts. The channels and lobes on the rim have a hummocky appearance, which led Guest (1973) to map it as such. This unit contains the greatest proportion of impact melt flows and ponds.

**2.3.2.4 Impact Melt Features**

Impact melt features are ubiquitous around the Aristarchus Crater and are observed both interior and exterior to the crater rim. The work of early researchers, most notably Howard and Wilshire (1975), showed that the distinctive morphologic features of impact melt deposits can be observed in four groups. Melt can be observed as a superimposed, smooth veneer over irregular surfaces; as flow lobes and channels; as small ponds on crater walls and rims; and as complexly
fractured pools on crater floors (Howard and Wilshire, 1975). Impact melt that has moved down slope is observed to be both erosive and depositional. Depositional lobes have distinct end scarps and appear as flow fronts. When the melt is more viscous, the flow can deposit levees on the sides of channels. In steeper terrain, when the flows have moved quickly, it appears that the impact melt can erode into the walls of the crater (Bray et al., 2009). In order for this to occur, the flow of material must be fast, turbulent, and of low viscosity (Hulme, 1973), but it is unclear if the erosion into the wall is mechanical or thermal, or a combination of the two.

For the purposes of mapping, we identify impact melt ponds as flat areas of dark material resembling lakes or lava, which often contain cracks and fissures, likely related to cooling. Dark flows are also observed, which resemble lava flows or long runout landslides. We also consider units that have morphologies similar to low viscosity fluid flows, “splash”-forms, and globular textures with similar albedo an appearance to ponded areas to be melt. We describe these units and distribution in detail and provide context figures for previously undescribed or unique melt features of Aristarchus (e.g., Strom and Fielder, 1970; Howard and Wilshire, 1975; Schultz 1976; Guest and Spudis, 1985; Mustard et al., 2011). Within our map, three units are related to the distribution of impact melt. Ponded areas of smooth melt are marked in bright red, enigmatic areas of a globular and mounded melt, which have a globular-like appearance, are marked in deep red/burgundy, and surfaces that appear to be covered in a thin melt veneer are marked with a red dotted pattern. Here we describe and give examples of each unit individually.

**Interior Melt Features**

Melt pools within the interior of the crater are found on terraces between major fault blocks and among the hummocky deposits of wall-derived material near the crater floor. They vary greatly in size, and are controlled by area available on the terrace in which they form and by
the amount of melt available at the time of their formation. The surface of the ponds is normally smooth, but cracks and fractures can be found on almost all ponded surfaces. These cracks are, in most cases, probably related to thermal contraction of the surface during cooling, but may also be related to extension of the surface if underlying molten material is drained away. Fresh appearing small blocks and down-wasted debris from the adjacent crater walls can commonly be found on the pond surfaces, indicating that they had already solidified. Impact craters on the surface of the melts can vary considerably in their morphology, and crater density can vary greatly, even between adjacent units (Zanetti et al., 2012).

The largest terrace melt pond is shown in Fig. 19. It is located approximately 400 m above the crater floor, along a portion of the northern wall of Aristarchus Crater. The melt pool lies approximately 300 meters above the crater floor, and 2.8 vertical kilometers below the crater rim. The pool is approximately 800 m wide and 2.3 km in length along the terrace, and has a surface area of ~ 1.5 km² (Fig. 19). This pool is unique within Aristarchus in that is shows evidence of a collapsed roof following drainage of the pond. Large channels flow into the pool and out of the pond (red arrows). Within the pond itself are sharply defined scarps and extensional fractures and faults that provide evidence of collapse. The main elevation of the pond surface was at -2765 m, as evidenced by a high-stand plateau of melt. The area within the pond has drained away to the southeast, and the pond floor is now at ~-2785, indicating the roof collapsed ~20 m in some places. Approximately 0.025 km³ drained from the pond allowing for the collapse.

This collapsed impact melt pond is important because it suggests that the crater was still being modified for a considerable period of time after the main modification stages. For this scenario to develop, the main terraces of the crater had already developed and enough impact...
melt had drained from the walls to fill the 1.5 km² pond. Additionally, the upper surface had cooled considerably and possibly as much as 3 meters or more of crust had developed, as indicated by shadow measurements on the scarps and the wall thickness of a small collapse pit found with the feature.

**Exterior Melt Features**

Exterior melt pools share almost all of the characteristics of interior melt pools (flat surface, mantled boulders, cracks and fractures), and no distinction in the map unit is made. Exterior melt ponds are occasionally found along the crater rim, but are most often found at the distal end of linear channels radial to the rim (Fig. 18). These channels are often kilometers long. The melt forms interconnected pools in depressions adjacent to these channels. The pools cover large areas, periodically broken up by larger ridges, and have a braided or anastomosing appearance.

**Globular Melt Features**

An enigmatic melt texture is seen in a number of areas on the northern wall of the crater (Fig. 20). It has a globular/mamillary, smooth, mounded texture, and occurs on flat areas of terraces and clinging to walls of channels. We call this texture globular melt, because in many instances it looks like an impact melt pond has bubbled up. We interpret this texture to be related to impact melt, but it has distinctive characteristics. Like the melt it has a smooth surface, but these deposits form isolated islands that are commonly tear-drop shaped or irregular clusters. They are often found in association with large impact melt ponds or channels in which impact melt flowed. They are small structures, typically a few meters or tens of meters in diameter, and some appear to have coalesced together. These features are very similar in appearance to irregular mare patches (IMPs, Braden et al., 2014), but are typically smaller and found associated
with melt features and not lava. Globular melt is only observed in the northwest and northern crater walls, and only on the shallower upper third of the crater wall above a steep break in slope. This type of melt is often associated with smooth, mantled walls in the north, interpreted to be melt veneer.

Our interpretation is that the globular texture results from remnants of melt that remain after a larger mass of melt has flowed downhill, similar to the adhesion of water puddles on a waxed car. As a low viscosity melt flows downhill, small amounts adhere to the walls of channels or remain as cohesive blobs of melt on smooth areas. An excellent example of the blobs, which are smooth, raised mounds is shown in Fig. 20. That globular melt is only preserved on shallow slopes above steeper drop-offs supports our interpretation that the viscosity of the material played a role. The material was fluid enough to flow downhill and form leveed channels and flows, but still viscous enough that blobs of material could be left behind. The features do not resemble boulders mantled by melt. If the mounds were simply covered rocks, we would expect to see some indication of a buried block, either an outline, or a fraction of the block sticking out of the mound, neither of which has yet been observed. Areas we now interpret to be globular melt were also imaged by Lunar Orbiter V and briefly mentioned by Schultz (1976). His interpretation was that the surface appeared blistered, and suggested that they represent trapped volatiles, buried blocks, or spatter remnants from molten secondary craters.

Melt Flows

Dark Flows

Large dark flows of impact melt can be seen in a few areas around the crater, with the largest single flows occurring in the northwest and southwest regions, on the elevated plateau. The largest flow by area and volume can be found northwest of the crater rim (Fig. 21). The
flow covers an area of approximately 16.5 km² and ends in a lobate scarp at its distal edge. The flow is relatively thin, with a maximum thickness of ~15 m, grading to ~1 m thick at its distal edge, based on shadow measurements, with an approximate volume of 0.13 km³ (assuming 7.5 m average thickness). It also appears to have flowed over previous flow-like ejecta deposits. A small window of underlying material can also be seen in the northwestern part of the flow. The flow contains many small fractures, which are typical of impact melt pools and melt veneer. Parts of the flow appear to have experienced multiple phases of activity, evidenced by lobate flows on top of the main flow. Additionally, some small craters are observed within the flow that are oblate and appear to have been deformed by relaxation of the melt after the crater formed.

The flow feature appears to emanate from a series of theater-headed sources along the hummocky rim deposits, as well as being sourced from a network of channels of ponded melt from topographically higher positions on the rim. This flow was previously documented in Lunar Orbiter V images (e.g., Strom and Fielder, 1970; Guest, 1973; Schultz, 1976), and Guest (1973) suggested that it appeared the material was being sourced from within the hummocky terrain at the rim of the crater. Guest (1973) inferred that the melt was located within a portion of the overturned flap, and that subsequent impacts were responsible for breaching the area and letting the material spill out. Strom and Fielder (1970) used this feature as evidence for multiphase volcanism in the region, and interpreted the flow as lava. Our observations conflict with the interpretation that the material was sourced from within the hummocky terrain at the rim. We suggest that the flow was sourced entirely from melt ponds on top of the hummocky terrain.

*Flow Crossing Ray*

Images from the LRO NAC show the mobility and longevity of impact melt flows (Fig. 22) and allow us to infer the stratigraphic timing of emplacement. Melt flow features extend
more than 14 km from the rim to a series of channels and ponds at the edge of the continuous ejecta blanket and continuing into the hillocks of the distal ejecta. Much of the flow appears as sheet flow or fast moving radial flow, with melt ponding in local depressions. A second melt feature, beginning in a small catchment area on the rim and flowing ~6 km, ending in a lobate front, formed after the longer sheet-like flow. This later, lobate flow resembles a debris flow, with a channelized region between the catchment and lobe, and crosses a bright ray of material. This example provides evidence that melt remained mobile for a relatively long period after the ejecta blanket was emplaced and that small catchments of melt can coalesce to form more viscous lobate flows. That the lobate flow also crosses a ray of bright boulders also indicates the late arriving timing though its stratigraphic position. Figure 22b shows a close-up view of the flow and rays of both bright and dark boulder streams.

*Splatter Flows*

Flows which appear to emanate from small, irregular depressions are observed almost exclusively on the southwestern ejecta blanket. They are typically found a few kilometers from the rim and have a ropey appearance, which is sometimes seen in a herringbone shape (Fig. 23). Strom and Fielder (1970) suggested that these features result from flank eruptions of lava from vents. Schultz (1976) tentatively suggested that these and other similar flows are the result of trapped subsurface molten ejecta, possibly released by late arriving secondary craters. We refer to these features as splatter flows because it appears that late-arriving molten ejecta material impacted the already-formed ejecta blanket, leaving a small depression and flowing radially away from the crater. This hypothesis was also put forth by Schultz (1976).
**Melt Veneer**

Impact melt veneer is difficult to map clearly, as it is ubiquitous within the crater walls and ejecta blanket. Nearly every surface of the crater walls have the appearance of a thin coating of a melt veneer, and areas that are not covered by clear channelized melt usually have the appearance of sheet flows. Areas that do not have a strong appearance of veneer are limited to the terrace wall scarps and structural surfaces that formed after melt was emplaced. Impact melt veneer was mapped based on the presence of a distinct morphology of small impact crater that is only found on impact melt ponds and surfaces with impact melt veneers. These small craters are shallow, irregular circular features with raised rims, like regular simple craters, but lack ejecta and often have a small mount in the center of the crater floor. They resemble impact into molten material (Oberbeck and Quaide, 1967) and may be related to late-arriving impacts of self-secondary ejecta fragments. Similar features are referred to by Schultz (1976) as pan craters and also resemble bench craters. Bench craters are often found in units with a thin layer of regolith overlying a denser substrate, leave a shelf or bench on the crater wall. Pan craters are similar in appearance to those shown in Fig. 24b, and are found on the King Crater ejecta blanket melt sheet in Al-Tusi Crater (Schultz, 1976). The small irregular craters as shown in Fig. 24b are only found on impact melt surfaces, including impact melt veneer. These irregular craters are also found together with normal appearing craters of the same size, and smaller. The formation of normal appearing craters suggests that the melt units can form traditional crater morphologies and that the formation of the irregular craters could be related to the properties of the melt unit when they formed, i.e., the melt had not yet completely crystallized. Indeed, the small mounds or boulders within the irregular craters may be the bolide that formed the irregular depression. We have used craters with this morphology to map the extent of the melt veneer in
the ejecta blanket (Fig. 22a). However, these features are very small, and only clearly identifiable in NAC images with the highest resolution (< 0.5 m/pixel). As such, there are gaps in coverage for mapping the extent of these features, and thus melt veneer. Regardless, the distribution of melt veneer identifiable with these craters appears to only extend as far as the edge of the continuous ejecta blanket and the Melt-rich Channels and Lobes unit of our map.

**Melt Area and Volume**

Ponded melt and other melt features, including dark flows and other types of melt, have a combined surface area of ~118 km². Without knowing the pre-deposition geometry of the depressions that are infilled, accurate estimates of melt volume are difficult. The largest flows have thicknesses of ~10 – 15 m, and the thickest interior terrace melt deposits are >15–20 m thick. Exterior ponds that embay the concentric-ridged terrain are only 1–2 m thick. More distal ponds that have pooled at the base of the ejecta slope on in the east are on the order of 5–25 m deep. If we assume an average melt thickness of 10 m, regardless of location, we suspect that ~1.2 km³ of melt has accumulated in ponds and obvious flows around the crater. This estimate does not include the smooth floor, which is the main mass and volume of melt, nor does it include obvious melt features on the crater walls and ejecta in the Channeled and Veneered walls and Melt-rich Channels and Lobes ejecta.

### 2.3.2.5 Boulders

Surrounding the rim of Aristarchus Crater are numerous groups of blocky ejected material containing boulders of various sizes. The ejecta blocks form chains that can be long (e.g., 5 km) or short (e.g., several hundred m), but are contained completely within the proximal ejecta blanket. Boulders are strewn all over the proximal ejecta blanket, but the areas that have been mapped are composed of concentrations of boulders that appear as elongated chains radial to the
crater center. These concentrations of boulders are easily identified, as the surrounding terrain is relatively boulder free. Some areas containing concentrated boulder chains are also found in somewhat less developed boulder fields, which are also related to the impact ejecta. The Boulder Field unit represents areas of contiguous and adjoining blocks of ejecta. We have not included boulders that appear to be related to weathering of mounds or formed by fracturing of melt units, nor as part of ejecta of craters that post-date Aristarchus.

Our intention was to map ejecta blocks to determine the distribution of blocks that were emplaced ballistically. Nearly all of the boulder fields and streams mapped form radial chains on the crater walls and in the ejecta blanket. They are typically emplaced on top of the surrounding ejecta blanket, but some streams show evidence for radial flow and entrainment of melt and surface debris. With few exceptions the boulder fields are found within ~10 km of the rim, and the length of the chains are typically 1–2 km, but can be less than a few hundred m to more than 5 km, and are rarely more than a few hundred m across. Few blocks larger than ~100 m diameter are found within the blocky chains. Most of the largest blocks in the chains appear to be fractured, and it is likely that the lack of very large blocks is due to break up during the ejection stages or on impact during emplacement.

Stratified Blocks

Owing to the high-resolution of the NAC cameras, fine details can be seen on large boulders in and around the crater. One of the most significant is the observation of stratified ejecta blocks (Figs. 25 and 26). The blocks range in size from a few meters to more than 150 m in diameter. The strata alternate between dark-toned (less reflective) and light-toned (more reflective) layers, where the bright layers are typically thicker than the dark layers. Bright layers vary in thickness but are most commonly 2–3 m thick, with a few layers up to 10 m thick (Fig
Dark layers may more accurately be described as banding, and are less than 1.5 m thick in all observed cases. The blocks are randomly oriented on the northern and northeastern crater wall, and appear to have moved down slope from the crater rim; however, a few small stratified blocks are observed on the outer rim sloping away from the center. The blocks appear to both lie on top of and within impact melt on the crater wall, and in some cases appear to have been transported down slope by the flow of melt. The boulder field is approximately 2.8 km downslope from the first major crater wall terrace, and five kilometers from the main crater rim. Stratified blocks are only clearly identifiable in <0.5 m/pixel coverage and under low incidence illumination, and as such, due to limitations in NAC coverage with favorable conditions, it is possible that we have not identified all stratified boulders within the crater. However, adequate coverage exists to establish that nearly all of the stratified boulders are found in the northeastern part of the crater wall.

Figure 26 shows one of the best examples of the stratified blocks on the Aristarchus Crater wall. The block measures 95 m in length by 65 m at its widest point, perpendicular to bedding, and is at least 35 m thick based on shadow measurements of its height. It contains at least 20 individual layers of bright material, separated by thinner bands of dark material. Bright layer thickness varies between one meter and six meters, and dark layers have a maximum thickness of ~ 1 m. The layers appear slightly wavy, and the block appears to be covered by some dusty mantling.

2.3.2.6 Structural Units

The mapped locations of major structural units are limited to terrace faults, prominent leveed channels and melt conduits, and floor fractures. Terrace faults are generally inferred by the presence of a scarp, where the fault is mapped on the down-dropped block at the base of the
scarp. Sinuous and curvilinear channels are mapped primarily within the Channeled and Veneered Walls, and to give radial motion context to melt flows in the ejecta blanket units. Floor fractures are contained entirely within the smooth floor deposits and are often circumferential to the crater walls. Floor fractures near the walls likely indicate a “bath tub ring” and represent the drainage of melt after the top of the melt sheet cooled, leaving extensional fractures. Many fractures are hundreds of meters wide and can be interpreted as pits (e.g., Wagner et al., 2014).

2.3.3 Geologic Map from Image Classification

The geologic map created using image classification techniques of the WAC color can be broken down into ten major units, eight of which are found associated with the excavation of material from Aristarchus, and two of which represent pre-impact surfaces.

The central peak (unit CP) is the only area in the map designated in red. As indicated in the principal component image discussion, this area has the largest variance, and also has the lowest FeO concentration (<4 wt%), and a short CF position (7.7 µm). Unit CF (Crater Floor; dark pink color) surrounds the central peak and comprises most of the crater floor and areas on the southwest wall, with FeO content <9 wt% and CF positions typically much shorter than 7.9. Morphologically, hummocky floor units are well correlated with very low (<7.7 µm) CF positions in the unit. Unit SWE (Southwest Ejecta; light pink color) makes up much of the southwestern bright ejecta ray and the eastern high-albedo ejecta, and is highly correlated with boulder fields in the eastern ejecta and with rocky areas of the scarp and bright ray. FeO contents in Unit SWE are less than 9 wt%, with short CF positions (<7.6 µm) and high Th contents. Impact melt features (splatter flows) are also found associated with Unit SWE.

Two units compose the crater walls. Unit WW (West Wall; yellow color) is highly correlated with boulder fields and wall units in the west and southeast, and corresponds to areas
in yellow in the PCA image. Unit WW has intermediate FeO contents (8.5-13 wt%) relative to the other units in this map, and has CF positions close to 8 µm. Unit NW (North Wall; dark green color) occurs on the northeastern wall, and much of the northern ejecta blanket, and is characterized by high FeO content (14-17 wt%) and CF positions indicative of pyroxene (8.15 – 8.25 µm) and/or olivine. There is a sharp contact between units WW and NW in the eastern wall. Unit NW also makes up a more diffuse unit of the ejecta blanket on the plateau and areas in the SW not dominated by Unit SWE.

The ejecta blanket is comprised of 5 major units. Units SWE, in the southwest and eastern ejecta, and Unit NW, predominantly in the north, are discussed in preceding sections. Unit WE (West Ejecta) occurs exclusively on the plateau in the distal northern and western ejecta blanket, and is highly correlated with secondary crater fields. Unit WE has a relatively high FeO content (14-17 wt%) and mafic CF positions (8.1 – 8.195 µm). Unit SE (Southeast Ejecta) occurs in the southeastern ejecta and upper crater wall, and corresponds to olivine-rich areas identified by previous researchers (e.g. McEwen et al., 1994, LeMouélic et al., 1999). The FeO content of Unit SE is 13.5 – 14.5 wt %, and CF position is 8.125 – 8.225 µm. Although typical olivine-rich CF positions are greater than 8.5 µm, it is clear from M3 (Mustard et al., 2011; Isaacson et al., 2011) and Clementine UVVIS spectra (McEwen et al., 1994; LeMouelic et al., 1999; Chevr nel et al., 2009) that this area is olivine-rich, with a minor pyroxene component. Unit NE (Northeast Ejecta) is a large dark region in the northeastern ejecta blanket, with a predominantly mafic composition with high FeO content (16 – 18 wt%) and CF positions greater than 8.2 µm.

The pre-impact surfaces are the bright-blue colored unit WP (Western Plateau), which represents the current surface of the plateau, and the light blue unit SM (South Mare), which represents the current mare surface. As commented on by a number of previous researchers, the
plateau surface is dominated by > 30 m thick dark mantling deposits, overlying either basalt flows, Imbrium ejecta, or ancient highlands crust (e.g., Zisk et al., 1977; Gaddis et al., 1985; McEwen et al., 1994). The blue colored unit WP has high FeO content (>16 wt %), and high CF positions (> 8.20 µm) indicating it is a mafic unit, dominated by Px-rich components (Chevrel et al., 2009). These values are consistent with a dark mantling deposit and the general volcanic nature of the plateau. Unit SM (light blue) occurs predominantly in the East and South of the map, with high FeO content (>15 wt%), high CF positions (> 8.20 µm), and has the smooth, flat appearance of mare flood basalts.

2.4 Discussion

We use the following sections to discuss the geomorphologic map (Section 2.4) and the general geology of units excavated by Aristarchus Crater (Section 2.5). In section 4 we discuss our geomorphologic mapping, and suggest that the pre-impact topography has played a role in the morphometry of the crater (Section 2.4.1) and the distribution of ejected materials (Section 2.4.2). The morphology of the central peak is briefly discussed in Section 2.4.3. The observation of stratified blocks was first made at Aristarchus, and we discuss possible modes of formation for these boulders in Section 2.4.4. A surprising finding from our geomorphologic mapping is that there is little correlation between morphology and composition, which we discuss further in Section 2.4.4.

2.4.1 Influence of pre-existing topography

Effects of the pre-existing topography on the Aristarchus cratering event have been discussed in a number of previous works. In their morphologic mapping, Guest (1973) concluded that hummocky ejecta on the plateau side of the crater was likely related to the deposition of plateau materials there, whereas ejecta units on the mare side reflected deposition of mare units,
an hypothesis supported by mapping of Guest and Spudis (1985). Our mapping and observations confirm a difference in the nature of ejecta morphologies between the two elevations. We further suggest that differences in topography and in bedrock density between the units are also reflected in the distribution of impactites (melt and boulder fields) and in asymmetries in crater morphometry.

Topographic cross-sections (Fig. 5) show the difference in morphology of the crater rim between the western and eastern units. The plateau side (mainly to the north of the crater) has a poorly defined rim and the ejecta blanket appears nearly flat. For the most part melt on the plateau appears as a thick veneer, and there are few well-developed melt ponds. In areas where melt was mobilized, as in the large dark flows in the NE and in the SW (on the hanging wall of the plateau scarp), there were no steep slopes for the melt to flow down, so large sheet flows developed. Melt flows above the scarp in the southwest fed narrower sinuous channelized flows that ran down the scarp, and the steep topography led to a different melt morphology. In the east, the steeper slopes on the ejecta blanket led to channelized flows sourced from small catchments on the rim and incising the concentric ridged unit. Numerous melt ponds formed at the break in slope at the base of the ejecta.

The interior walls of the crater also display different regional slopes, where the plateau wall is steeper than the mare wall (~17° vs. ~14°, respectively), and measurable effects of the different bedrock target materials can be estimated from terrace widths. The width of terraces, particularly first order terraces (those closest to the final crater rim), can be approximated by calculating the overburden pressure required to match the bedrock shear strength in a simple geometry (Melosh, 1977) and can be approximated by $W = \frac{c}{\rho g} \left(\frac{1+16\lambda^2}{16 \lambda^2}\right)$, where $W$ is the predicted width of the terrace, $c$ is the bedrock cohesion (shear strength), $\rho$ is the bedrock
density, \( g \) is the gravitational acceleration, and \( \lambda \) is the depth diameter ratio of the transient crater (Pearce and Melosh, 1986). Following this model, which neglects inertial forces, the difference in terrace width between the plateau and mare sides of the crater can be reproduced by assuming different material properties for cohesion and bedrock density that reflect the assumed properties of the target. The first terrace widths (~1.3 km in the east, ~2.0 km in the west) can be reproduced if we solve the above equation and change the bedrock density to reflect ~3300 kg/m\(^3\) for mare materials in the eastern terraces, and ~2500 kg/m\(^3\) for highlands-like materials in the western terraces, and assume the same shear strength for both materials (15 bar). Although shear strength is likely to be different between mare and highlands materials, this parameter varies by less than a factor of 2, and variable cohesion does not need to be invoked to explain the variation in terrace widths (Pearce and Melosh, 1986). In the case of Aristarchus, the variation in terrace width between the mare and plateau sides can be explained by a change in bedrock density.

### 2.4.2 Impactite distribution

An important result of the geomorphologic mapping is the recognition of a difference in the distribution of impact melt ponds and boulder fields between the plateau and mare regions of the crater. Few boulder fields are found on the high plateau ejecta, whereas many occur on the ejecta blanket of the lower elevation mare side. Conversely, many boulder fields occur on the western plateau wall within the crater, whereas few boulder fields are found on the mare-side wall. This asymmetric distribution of materials is also seen in impact melt ponds. We suggest that the boulder field and impact melt distribution dichotomies could be related to ejection dynamics. The boulder fields are sourced from relatively shallow depths, as inferred by their location proximal to the crater rim, and were likely emplaced at low velocity. Owing to the
elevation difference between the plateau and mare, it is possible that these ejecta streams did not have enough momentum to crest the high plateau wall and instead impacted the wall of the crater. On the mare-side, the ejecta blocks had no impediment during excavation and instead landed within the ejecta blanket. Some boulder fields on the western wall have a tear-drop shape indicating they fell down the wall, but many others appear to remain where they were emplaced. The superposition of boulders on the wall would indicate that the boulder fields were emplaced before the formation of the terraces, and provides evidence that terrace formation began after most of the ejecta blanket was emplaced.

The target materials at Aristarchus are spatially inhomogeneously distributed, and it is possible that some lithologies were preferentially incorporated into different parts of the impact melt and ejecta stream such that the pre-impact distribution is now reflected in the ejected distribution. This situation differs from that commonly observed or inferred for impact craters, which usually produce very well mixed and homogenized impact melt. Studies of terrestrial impact melt rock distribution and composition generally show that rock compositions are homogenized by the impact melting and excavation processes, and that final melt compositions from different areas around a crater tend to have very similar melt compositions (Grieve 1975; Dressler and Reimold, 2001). The well-mixed character of impact melt is a result of superheating on the order of 1500 – 2500 °C (Grieve et al., 1977), with very low viscosity, and vigorous mixing followed by rapid cooling (Phinney and Simonds, 1977). However, a detailed study of the melt rocks preserved at Mistastin Lake Crater (Marion and Sylvester, 2010) showed that heterogeneities can exist at microscopic and outcrop scales; and although the heterogeneities are slight (most outcrops have very similar bulk compositions), matrix and bulk compositions can reflect different proportions of the target rocks. Dhingra et al. (2013) used M³ data to show
evidence for compositional heterogeneity in impact melt flows in Copernicus crater (e.g., suggesting that impact melts might not be as well mixed as previously thought.

2.4.3 **Central peak Characteristics**

On the basis of morphology and relative albedo differences, we infer that the central peak is probably not composed of three distinct materials or rock types (Fig. 15), but rather the albedo differences result from differences in mass wasting, revealing true central peak materials from beneath a thin veneer of melt. The northwestern portion of the central peak, which appears to be covered in melt, suggests that possibly the entire peak was covered at one time, and most of this material has slid off owing to steeper slopes. The contact between the northwestern melt-covered area and the bright northern patch is littered with boulders, indicating that the melt cover is being weathered off the slope. The gray middle portion is still covered by many smaller boulders, which might explain its slightly darker albedo.

2.4.4 **Stratified Boulder Observations and Formation**

Stratified blocks are observed on the northeastern wall and range in size up to 150 m in diameter. The stratified blocks observed on the northeastern wall clearly show alternating bands of light and dark albedo material in NAC imagery (Figs. 25 and 26); however, the origin of the banding remains unclear, and may be explained by a number of different geologic processes. Possible origins for the stratified materials and the repetition of dark and bright albedo layers have been proposed by Zanetti et al. (2011) and by the Kickapoo Research Team and Kramer (2014) : 1) The banding represents layers of basaltic lava (higher albedo layers) that are capped by periodic pyroclastic dark mantle deposit (DMD) eruptions; 2) the dark layers represent the development of a regolith layer on top of basalt flows; 3) the dark layers represent the formation of a glassy rind or vesiculated crust formed by cooling of the surface in a vacuum; (models 1-3
from Zanetti et al., 2012); 4) the dark and bright regions represent a sequence of alternating felsic and mafic materials from a layered mafic intrusion (Kramer et al., 2014).

As suggested by Zanetti et al. (2011), the dark layer could be related to the deposition between basalt layers of some interstitial material, either pyroclastic material (given the proximity of the crater to known dark-mantle pyroclastic material on Aristarchus Plateau) or by the regular development of a meter or so of regolith between each successive lava flow. The development of a regolith 1 m thick between each episode of lava emplacement can be ruled out by insufficient time. Regolith development rates are extremely slow, on the order of 1 m/10^9 years (Melosh, 1989), and there is simply not enough time in which to develop interleaved bands such as those observed in figures 25 and 26. Regolith growth would also not explain the uniformity of the dark layer thicknesses (1-3 m?). Pyroclastic deposition offers an interesting possibility, but has drawbacks. The deposition of a dark mantle material neatly explains the occurrence of the albedo variation between the layers, but it is not expected that meters-thick pyroclastic deposition would occur after every successive lava flow. It is unlikely that pyroclastic deposition would uniformly thick deposits of material after every successive lava event. Moreover, it is difficult to envision that each of the depositional events would be of a uniform thickness, such as what is observed.

Considering the geology of Aristarchus, it is not unreasonable to postulate that the stratified blocks could be remnants of a shallow layered mafic intrusion that was impacted and excavated by the Aristarchus event, as suggested by Pieters (personal communication, 2012; 2013) and further proposed by Kramer et al. (2014). The Kickapoo students^2 measured albedo

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^2 M. Zanetti originally advised the Kickapoo High School students on the three lava flow emplacement scenarios (and suggested a possible LMI, to their advisor Richard Snyder, in a phone conversation in Feb. 2012) prior to their official involvement with the LPI and G. Kramer. MZ had no involvement with the measurements or conclusions presented in Kramer et al. (2014).
differences between bright and dark layers for 29 blocks and determined that average bright values are on the order of 0.45 and dark values average ~0.32, and both tend to fall intermediate between highlands material and basalt (Kramer et al., 2014). We have been unable to reproduce their measured albedo values, even within the same image and on the same blocks. Using photometrically calibrated NAC imagery, we measured average bright region pixel I/F values of 0.11, with maximum, sun-facing slope values of ~0.13. Dark layer I/F values average 0.09. For comparison, the darkest (least reflective) melt ponds in the same NAC frame have I/F values of ~0.07 (shadows are 0.02); and the brightest (most reflective) areas of the central peak, one of the brightest features on the Moon, are ~0.4. Measurements by Kramer et al., (2014) appear to over-estimate the albedo of the surface by a factor of 3, and possibly indicate a significant error in data preparation.

In addition to their work at Aristarchus, Kramer et al. (2014) also identified stratified blocks in Mare Undarum and posited that those features were also derived from layered mafic intrusions. The Mare Undarum stratified blocks, however, occur in 300–1000 m diameter craters in a mare-basalt target, indicating they were excavated from less than ~100 m depths. It is unlikely that the stratified blocks there could have sampled a shallow pluton, considering the mare thickness based on buried ghost craters is ~300 m. Furthermore, stratified blocks have also been reported in other strictly mare-target crater ejecta at Rumker E, Bose, and unnamed craters in Mare Australe (Julie Stopar, personal communication, July, 27, 2012).

Another hypothesis for the stratified boulders, in addition to the ones listed above, is that the dark and light banding results from textural differences in basalt flows, where the dark regions represent chunkier, A’a-textured surfaces surrounding a core of higher albedo, crystalline basalt. Although this hypothesis is similar to the first 3 models involving deposition of
successive lava flows, we base this interpretation on observations of terrestrial lava flows (e.g., Lockwood and Hazlet, 2010). Basaltic lava flows like those on Mauna Kea, HI, and near Sunset Crater, AZ, (Fig. 29) have fragmental, rubbly basalt textures above and below a crystalline central core. As an ‘a’a flow advances, it buries its own surface rubble, leaving a blocky zone both above and below the core. The different textures also have a different color and albedo, as the rough ‘a’a is darker than the lighter crystalline core. Figure 29 shows the Bonito flow in Sunset Crater National Monument, Flagstaff, AZ, where two, 1.5 – 2 m thick, bright-albedo, crystalline basalt cores are seen separated by ~1 – 1.5 m of darker albedo ‘a’a textured material. The vertical dimensions of the Bonito Flow units are very similar to what we observe in the stratified blocks at Aristarchus. The ‘a’a texture model can also explain the differential weathering observed between the layers, where the darker, possibly more friable ‘a’a material could be more easily removed, resulting in the ladder step appearance seen on the margin of the block in Fig 26.

Suspected lava flow layering, with similar thickness and albedo variation, is observed in Kepler Crater (Ohman et al., 2012), and within the walls of Vallis Schröteri close to the margin of the plateau (Garry et al., 2010). Further evidence for an origin as lava flows can be inferred from the spatial distribution of stratified blocks within the crater. Nearly all of the stratified blocks are observed in the northeastern wall, which is likely to have been a shallow mare unit at the margins of the plateau prior to the impact. Their position high on the wall and very near the rim indicates that they represent very shallow excavation depths, and that the blocks have tumbled downhill during late stages of crater modification.
2.4.5 **Morphology vs. Composition**

One of the major goals of creating the geomorphologic map was to identify compositional dependencies for specific morphologies. A surprising result from comparisons of composition with morphologic units is the apparent lack of a uniform impact melt signature in compositional datasets, such as $M^3$ and Clementine FeO, as well as in WAC color ratio images. For example, the large dark melt flow in the NW is compositionally indistinguishable in WAC color and $M^3$ parameter maps from the ejecta surrounding it, as are radial melt flows in the northeast and most radial melt channels.

Areas where morphology does correspond to specific compositions are limited. The crater floor and the central peak are well correlated, with Low FeO ($< 9$ wt%) and short CF positions ($<7.8$ µm). Some areas of impact melt, as suggested by Mustard et al. (2011, 2012), can be correlated with both short CF position, for example splatter flows in the SW, and with olivine-rich areas in the SE. However, we observe that these relationships are very rare, and may be coincidental.

The morphology that most often corresponds with specific compositional information is our boulder field unit. Frequently, in areas where spectral datasets can resolve boulder fields, their compositional character can be discerned. Even with 400 m/pixel resolution, the color of large boulder fields can be discriminated and related to a specific composition through comparison with other datasets. Fresh appearing boulders dominate areas that have low FeO and short CF positions; such as hummocky floor unit, where boulders cap most of the mounds, and the bright eastern ejecta tongue, where a high concentration of boulders can be well correlated. Individual boulder fields of different compositions can be distinguished from one another, even in very close proximity. In the SE, bright boulder fields, with low FeO and short CF positions are
within a few km of a dark boulder field with olivine-rich compositions (Fig. 22). We discuss additional examples in subsequent sections.

2.5  Geologic Map Discussion:

In Section 2.5 we discuss the results of our PCA image classification mapping, and our geologic map. A description of each of the ten units in the geologic map is presented, focusing on spatial distribution and correlation with compositional data. For each unit, we attempt to classify the rock type exposed based on available compositional data and relative position within the ejecta. We emphasize that our results are based on remotely-sensed data with a wide-range of resolutions, and on the informed interpretations of previous researchers, but are reasoned speculation. Based on the image classification geologic mapping, we suggest that Aristarchus may have excavated a shallow, silica-rich pluton, and that other areas on the plateau, notably the Cobra Head volcano, may be related to silicic volcanism. We devise a new scenario for the evolution of the southern Aristarchus plateau based on these inferences.

Principal component analysis of WAC color bands and the wealth of compositional data and previous research show that the materials excavated by Aristarchus are compositionally diverse, and that the geologic history of the area is complicated. The following sections discuss the units of the geologic map from image classification of the WAC-color PCA analysis (Geologic Map: Fig. 30; Legend: Table 3). We attempt to narrow down possible rock types based on the spatial distribution of materials, any dominant morphologic character, and compositional information from FeO and CF position (Figs.32 and 33), and Th abundance (Fig. 34). Mapping and observations were informed by the image classification of the PCA, analysis of WAC color-ratio imagery, and knowledge of crater morphometry. Suspected super-position of units in the pre-impact target is inferred from position of ejecta units relative to the crater rim,
with near-rim units originating from a deeper source and emplaced as low velocity ejecta (Shoemaker et al., 1963; Oberbeck and Quaide, 1968). We also use interpretations from previous spectral studies to constrain mineral content (e.g. Chevrel et al., 2009; Mustard et al., 2011) to further narrow the possible rock types excavated by the crater. We emphasize that our interpretations are inferences about the possible rock types and their sources based on available data. Most of the units mapped in Fig 30 represent mixtures of ejecta units with one another, and mixtures with surface material where the ejecta landed. As such, the units defined in the geologic map are likely more diffuse than indicated. For example, although the crater floor is mapped entirely as unit CF, much of the smooth floor is clearly impact melt, and our rock type interpretation applies to hummocky mounds with clear boulders that are not impact melt. Although it is likely that a significant component of the “bedrock” mounds was melted and formed the melt sheet, and the melt may be compositionally very similar, the simplified map units do not discriminate this fact. When appropriate, we call attention to the areas used to define the unit through image classification. Despite these caveats, we can deconvolve useful information about the possible lithologies, the current distribution, and probable subsurface provenance.

2.5.1 Central Peak, Crater Floor, and SW Ejecta (Units CP, CF, SWE)

Although it is spectrally distinct, the central peak is compositionally similar to areas of the crater floor and areas of the East and Southwest ejecta. Units CP, CF, and SWE are likely related to the same assemblage of rocks that was excavated and dispersed asymmetrically in the ejecta blanket. They all share the same compositional characteristics, with low FeO contents, short CF positions, and high Th content. Additionally, all three units have generally high albedo.
Unit CP is dominated by a very bright region on the central peak (see Fig. 15), which likely is the bedrock surface exposed beneath a thin, exfoliating melt veneer. The approximate upper limit for the CF position of the central peak is 7.72 µm (Glotch et al., 2010; Song et al., 2013). Hills and mounds in the hummocky floor unit share a similarly high albedo and can be directly correlated with low FeO regions (Figs. 32a, 33a) and short CF position “white-spots” (Figs. 32b, 33b). The ejecta blanket to the east is dominated by high albedo boulder fields, and although more diffuse, the southwest ejecta ray is also higher in albedo than surrounding terrain. Although the southwestern ray is probably composed primarily of fine-grained material, large boulders resolvable in NAC imagery (i.e., > 3 m) are abundant and likely contribute to the spectral characteristics of the unit. Individual “white-spots” of very low CF position can be well correlated with diffuse boulder fields (Fig. 33b).

Previous interpretations for these units indicate they are feldspar-rich. Chevrel et al. (2009) propose a dominant anorthositic component with minor contributions of clinopyroxene and olivine, interpretations based on telescopic investigations by Lucey et al. (1986). Tompkins and Pieters (1999) regarded the central peak as difficult to classify and grouped it into their gabbroic-noritic-troctolitic-anorthosite (GNTA1, 2) scheme. Ohtake et al. (2009) listed material from the crater floor as PAN (>98% anorthosite) and claimed to have identified a prominent 1.25 µm feature. Most recently, M^3 data indicates no Fe-absorptions, likely indicating a feldspar-rich rock sourced from the upper plagioclase-rich lunar crust (Mustard et al., 2011).

Based on the low FeO content (< 9 wt%), high albedo, and lack of mafic signatures in M^3 spectra, it is likely that feldspar is a dominant component in the rocks that make up CP, CF, and SWE. However, we ask: what is the character of the feldspar? Are the exposed rocks typical Ca-rich plagioclase, common to the lunar highlands, or is there evidence that they are more alkaline?
In the following sections we make the case that the rocks of units CP, CF, and SWE may be evidence of highly evolved, alkalic (or possibly granitic) material. Although similar suggestions have been presented in the past (e.g., Jolliff, 2000; Hagerty et al., 2009; Glotch et al., 2010), we present the strongest evidence to date, using multiple compositional datasets, and show the spatial extent of the materials.

2.5.1.1 Low FeO – High Silica – High Th Correlation

Areas with <10 wt % FeO, < 8.0 µm CF positions, and > 11 ppm Th content are very highly correlated. Areas with the shortest CF positions directly correspond with areas of low FeO content (Fig. 36). A plot of FeO concentration versus CF position is shown in Fig. 36a, fit with a moving average trendline (which smooths variation by averaging the next 50 values on a point by point basis) to show the general trend more clearly. Areas of this graph with FeO values less than 11 wt % and CF positions shorter than 8 µm is shown in Fig. 36b. These areas are directly correlated with the CP, CF, and SWE regions in the geologic map. A similar correlation exists with FeO content and Th concentration, albeit with a much lower spatial resolution data set (Fig. 37). Plotting Th abundance versus FeO content produces a graph with a distinctive tail of areas with lower FeO content and high Th concentration (>11 ppm) (Fig. 37b). The spatial distribution of this material also covers the same areas as FeO and CF positions in Fig. 36b and the map units CP, CF, and SWE. The strong correlation of compositions with these units provides good evidence that they are genetically linked.

2.5.1.2 Correlation with \( M^3 \) spectrally indistinct unit

There is nearly 1:1 correlation of spectrally indistinct units in \( M^3 \) with silica-rich regions (Mustard et al., 2012), and we would add that this is also true for areas of low FeO content. The interpretation by Mustard et al. (2011, 2012) is that the spectrally indistinct areas are likely
impact melt glass, and they point out that these areas are well correlated with impact melt features (e.g., splatter flows and large area flows in the SW of our map). While this correlation is true, and glass material is expected to be a component of impact melt and the ejecta blanket, we suggest an alternative explanation for the correlation of M³ parameter mapping with CF position and FeO content. The most strongly correlated with areas of lowest CF position and regions of the lowest average FeO content in the southwest are boulder fields. Additionally, areas of the central peak and hummocky crater floor, particularly mounds capped by boulder debris, share a similar correlation between the three datasets. If the boulder fields and debris-covered hills indicate exposed low FeO, silica-rich rocks, then the interpretation of these regions as dominated by glassy impact melt cannot be entirely correct. The distribution of these materials along a bright ray to the southwest could indicate that they are fine-grained particulate matter, as opposed to glassy material. We speculate that the spectrally indistinct material on the crater floor and in the southwestern and eastern ejecta is better explained as a compositional lithology that is not identifiable in M³ spectra. Spectrally indistinct minerals in the near-IR, such as quartz and K-feldspar, both of which may be present in low FeO, silica-rich materials, offer a possible alternative.

2.5.1.3 Possible Rock Types

The short CF positions in these areas indicate materials that either contain quartz, silica-rich glass, or alkali feldspar (Glotch et al., 2010). The CF positions measured in the CP, CF, and SWE units fall between 7.5 and 7.9 µm, which may, based on laboratory CF position measurements, indicate the presence of sodic minerals, such as labradorite or albite (Fig. 35 from Glotch et al., 2010). In comparison with detailed analyses of returned samples, primarily from Apollo 12, the CP, CF, and SWE units fall generally in the range of KREEP basalts, with the
lowest FeO and highest Th areas (red areas in Fig. 37) trending toward alkali anorthosite compositions (Fig. 37b). Although the dataset is very low resolution, Lunar Prospector GRS potassium maps indicate that the Aristarchus Crater and areas slightly to the west (SWE) have elevated K concentrations, on the order of 3300 ppm (and among the highest K concentrations on the Moon; Gillis et al. (2002); not pictured).

Taken at face value, these compositions indicate an alkali anorthosite composition containing significant (in relation to lunar rocks) K-feldspar. The relative location of CP, CF, and SWE units on a plot of LP-GRS derived Th vs. Clementine FeO shows that they fall within the range of KREEP-rich and trend toward low FeO and high Th alkali-anorthosite (Fig. 37). The very high FeO content and high Th units in the left side of figure 37 suggest a mixing of high Th material with high FeO, which can be interpreted as a simple mixture of CF-like material with local mare basalts. However, linear mixing of a CF-like material and basalt alone cannot explain the elevated proportion of FeO compared to Th-content. The highest FeO, high-Th materials observed at Aristarchus plot along a mixing line between FeO-rich Apollo 12 basalts and Apollo 15 impact glass that trend toward more evolved rocks like monzogabbro and granite (Fig. 37, modified from Jolliff 2004).

All of these rock types are consistent with the major mineralogy observed at Aristarchus. High CPx concentrations (>20%; Shkuratov et al., 2005), and high-Th and –K can all be expected in lunar quartz monzodiorite or monzogabbro (Jolliff et al. 1991; 2004). The lack of strong mafic absorptions in the central peak, and generally unremarkable spectra seen in M3 parameter mapping (Mustard et al., 2001), may be related to silica-rich phases with weak to absent spectral characteristics mixed with basaltic material, rather than their preferred interpretation of impact glass. Owing to the broad spatial response function of the LP-GRS,
which affects its response to both Th and K, and the fact that the ejection process has dispersed materials over a wide area, it is difficult to know their true concentrations. Hagerty et al. (2009) suggested modeled Th abundances for Aristarchus Crater as high as 15 ppm, but it is possible that individual small areas, in particular the central peak and bright hummocky mounds on the floor of the crater may have significantly higher concentrations of these elements. Granitic compositions could be present in small quantities, widely dispersed in the ejecta, and contribute to the Th signal in LP-GRS, but large outcrops are not probable.

2.5.1.4 Highly-evolved source?

The high Th and silica-rich character of the central peak and associated units, and the relatively shallow sampling depth of Aristarchus (4 km for CP, <3 km for CF and SWE) argue against the central peak having sampled an ancient plagioclase flotation crust related to lunar magma ocean (LMO) crystallization buried beneath the mare basalts or plateau (e.g., Thompkins and Pieters, 1999; Ohtake et al., 2009; Mustard et al., 2011). We argue that more evolved lithologies are necessary to explain the observations, and that Aristarchus may have excavated a shallow alkali anorthositic pluton or an alkaline-suite igneous complex consisting of intrusive and extrusive rocks. In order to form relatively evolved rock types on the Moon, a mechanism for generating silica-rich magmas is necessary.

Two possibilities, silicate-liquid immiscibility (SLI) and basaltic underplating, are methods of separating silica-rich material from the parent melt. SLI occurs at very late stages of basaltic magma fractional crystallization (>90%) (Rutherford and Hess, 1975). During the fractional crystallization process, as temperature, pressure, and bulk compositions change, the remaining basaltic liquid becomes increasingly Fe-rich and enters a field of silicate liquid immiscibility. It then separates into two coexisting, immiscible melts (Hess, 1989). One melt is silica-poor and
Fe-rich, and the other is silica-rich and alkali-rich (Hess, 1989). Although SLI has been proposed for the formation of lunar granites at the microscopic scale (e.g., Jolliff, 1991; Seddio et al., 2014), it is not known if significant quantities of granitic, felsic, or alkali-suite rocks can be created through this process (e.g., Jolliff et al., 1998).

Basaltic underplating is a process that involves intrusion of a basaltic magma into the crust, which if it is significantly hotter than the liquidus of the rocks into which it intrudes (or those rocks are already sufficiently hot) can cause melting or partial melting, forming a silicic magma (Hildreth, 1981). On Earth, basaltic underplating has been suspected of playing an important role in the production of rhyolites, through injection of basaltic magma into continental crust, and can produce large amounts of silica-rich melt on short timescales (Hildreth, 1981). The problem with the Moon is that known crustal rock types are essentially anhydrous and have very high melting temperatures. An exception is KREEP-rich material, which would begin to melt around 960-980 °C. Basaltic underplating of a KREEP-rich deposit or crustal concentration could produce a partial melt. In the vicinity of Aristarchus, this scenario may be plausible.

2.5.2 **Crater Wall Units (WW and NW)**

Based on the WAC-Color PCA image classification map, the walls of the crater can be defined by two units, WW and NW. Unit WW was classified based on the distribution of yellow-colored boulders in the WAC PCA image, found on the western wall and large outcrops within the north wall (Fig. 10b). PCA classification determined that areas of the southeast wall are similar to the western wall, although they appear mantled by olivine-rich units (Unit SE) and southwest ejecta units (SWE). Analysis of representative spectra from Unit WW shows that it is the same shape and has the same reflectance as Unit CF, with a slightly lower reflectance in the 321 nm band.
Unit NE within the northeastern crater wall is clearly distinct from other wall units, and can be recognized as a darker albedo unit in WAC mono images, as well as in WAC-color and other spectral datasets. A striking feature of this unit within the crater is a sharp linear contact with Unit WW, which appears to reflect a difference in major lithology, and is not a result of mantling by boulders or melt veneer. Although unit NW within the crater does contain melt channels, the dominant morphology is wall terracing, indicating exposed bedrock. Within the ejecta blanket patches of Unit NE can be seen in the southwest, but the unit is predominantly found in the north, west, and southwest.

2.5.2.1 Possible Rock Types (WW and NW)

The intermediate FeO content (range) and mid-range (~8 µm) CF positions seen in Unit WW indicate a generally feldspathic composition. M^3 spectra of an area on the northwest wall indicate a high-Ca pyroxene component, and have been interpreted as indicative of gabbroic material or mare basalt (Mustard et al., 2011). Telescopic observations also suggest a feldspar and high-Ca pyroxene component (Lucey et al., 1986). The nearly identical WAC color spectra of unit WW and unit CF indicates they are very similar, and based on previous observations of clinopyroxene, we speculate the differences between these units could be related to a CPx component in unit WW. The observation of these materials in the crater walls suggest a large exposure of the parent rock type, and could represent plateau material. If these are plateau materials, then the rocks exposed on the crater floor mounds, in the SWE ejecta, and WW walls suggest an areally large deposit of compositionally evolved rocks.

Unit NW is clearly compositionally distinct from other wall units, and is considerably more iron-rich. FeO contents and CF positions indicate a pyroxene component. Chevrel et al. (2009) suggested the unit was anorthositic with more orthopyroxene than clinopyroxene (their
unit AER). M^3 observations suggested a noritic composition, i.e., dominated by low-Ca pyroxene. The wide distribution of this unit and its stratigraphic relationships with other units (i.e., overlying units SWE, SE, and WW) make this a candidate unit for a km-thick Imbrium ejecta facies. It is also possible, given the outcrops in the crater terrace walls, that this unit could be an orthopyroxene-rich basalt, and compositionally linked to Unit NE. However, we prefer the interpretation that it is Imbrium ejecta, as also suggested by Chevrel et al. (2009). Our interpretation differs from that of Chevrel et al. (2009) with respect to their AER unit as a component of Cobra Head, Herodotus, and Väisälä, and is further discussed in section 5.5.

2.5.3 Ejecta Units (WE, SE, NE)

Five distinct ejecta units are found surrounding Aristarchus. Unit SWE, likely an evolved rock type, occurs in the southwest and east, and is discussed in previous sections. Unit NW, a probable Imbrium ejecta facies, is also abundant in the northern and southwestern ejecta blanket. In these areas it makes up much of the continuous ejecta deposits on the north rim, and is diffusely distributed around the SWE deposit in the southwest. The spatial distribution of Units NW and SWE is consistent with a layer of Unit NW overlying Unit SWE, such that during overturn during the ejection process, Unit SWE stratigraphically overlies Unit NW. Alternatively, based on its distribution on the plateau, Unit NW could be plateau surface volcanic units. However, based on the morphology of the unit, which contains clear radial striations and secondary crater chains, it is more likely that this is an ejecta unit sourced from a relatively shallow depth within the crater.

Unit WE occurs exclusively in the northern and western ejecta on the plateau. High FeO content and intermediate to long CF positions suggest a pyroxene component. Chevrel et al. (2009) have suggested this unit to be anorthositic with clino- and orthopyroxene constituents,
and note that this unit is spectrally similar to Unit NW (their unit HE), but with more Cpx. Unit WE is highly correlated with secondary crater chains and boulders from Aristarchus on the western continuous ejecta blanket. It is also found on the plateau between Aristarchus and Väisälä. It is likely that this unit is a mixture of excavated materials from Aristarchus and surface materials on the plateau (i.e., tens of meters of dark mantling deposits, and Imbrium ejecta materials). If the materials excavated by Väisälä are similar to materials excavated by Aristarchus, as we hypothesize in a later section, then this northern unit may be related to mixing of the ejecta from Väisälä and the plateau (with some component from Aristarchus, based on secondary crater chain occurrence).

Unit NE occurs as a large patch of dark albedo material at the distal margin of the continuous ejecta blanket in the northeast. It is distinguishable as a dark unit in WAC-albedo and as a dark-blue colored unit in the PCA image. The morphology of this unit is smooth, with no large boulder fields or melt morphologies. Small patches of this unit also occur in the western ejecta. Based on its appearance in the distal part of the continuous ejecta blanket, Unit NE likely was originally stratigraphically overlying Unit NW. The high FeO content and long CF positions indicate a large mafic component. In telescopic observations by Lucey et al. (1986) this unit was interpreted generically as materials with significant mare contamination. In M$^3$ parameter maps, this unit is bright green, indicating a strong low-Ca pyroxene, likely noritic component.

Unit SE is localized in the southeastern ejecta blanket, and represent olivine-rich materials that have been identified in spectral observations from telescopic studies (Lucey et al., 1986), Clementine (McEwen et al., 1994; LeMouelic et al., 1999; Chevrel et al., 2009), and M$^3$ observations (Mustard et al., 2011; Isaacson et al., 2011). The main area of this unit is a large tongue extending radially away from the rim, but also encompasses parts of the upper rim and
terrace walls. The tongue of Unit SE has no strong morphologic correlation, including no obvious melt features or boulder fields. More distal surface exposures are reportedly correlated with melt ponds (Mustard et al., 2011), although they report no olivine-rich units associated with specific ejecta blocks or breccia units. However, our morphologic mapping as revealed boulder fields on the rim and within the crater that correspond to olivine-rich areas in M3 and corresponding WAC-color regions (Fig. 22).

2.5.3.1 Possible Rock Types (WE, SE, NE)

Compositions and interpreted mineral content suggest unit WE may be a mixture of unit WW and surface plateau materials (Unit WP) or Imbrium ejecta (or both). The spatial distribution of this unit, contained only on the plateau, suggests it may be excavated plateau material from the western part of the crater and re-deposited or mixed with surface plateau material. If this unit is plateau material, it was sourced from deeper parts of the crater and it is possible that Unit WE was stratigraphically overlain by Unit NW material in the pre-impact stratigraphy.

The compositional similarities and spatial distribution of unit NW and NE suggest they are possibly related to the same source. Unit NE may be part of the original target surface. The composition of NE is suggested noritic (Mustard et al., 2011), and it is possible that this unit is an orthopyroxene-rich basalt. The location in the ejecta blanket, and the deposition of unit NW on unit NE, suggest unit NE was a pre-impact surface unit, and was probably the mare surface. The interpretation of Mustard et al. (2001) that this unit is noritic, may be the result of an orthopyroxene-rich component of Imbrium ejecta mixed with the mare basalt.

The composition and mineral content of unit SE indicates that although it is olivine-rich, there is a likely another component (pyroxene or possibly spinel, Isaacson et al., 2011). The area
covered by unit SE is not anomalously high in terms of FeO content or CF position, and cannot be readily distinguished in these datasets from other basaltic units. We interpret Unit SE as an olivine norite or gabbro, or alternatively an olivine-bearing basalt. Although uncommon, olivine-rich basalts to the extent that olivine dominates the NIR spectrum have been reported in nearby volcanic regions in Western Procellarum basalts, 250 km north, and the Marius Hills, 250 km south of Aristarchus (Staid et al., 2011; Besse et al., 2011). The petrology of this unit is difficult to explain in regional context with the silica-rich units that are closely associated with it (olivine-rich SE boulder fields are found within hundreds of meters of CF and SWE boulder fields). The location on the rim, and the radial distance away from the rim in the southeast of both Unit SE and Unit SWE suggest a deep source suggest a very close spatial relationship in the subsurface.

2.5.3.2 Source of Olivine-rich Units

The three leading candidates for the olivine-rich source materials are 1) a small, shallow olivine-rich pluton excavated by Aristarchus (e.g., LeMouelic et al., 1999; Chevrel et al., 2009); 2) re-excavation of Ol-rich Imbrium ejecta facies (e.g., Yamamoto et al., 2010 (implied); Wiseman et al., 2012); 3) melting and excavation of olivine-rich Procellarum basalts (e.g., Mustard et al., 2011; Issacson et al., 2011). LeMouelic et al. (1999) concluded that the olivine was sourced from shallow crustal depths, as evidenced by its location on the rim of the crater, and was thus derived from a shallow pluton, possibly related to the formation of the plateau and pyroclastic deposits. A troctolite or dunite rock type was proposed (LeMouelic et al., 1999). The proximity of the olivine-rich areas to the crater rim suggests a very shallow source. We estimate the excavated olivine must be sourced from the upper ~< 2.5 km of the pre-impact surface. Chevrel et al. (2009) concluded that the olivine was excavated from an olivine-rich layer that sits above an anorthositic crust, but beneath the noritic Imbrium ejecta. SELENE Multiband Imager
mapping of global olivine deposits by Yamamoto et al. (2010) observed that olivine-rich deposits are found clustered around large impact basins and suggested that the global deposits are mantle-derived ejecta, although they do not specifically state this as the origin of olivine at Aristarchus. Wiseman et al. (2012) noticed a similar global distribution of olivine-bearing materials in M3 data in the vicinity of Imbrium and that some of these deposits have been exposed by subsequent impacts, such as Aristillus. An investigation of the olivine deposits on the Moon using M3 by Isaacson et al. (2011) show that the Aristarchus olivine-rich areas are not Fe-rich and are spectrally distinct from other circum-Imbrium deposits; in that they have features that suggest contributions from other phases, and they are likely formed in a different geologic context to the olivine deposits around, for example, Copernicus and Moscoviene.

Based on the spatial relationships and morphologic correlation with impact melt ponds in the SE ejecta, both Isaacson et al. (2011) and Mustard et al. (2011) suggest the olivine-rich areas are formed by the crystallization of a basalt-source impact melt, a formation process that would help explain the presence of pyroxene, possibly Cr-spinel, and other phases that distinguish the Aristarchus olivine deposits from other circum-Imbrium olivine deposits. Isaacson et al (2011) also noted that Fe-bearing glass, which is likely found associated with the plateau pyroclastics, could be a contaminant affecting the appearance of the olivine spectra at Aristarchus. Some western Procellarum basalts (Staid et al 2011) and Marius Hills volcanics (Besse et al., 2011) are apparently olivine rich, and the olivine-rich regions at Aristarchus may be related to similar composition mare basalt flows. However, as pointed out by Wiseman et al. (2012), these materials also have significant pyroxene contributions to the olivine spectra, which are not seen in the Aristarchus units, despite the impure nature of the olivine-rich materials there (Isaacson et al., 2011).
Ol-rich boulder streams are present on the rim of the crater in the south and southeast ejecta. The olivine signature seen in the impact melts in this area may be related to embedded clasts in the melt flows, and not intrinsic to the melt itself. The presence of Ol-rich boulders suggests that the olivine source is a buried target rock, which implies either thick Ol-rich basalts or a hypabyssal intrusive. Based on a relatively dark albedo and distribution seen in WAC color, these Ol-rich boulder fields are plausibly mare basalt units (although clear stratification, seen in proposed basalt units in the northeast, is not seen). The interpretation of spectral data of olivine-rich basalts in other, nearby, Procellarum volcanic regions (Staid et al. 2011; Besse et al., 2010) also contributes evidence for an olivine-rich basalt extrusive source.

A shallow plutonic source cannot be ruled out. However, if Aristarchus has excavated highly silicic material, also likely related to the excavation of a hypabyssal intrusive, a petrologic explanation that would relate the silicic and olivine-rich lithologies is unlikely. With this consideration, although it is chemically slightly different from other circum-Imbrium olivine-rich deposits, we prefer the interpretation that the olivine-rich material is re-excavated Imbrian Basin ejecta (Wiseman et al., 2012).

2.5.4  **Pre-impact Surface (WP, SM)**

Unit WP occurs beyond the Aristarchus ejecta blanket on the west plateau. The unit is FeO rich with long CF positions, consistent with the interpretation of Chevrel et al. (2009) of pyroxene-rich materials contaminated with mare materials. A DMD glass component is also likely, as most of the plateau is mantled in > 10 – 30 m thick pyroclastic deposits (Gaddis et al., 1985; McEwen et al., 1994; Weitz et al., 1998). Unit SM occurs on the smooth, flat, Oceanus Procellarum flood basalts in the east and south of the crater. High FeO and long CF positions are consistent with mare basalt. The extensive ray system from Aristarchus also lightly mantles
much of this unit, and has the effect of slightly lowering the FeO content (~2-3 wt%) compared to mare basalts beyond the rays. Unit SM also appears associated with the infilled Herodotus Crater and adjacent to Vallis Schröteri, consistent with a mare basalt interpretation. It is likely that both of these materials represent the pre-impact target surface, and were a covering over materials ejected by the crater. Both of these materials are expected to have mixed with ejecta units to some degree.

2.5.5 **Anomalous Plateau Units Correlated with Aristarchus Ejecta**

Three anomalous regions on the plateau, the western wall of Herodotus, the eastern wall and areas of the floor of Cobra Head, and Väisälä crater, share compositional characteristics with ejecta units CF, SWE, and WW. Each region on the plateau is high albedo, low FeO, and has short CF positions (Figs. 32, 36). Additionally, these areas correspond to $M^3$ parameter maps spectrally indistinct units (Mustard et al., 2012), and units in Clementine UVVIS mapping (Chevrel et al., 2009).

Chevrel et al. (2009) noted that these regions are spectrally similar to their endmember iterative linear mixing model unit AER (Anorthositic with Cpx and Opx; located predominantly in the northeast wall), as well as their unit SC (Anorthositic with pyroxene, similar to the central peak, but with less feldspar and pyroxene). The occurrence of these units in the ejecta materials of Aristarchus and in the three regions on the plateau was taken as an indication of widespread distribution within the southern part of the plateau. They concluded that their AER (our NW) was a regionally extensive layer of Imbrium ejecta. They also inferred a relationship between the southwestern eject unit (our SWE, their SC) and the olivine-rich unit (our SE, their OL), which places these units within a laterally widespread layer, consistent across the plateau. Mustard et al. (2012) noted the correlation of these anomalous regions with $M^3$ parameter mapping results
suggesting spectrally indistinct material and indicating glass. Their analysis stemmed from their
interpretation of the materials in the southwest ejecta being impact melt glass, and they
interpreted the anomalous areas as evidence of pyroclastic glass due to their similar spectral
shape in $M^3$ (Mustard et al., 2012).

It is worth noting that along nearly the entire length of Valles Schröteri immature surfaces
and fresh weathering of boulders occurs. The walls of the rille have the same steep angles and
have been imaged at the same viewing geometry as the anomalous non-basalt exposure in Cobra
Head, but these areas do not have the same anomalous composition. Indeed, that these regions
are anomalous in multiple datasets, despite being observed by a number of different spacecraft
under different viewing geometries, may be significant. If walls of Valles Schröteri share the
same maturity (i.e., freshness), weathering (i.e., grain-size), and slopes (i.e., viewing geometry)
to the anomalous region in Cobra Head, then we may only be left with a compositional
difference as an explanation. Because Cobra Head is the peak of a volcanic construct, it must
have formed over millions of years or longer, and clearly post-dates the plateau. Clear layering
of lavas, similar to that observed in the walls of large mare craters, such as Kelper (Ohman and
Kring, 2012) and in mare pits (Wagner et al., 2014) are not seen in the walls of Cobra Head, but
are observed in distal areas of Valles Schröteri.

We interpret these anomalous areas on the plateau to represent separate occurrences of the
silica-rich rocks that are present in the ejecta of Aristarchus. This interpretation implies that CF-
like units are areally widespread in this region, either as a coherent layer beneath a cover of
Imbrium ejecta and/or DMD; or as a series of discrete subsurface intrusions. The morphology of
the Cobra Head volcano is a prominent topographic high, and is the source of Vallis Schröteri,
the largest basaltic sinuous rille on the Moon. However, the steep sides of the volcano (relative to
other basaltic volcanoes on the Moon, such as the shield-like Hortensius domes), and numerous nearby mounds of presumably constructional volcanism, suggest that the topography may be related to other processes than basaltic volcanism. Silicic volcanic structures on the Moon, such as the Gruithuisen domes, and the Compton Belkovich Volcanic Complex (CBVC) (Chevrel et al., 1999; Jolliff et al., 2011), share morphologic similarities with the Cobra Head volcano, with volcanic constructs that include steep topography and strong compositional anomalies seen in Diviner CF. Additionally, Väisälä Crater is perched on what may be interpreted as a constructional mound. If units CF and SWE make up a large component of the plateau material in the southern part of the Aristarchus Plateau, then an origin of these units as extrusive volcanics, rather than intrusive plutonic rocks, may also be possible. The CF positions indicating silica-rich compositions are also consistent with obsidian (Glotch et al., 2010), which may have contributed to the construction of Cobra Head, and is the material currently being exposed.

2.5.6 **Regional Emplacement of Evolved-Lithologies**

If the rocks exposed in the floor and ejecta of Aristarchus crater and other regions on the plateau are, in fact, evolved silica-rich rocks, then what emplacement scenario can produce them? A case for the role of basaltic underplating in forming non-mare volcanic units on the Moon was made by Hagerty et al. (2006). Here, we summarize his model for the formation of volumetrically significant amounts of silica-rich magmas on the Moon, and apply it to our observations.

In the basaltic underplating model (from Hagerty et al., 2006; adapted from Hildreth et al., 1981) a hot, mantle-derived plume of basaltic magma rises and intrudes into the crust and ponds at some level where it can heat and initiate melting in overlying rocks or rocks into which it intrudes, producing partial melts that have high-silica contents. This process requires “fertile”
crust, i.e., rocks of a composition or initial thermal state such that heat supplied by the basalt can cause sufficient partial melting. The high-silica partial melt is less dense than the surrounding crust and can rise and further intrude the crust to shallow levels or possibly extrude. If the basaltic underplating occurs in an area of incompatible-element-rich rocks such as the Procellarum KREEP Terrane (PKT; Jolliff et al., 2000), then partial melting is more likely and such melts will be rich in incompatible elements including K, Th, U, and REEs. Hagerty et al. (2006) proposed this model as a mode of emplacement for silica-rich volcanic regions like the Gruithuisen domes, Hansteen Alpha, and other lunar red spots. Hagerty et al. (2009) further proposed basaltic underplating as the likely mode of Th-enrichment of Aristarchus Plateau DMD glasses and the Th-enriched material excavated by Aristarchus Crater.

In this model, the rock of units CP, CF, and SWE (as well as WW, and possibly WE; see unit discussion) represent plutonic rocks formed when basaltic underplating occurred under KREEPy rocks beneath the plateau. Essential to the basaltic underplating model is a close spatial association with large-scale basaltic volcanism that can serve as a source of heat for the underplating process (Hagerty et al., 2006). As the source of the largest concentration of lunar sinuous rilles, Th-rich pyroclastic DMD deposits, and extensive mare volcanism, the Aristarchus Plateau meets this requirement. A KREEP-rich body of melt, produced by partial melting, might also have undergone fractional crystallization to produce a gabbroic layer through gravitational settling, an alkali anorthosite layer in a neutrally buoyant zone, and lastly a silica-rich layer or zone in which silicate liquid immiscibility occurred during the last phases of crystallization. When this pluton was excavated by the Aristarchus forming impact, material from the upper silica-rich and incompatible-element-rich layer, and the deeper, alkali anorthosite layer were exposed in the crater floor and ejecta.
In this scenario, the wide dispersal of incompatible elements (evidenced by the distribution of K and Th) are found in units SWE and possibly as the monzogabbro admixture inferred from FeO-Th concentrations on the mare. The areas of the crater floor with the shortest CF positions (i.e., most-silicic) and lowest FeO content might indicate exposures of more granitic compositions. The central peak (unit CP) would have sampled most deeply and exposed the alkali anorthosite layer (evidenced by albite-like CF position and very low FeO content). Materials into which the pluton intruded; the plateau, Imbrium ejecta layer, and mare flood basalts, were also excavated. The close proximity of much of the silica-rich, unit CF materials to the rim of Aristarchus imply a shallow source in terms of position within the lunar crust (<3 km), but from a deep area of the excavated zone (i.e. close to limits of excavation; 2-3 km). If these materials are plutonic, then they were intruded very close to the surface, and are only mantled by ~2 km of Imbrium ejecta and ~500 m of mare basalt.

If an evolved pluton intruded the plateau shallowly enough to be excavated by Aristarchus, then the other outcrops of units SWE and CF in the anomalous regions of the plateau can be explained by either very-near-surface shallow plutonism, or possibly rhyolitic domes. If the silicic material was extrusive, possibly from more than one vent, this can explain the steep-walled topography of the Cobra Head volcano, and its similar morphology to other suspected regions of silicic volcanism such as the Compton-Belkovich Volcanic Complex. If the outcrops of units CF and SWE in the anomalous regions on the plateau are evidence of this here-to-fore unrecognized silicic-volcanism in this region, then it is possible that the units are much more widespread, but buried under hundreds of meters of Imbrium ejecta and DMDs that have obscured this evidence.
However, the cobra head volcano shows clear evidence of basaltic volcanism. Valles Schröteri is a sinuous rille that is sourced from Cobra Head and extends for more than 150 km to the northwest. The huge volume of DMD glass is also likely partially sourced from Cobra Head, in addition to other basaltic volcanic vents on the plateau. It is possible to reconcile both basaltic and rhyolitic eruptions from the same source if Cobra Head has undergone bi-modal volcanism. Bi-modal volcanism is a common bi-product of basaltic underplating and the production of rhyolite domes from mantle plume hot-spots on Earth, and would allow for both basaltic and rhyolitic eruptions from the same vent, but from different magma chambers (Freundt-Malecha et al., 2001).

2.5.7 On the possibility of a South AP Crater

Using LOLA DEMs, Mustard et al. (2011) have suggested a previously unidentified 110 km diameter, heavily degraded impact crater directly south of the Aristarchus Plateau (AP), which they named South AP crater. Because the ejecta from a heretofore unidentified crater could be re-excavated by Aristarchus, and therefore play a role in our geologic interpretations, it is important to consider this possibility in detail. The northern wall of this crater is purportedly the reason for the occurrence and location of the scarp along the southern edge of the Aristarchus Plateau, and materials excavated by this pre-Imbrian aged crater would have been emplaced over the pre-impact site of Aristarchus Crater, as well as that of Herodotus Crater. As a result, materials excavated by Aristarchus would be predominantly composed of South AP ejecta, with only minor components of Imbrium ejecta at this location. Evidence presented in favor of the existence of South AP is limited to the slightly curved nature of the plateau scarp, and that the nearby 45 km crater Prinz is heavily infilled by mare (Mustard et al., 2011). In support of a South AP crater we note that a positive circular Bouguer gravity anomaly occurs at the base of
the plateau, centered approximately 60 km from the plateau scarp, and could be interpreted as

evidence for a ~110 km crater that had been filled by relatively dense mare basalt.

However, a number of observations do not support the presence of South AP. Although the
plateau scarp appears curved, owing to a buildup of Aristarchus blanket in the southwest, the
scarp can actually be seen as a slight ridge below the ejecta extending linearly until it intersects
the Aristarchus Crater rim. If South AP was indeed buried by mare basalts, they would have to
be >1.2 km thick, however, DeHon (1979) report <500 m of basalt flooding the southern part of
the plateau. Moreover, in all other cases of very large craters that are infilled by mare, for
example, the Flamsteed ring and Letronne, two craters that are ~110 km in diameter, large parts
of the crater rim remain visible above the mare. Directly within the center of South AP lies a
small volcanic source vent, and in the region of the western ejecta is a small shield volcano, both
of which would have had to form through South AP crater facies. Additionally, although there is
a circular positive Bouguer gravity anomaly, the margins of the plateau are linear in areas not
associated with Aristarchus Crater. To the east of the plateau, extending to the infilled Prinz
Crater and wrapping around the southern edge of the plateau, free-air gravity anomalies suggest
the presence of thick surface flows and subsurface intrusions of high-density mare basalt (Kiefer
et al., 2013).

Although we cannot rule out the presence of a South AP crater, we find it unlikely that
such a large crater could be completely buried as to leave no evidence except for the 1 km scarp
on the plateau, and possibly a gravity signature. Although not suggested by Mustard et al.
(2011), it may be that NE-SW trending wrinkle ridges on the mare plains could be mistaken for a
completely buried crater rim. Irrespective of whether or not South AP exists, we do not think its
presence would have a great effect on the materials excavated by Aristarchus. As South AP
would have formed pre-Imbrium, it would have been subjected to the same 1-4 km thick ejecta deposition as seen around other parts of the rim of Imbrium (Haskin et al., 2003), and further buried by ~500 m of mare basalts. As the excavation depth of Aristarchus is < 3 km, materials in the ejecta of Aristarchus are possibly sourced entirely from Imbrium ejecta and mare basalts. Furthermore, as the materials in the central peak are nearly identical to those in the ejecta, they likely share the same source.

2.6 Geologic History of the Southern Aristarchus Plateau

If our interpretations are correct, the previously interpreted geologic history of this region outlined by Zisk et al. (1977) and modified by Chevrel et al. (2009) and Mustard et al. (2011), and presented in the introduction may require revision. We propose the following scenario:

A. A region of ancient highlands crust, located above a residual incompatible-element-rich layer of the Procellarum KREEP terrane, was uplifted and tilted during the Imbrium Basin forming event, and comprised the basement rocks of the Aristarchus Plateau. The plateau was mantled by < 2 km of Imbrium ejecta, based on the amount of material excavated by Aristarchus Crater (within the 1-4 km thick estimate of Haskin et al., 1998).

B. A mantle plume of basaltic magma rose beneath the plateau and intruded the ancient highlands crust. The intrusion or multiple intrusions were enriched in incompatible elements due transport of the basalt plume through a magma-ocean residual KREEP layer or concentration. The intrusions formed silica-rich or KREEP-rich partial melts that differentiated to produce alkali-anorthositic concentrations or cumulates beneath the southern margin of the plateau. The intrusion excavated by Aristarchus was intruded to at least 4 km below the mare surface (the sampling depth of the central peak). Outcrops on the plateau indicate that at least some material intruded Imbrium ejecta to the surface.

C. Herodotus formed on the plateau, exposing a small area of silica-rich material.

D. At this stage, intrusions are now very near the surface, within the layer of Imbrium ejecta. Cobra Head began forming, and may have formed prior to the formation of Herodotus. Current topography indicates growth of the Cobra Head volcanic construct after the crater
formed, evidenced by the formation of small domes in what would have been the ejecta blanket of Herodotus.

E. Either contemporaneous with, or subsequent to the major topography-forming growth of the Cobra Head, low viscosity, high temperature basaltic magma erupted from Cobra Head and from other sinuous rille source regions, probably with large scale fire-fountain style eruptions. At least 2, and possibly many more basaltic volcanic eruptions took place, the last forming the small secondary channel within Vallis Schröteri. The combinations of basaltic volcanism, and silicic intrusion or rhyolitic volcanism occur contemporaneously, likely covering the rising dome with thin basalt flows. Major volcanism occurred during the highest flux of mare magmatism, between 3.6 and 3.8 Ga (Shearer and Papike, 1999), however, crater size-frequency measurements of the basalts surrounding the plateau suggest basaltic volcanism persisted late into the Eratosthenian, ending as late as 1.2 Ga (Hiesinger et al., 2003; Stadermann et al., 2015).

F. Väisälä crater formed on a small dome north of Aristarchus, exposing silicic material.

G. Aristarchus Crater formed, exposing a large pluton buried within ~2.5 km of the surface and mantling the region with Th-rich material. Both Väisälä and Aristarchus occur during the Copernican period, within the last Ga. Aristarchus likely formed ~250 Ma (Zanetti et al., 2015; Chapter 3 of this thesis).

2.7 Conclusions

This paper uses multiple remote-sensing datasets map the geomorphology and geology of the lunar crater Aristarchus. Geomorphologic mapping has improved the detail of existing maps, and made possible a number of new observations regarding the distribution of ejecta and morphologic dependence with composition. Major results from geomorphologic mapping are as follows:

1) A complete map of the morphologic units around the crater, including floor deposits, wall units, and ejecta units; and 1:24,000 scale mapping of impactites (ponded impact melt and boulder fields).

2) The distribution of impactites is asymmetric with respect to topography, with fewer melt ponds (but larger area melt flows) occurring on the elevated plateau side of the crater.
Boulder fields are abundant on the interior wall of the plateau, but found only in the ejecta on the lower elevation mare side of the crater, probably a result of topography affecting ejection dynamics.

3) Impact melt can remain mobile long after the main phase of crater modification process has ended, evidenced by large dark melt flows, flows crossing ejecta rays, and impact melt pond collapse.

4) Splatter flows, likely related to impact and subsequent flow of molten ejecta bombs, indicate ejected material can be delivered to the ejecta blanket after the ejecta curtain and ballistic sedimentation has taken place.

5) Stratified ejecta blocks were first reported at Aristarchus (Zanetti et al., 2011), and are probably evidence of fragmented mare basalt units.

6) There is little correlation between crater morphology and composition. Boulder fields can be correlated to specific compositions, but morphologic units such as impact melt ponds and impact melt flows are nearly always compositionally similar to their surrounding ejecta units.

7) Impact melts can, however, be compositionally heterogeneous with respect to position in the crater, and may reflect local melt composition, rather than a homogeneous mixing of melt into a compositionally distinct unit. This is counter to assumptions about impact melt homogenization in terrestrial crater examples (Grieve et al., 1998).

The geology of the crater floor, ejecta, and southern Aristarchus Plateau is complex and heterogeneous, representing a diverse suite of rocks, including mare basalt, Imbrium ejecta, dark mantle deposits (DMDs), and evolved lithologies rich in incompatible elements. We used principal component analysis of WAC-color data, and WAC-color ratio imagery to produce classified image maps of compositionally distinct units. These maps were compared with compositional datasets including Clementine FeO, Diviner CF, LP-GRS thorium and potassium, Clementine UVVIS, and M^3 data, and a new geologic map of Aristarchus Crater and southern Aristarchus Plateau was created. Major results include:

8) A geologic map with ten major compositional units related to facies within the crater and its surroundings.
9) Identification and mapped distribution of compositionally evolved units in the central peak, crater floor, and southwest ejecta (units CP, CF, and SWE, respectively), possibly indicating the presence of silica-rich rocks such as: Quartz monzodiorite, monzogabbro, and potentially granite or rhyolite. Related rocks inferred from high FeO and high Th compositions may indicate monzogabbro.

10) Anomalous regions on the plateau comprised of low-FeO, silica-rich compositions are seen in the western wall of Herodotus, the eastern wall of the Cobra Head volcano, and the crater Väisälä, north of Aristarchus.

11) We relate observations of silica-rich materials to a regional emplacement scenario whereby basaltic underplating and intrusion of KREEP-enriched magma into the plateau caused the formation of numerous small (or possibly one large) alkali-anorthosite, and evolved lithology plutons forming domes similar to silica-rich volcanic complexes (e.g. Compton-Belkovich Volcanic Complex).

12) We suggest the Cobra Head volcano is a topographic construction built though possible bimodal eruptions of silicic/rhyolitic volcanism, and low viscosity basaltic volcanism.

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Figures:

Figure 2-1: A) The Aristarchus Plateau as seen in the WAC low-sun morphology mosaic. The plateau rises 1-2 km above the surrounding Oceanus Procellarum mare flood basalts. Numerous sinuous rilles emanate from volcanic vents on and around the plateau, and wrinkle ridges are abundant in the basalt plains. B) Clementine color ratio RGB composite image of the same scene. Note the red color of the plateau, indicating the presence of dark mantling deposits (DMD) and the sharp contact with the surrounding basalts. Aristarchus Crater and its ejecta rays are clearly visible in the southeast corner.
Figure 2-2: A) WAC mosaic of Aristarchus Crater and neighboring regions. B) WAC GLD100 stereo topography of the same scene. Topographic cross-sections A-A’ and B-B’ are shown in Figure 5.
Figure 2-3: WAC-color composite image (R: 689, G: 565 nm, B: 321). Note areas of high albedo in the central peak, the Cobra Head volcano, and
Figure 2-4: A) NAC low sun angle controlled mosaic created for morphologic mapping. B) NAC high-sun controlled mosaic. Note the albedo variation between the western and eastern walls.
Figure 2-5: Topographic Profiles across Aristarchus Crater. Both profiles cross the central peak. Locations given in Fig. 2-2.
Figure 2-6: Previous geomorphologic mapping efforts of A) Guest (1973); and B) Spudis and Guest (1985).
Figure 2-7: A) M3 parameter map from Mustard et al. (2011). Map is an RGB composite image with R: Integrated band depth (IBD) at 1 μm; G: IBD at 2μm, and B: reflectance at 1.54μm. Red areas are olivine-rich; green areas are pyroxene-rich (high-Ca); blue is “spectrally unremarkable”. B) Clementine UVVIS spectral parameter map of Chevrel et al., (2009). Color coded units are Red: AP (Aristarchus Peak); Blue: AER (Aristarchus Eastern Rim); Green: SC (Scarp); Purple: OL (Olivine); Pink: HE (Herodotus)
Figure 2-8: RGB composite WAC color ratio images. A) R:689/321 nm G:415 nm B:321/689 nm. A close-up of the crater is seen in Fig. 10a. B) Ultraviolet focused ratio image: R:415/321 nm G:360 nm B: 15/321 nm in blue
Figure 2-9: WAC color principal component analysis (PCA) RGB composite image. R: Principal Component 1; G: PC2; B: PC3. Areas with the most variance are red and pink, areas with the least are blue. A close-up of the crater is shown in Fig. 10b.
Figure 2-10: Close-up detailed views of the crater in A) WAC color composite (same bands as Fig. 8a) and B) WAC color PCA (same bands as Fig. 9b).
Figure 2-11: GRAIL Bouguer Gravity. Note the anomalously high circular region in the south. Plateau margins are marked with a yellow dashed line.
Figure 2-12: Geomorphologic Map. Units are described in the text. A fold out map is provided.
Figure 2-13: NAC controlled mosaic context image with map unit context figures labeled.
Figure 2-14: A) Smooth Floor; note large floor fractures and cracks (mapped as structural units in blue on fold out map). B) Hummocky floor; large hillocks of slumped wall material and possible uplifted material. Many hills and mounds correlate with low FeO and short CF position compositions.
Figure 2-15: A) Aristarchus Central Peak. Areas labeled a-e correspond to areas with varying albedo described in the text. Unit (a) is likely impact melt covering the peak. Unit (b) is the high albedo central portion. Plates of dark material on (b) appear similar to (c), indicating (c) may be a thin veneer over a high albedo unit.
Figure 2-16: The southwest crater wall. The crater rim is toward the lower right of the image, the crater floor is labeled in the upper left. Note the large number of melt features. Prominent channels are marked with dashed lines. Melt ponds are labeled (p). (NAC-DEM over NAC controlled mosaic).
Figure 2-17: Concentric ridged ejecta unit (r) is common in the eastern half of the crater, on the topographically low mare side of the crater. Ridges are often embayed by melt ponds (p). Crater rim is to the lower left.
Figure 2-18: Melt-rich channels and lobes found along a break in slope in the southwest ejecta. Channels and flows are radial to the crater rim (toward the upper left).
Figure 2-19: A) The largest terrace melt pond within Aristarchus, showing evidence of roof collapse and melt drainage. Yellow arrows point to scarps where the roof collapsed, red arrow show incoming and outgoing drainage channels. B) NAC stereo DEM showing pool and area of collapse. The pond area is ~1.5 km², and drained ~0.025 km³.
Figure 2-20: Example of “globular melt” texture. Small mounds are possibly remnant melt blobs that have remained behind after rapid melt withdrawal.
Figure 2-21: Largest melt flow within the ejecta of Aristarchus. Flow extend nearly 15 km from melt ponds at the rim, and shows evidence of multiple flow events.
Figure 2-22: Flow crossing ejecta ray. A) Melt streaks, channels (dashed lines) and ponds (p) extend to the edge of the continuous ejecta blanket. B) a lobate melt flow extends from a collection alcove and crosses a bright ray of ejecta. Melt flow has a channelized middle section. (dashed line at top of image is the crater rim). Note a field of dark boulders in close proximity to the brighter boulder field. The dark boulder field can be correlated with the olivine-rich unit (SE) in the WAC color PCA image, the M3 parameter map, and our classified image geologic map.
Figure 2-23: Splatter flows in the southwest ejecta blanket likely indicate late-arriving molten blobs of ejecta landing after the main phase of ejecta blanket emplacement. Flows form radial to the crater rim (upper right of the image).
Figure 2-24: Melt Veneer can be mapped based on the occurrence of very small <50m irregular impact craters on melt surfaces. A) distribution of craters (green dots) that resemble those seen in B. Red line marks the limit of recognizable impact melt veneer. B) Some irregular craters appear to have central mounds, possibly a remnant bolide that impacted into a still molten impact melt.
Figure 2-25: Ejecta blocks with alternating dark and light banding mantle the northeast wall of the crater, often embedded within impact melt ponds. A) large collection of layered material with many examples of stratified blocks. B) One of the clearest examples of stratified material, showing clear alternating bands of thick-higher albedo material and uniformly thin bands of dark material. C) a large fractured stratified boulder.
Figure 2-26: One of the largest clearly stratified blocks contains more than 20 bright layers separated by thin dark banding. Our leading hypothesis is that these represent blocks of fragmented lava packages.
Figure 2-27: Locations of stratified boulders identified in highest resolution NAC imagery (< 0.5 m/pixel). Stratified blocks have only been observed on the northeastern wall of the crater, suggesting a mare basalt unit was located there. Stratified blocks have also been identified at a small number of other mare target craters.
Figure 2-28: Locations of impact melt units (red) and boulder fields (yellow) on A) WAC stereo GLD100 digital terrain model (DTM) and B) Diviner rock abundance. Note the dichotomy in melt and boulder distributions between the topographically high plateau and lower elevation mare. Boulders mantle the western plateau wall, but few melt ponds are found on the plateau (although the largest melt flows are found there). In the east, boulders fields occur radial to the crater rim and melt ponds frequently occur at the break in slope between the ejecta and mare floor.
Figure 2-29: “Banding” in the Bonito lava flow field in Sunset Crater National Monument, Flagstaff, Arizona. Bright crystalline lava cores are capped and floored by a’a’ texture, formed as the lava flowed. Based on the similarity in morphology, dimensions, and relative albedo differences, we suggest stratified blocks may have been formed in a similar manner. Graduate student Ryan Nickerson for scale.
Figure 2-30: Geologic Map of Aristarchus Crater created through image classification of WAC principal component analysis. Ten units are identified and described in Table 3. Crater Floor (CF), South West Ejecta (SWE) and West Wall (WW) units are observed in anomalous regions on the plateau in the west wall of Herodotus, the southeast wall of Cobra Head, and the walls of Väisälä Crater in the north (other small craters on the plateau also expose similar units).
Figure 2.31: Interpretative and highly schematic classification scheme for rock types present in geologic units. Actual mineral compositions are not measured, and the diagram is only intended to aid interpretation.

- *Highly schematic classification of the inferred rock types at Aristarchus Crater.
- Actual mineral compositions are not measured.
- Diagram provided to aid interpretation.
Figure 2-32: A) Clementine spectral reflectance derived FeO concentration map (after Lucey et al., 2000). FeO concentrations range from < 5wt% in the central peak to >18wt% in the surrounding mare (and northeastern ejecta). B) Diviner modeled Christiansen Feature (CF) position map at 128 pixel/degree (~215 m/pixel). CF positions of CF and SWE units are often <7.7 μm, and correlated with boulders. Long CF positions (<8.3) are found in unit SE (Olivine-rich), NE (FeO rich ejecta) and the surrounding mare and plateau units.
Figure 2-33: Detail of the (A) FeO and (B) modeled CF positions around Aristarchus Crater and nearby ejecta. Concentration scale bars have been modified slightly to make differences more obvious.
Figure 2-34: Lunar Prospector Gamma Ray Spectrometer (GRS) Thorium abundance map. Aristarchus Crater is a prominent “hotspot” within the Procellarum KREEP terrain. Thorium abundance within the crater is measured at ~12 ppm, with modeled abundance ~ 15 ppm (Hagerty et al., 2009). Yellow dashed line marks the outline of the plateau.
Figure 2-35: A) Laboratory spectra of major rock-forming mineral and spectra convolved using methods for Diviner modeled CF position (red lines) (Modified from Glotch et al., 2010). B) Illustration of how modeled CF position is fit to Diviner CF channels (Ch3: 7.8 µm; Ch4: 8.25 µm; Ch5: 8.55 µm) provided to aid interpretation (Modified from Song et al., 2013)
Figure 2-36: A) Plot of modeled CF position (µm) versus FeO content (wt %) for whole scene. Moving average (50 points) fit to show the general trend of increasing FeO content with increasing CF position. B) Map of pixels (in red) that are <11 wt% FeO and <8µm CF. Note the 1:1 correlation with Units CP, CF, and SWE in our Geologic Map (derived from WAC PCA classification), and that anomalous regions near Herodotus, Cobra Head, and Väisälä Crater are also correlated.
Figure 2-37: A) Plot of Clementine FeO versus LP-GRS Th for regions of feldspathic highlands (green dots), mare basalt (blue dots), and KREEP-rich lithologies for Western Procellarum, including Aristarchus (from Jolliff et al., 2004). B) Same plot with data collected from our study region. Note the trend of red dots (corresponding to units CF and SWE) toward alkali anorthosite. Also note that high FeO and high Th also occur, and may represent mixing of an evolved lithology (see Fig. 38). C) Red pixels in correspond to red dots in B).
Figure 2-38: Plot of FeO (wt%) versus Th abundance (ppm) in laboratory analyzed Apollo samples (mostly Apollo 12) (modified from Jolliff et al., 2004). Range of values for the areas in units CP, CF, and SWE of the geologic map are shown in the red dashed oval. Range of values for high-FeO, high-Th pixels (Fig. 37b) are shown in the light blue circle. These materials fall on an inferred mixing line between evolved lithologies (e.g. monzogabbro) and more primitive basalts and glasses. That values occur in this field is significant because it implies high FeO, high Th material was excavated by Aristarchus and mixed with local surface units.
Tables:

Table 2-1: Aristarchus Crater morphometric parameter measurements vs predicted values

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<th>Parameter*</th>
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*from Melosh 1989 and references therein
Table 2-2: Values of the Principal Component Analysis of 7 WAC color bands

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**EIGENVALUES AND EIGENVECTORS**

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* (reported by Glocot et al., 2010; Song et al., 2013)
** (modeled by Hagerty et al., 2009)
*** (inferred from location of reported spectra in their Fig. 11c)
Unit abbreviations from Chevrel et al., (2009) in brackets
Chapter 3: Evidence for Auto-Secondary Cratering of Copernican-Age Continuous Ejecta Deposits on the Moon

Associated publication: M. Zanetti¹, A. Stadermann¹, H. Hiesinger², C. H. van der Bogert², J. Plescia³, B. Jolliff¹. ¹Washington University in St Louis, Earth and Planetary Science Department and the McDonnell Center for Space Sciences; ²Westfälische Wilhelms-Universität Münster, Institut für Planetologie; ³Johns Hopkins University Applied Physics Laboratory.

Icarus (in revision).

Abstract

Crater size-frequency distributions on the ejecta blankets of Aristarchus and Tycho Craters are highly variable resulting in apparent absolute model age differences despite ejecta being emplaced in a geologic instant. Crater populations on impact melt ponds are a factor of 4 less than on the ejecta and crater density increases with distance from the parent crater rim. Although target material properties may affect crater diameters and in turn CSFD results, they cannot completely reconcile crater density and population differences observed within the ejecta blanket. We infer from the data that auto-secondary cratering, the formation of impact craters immediately following the emplacement of the continuous ejecta blanket by ejecta from the parent crater, contributes to the population of small craters (<300 m diameter) on ejecta blankets, and must be taken into account if small craters and small count areas are to be used for relative and absolute model age determinations on the Moon. Our results indicate that the cumulative number of craters per unit area, $N(1)$, on the continuous ejecta blanket at Tycho Crater ranges between $2.17 \times 10^{-5}$ and $1.0 \times 10^{-4}$, with impact melt ponds most accurately reflecting the primary crater flux ($N(1) = 3.4 \times 10^{-5}$). Using the cratering flux recorded on Tycho impact melt deposits calibrated to accepted exposure age (109 ± 1.5 Ma) as ground truth, and using similar crater distribution analyses on impact melt at Aristarchus crater, we infer the age of Aristarchus Crater to be ~280 Ma.
3.1 Introduction

Large, complex impact craters are capable of resurfacing thousands of square kilometers in the areas surrounding the parent crater, thus offering what may be assumed to be a completely reset surface for the accumulation of new craters following ejecta emplacement. The process of emplacement can bury the area surrounding the rim in meters to kilometers of ejected material (e.g., McGetchin 1979, Melosh 1989), which is enough to bury craters hundreds of meters in diameter. Through the process of ballistic sedimentation (e.g., Oberbeck 1975, 1979) an even larger area surrounding the parent crater rim is thought to be disturbed enough to be considered resurfaced. Ejecta blankets offer the largest and most straightforward areas surrounding a parent crater on which to determine crater size-frequency distributions (CSFDs) and establish the age of the impact. Crater walls and terrace melt ponds are prone to loss of craters through erosion on steep walls and are commonly too small to record a statistically useful diameter range of craters over geologic time. The melt sheets on the floors of Copernican-aged craters are rugged, chaotic areas with blocky and ropey textures containing many cracks and crevasses, which hinder the identification of subsequent impact craters, and as such are difficult surfaces on which to count to confidently record complete crater distributions. Counting craters superposed on the continuous ejecta blanket should therefore give an estimate of the flux of craters since the large parent crater’s formation, a technique used on ejecta blankets of Copernicus crater (Neukum, 1983; Hiesinger et al., 2012), Tycho Crater (e.g. Hartmann et al., 1980; Neukum 1981; Neukum and Ivanov 1994; Hiesinger et al., 2012), North Ray (Apollo 16 site), and Cone (Apollo 14 site) craters to establish the recent (< 1 Ga) Solar System cratering rate and anchor the lunar cratering chronology (e.g., Hartmann et al., 1981; Neukum, 1983; Neukum and Ivanov, 1994; Neukum et al., 2001; Ivanov and Hartmann, 2006; Hiesinger et al., 2012).
The assumption that areas within the continuous ejecta blanket are cleanly resurfaced has allowed for the calibration of a lunar chronology to the size-frequency distribution of craters on the ejecta blanket. Tycho is one of the youngest large craters on the Moon and its continuous ejecta blanket shows no obvious overprinting of secondary craters from nearby sources. The SFD of craters on the continuous ejecta blanket should be indicative of the flux of incoming primary impact craters and distant secondary craters since the time of Tycho’s formation. If the cratering rate has been constant and the distribution of craters random and the target materials uniform, then there should be no difference in the regional-scale distribution of craters. However, CSFD measurements used to date individual impact craters have shown that impact melt ponds and flows consistently yield lower relative and absolute ages than the ejecta blanket unit surrounding them (Shoemaker et al., 1968; Hartmann, 1968; Strom and Fielder, 1969; Greeley and Gault, 1971; van der Bogert et al., 2010, 2013; Zanetti et al., 2012, 2013; Hiesinger et al., 2012; Wünnemann et al., 2012; Xiao and Strom, 2012; Krüger et al., 2015). This apparent discrepancy was first observed in Lunar Orbiter images, and was interpreted as evidence for impact-induced volcanism or multiphase development of craters (Hartmann, 1968; Strom and Fielder, 1969; Greeley and Gault, 1971; Young, 1975). The currently favored hypothesis for the CSFD discrepancy is that material property differences between the impact melt and ejecta blanket are different enough that cratering efficiency is affected, resulting in relatively smaller diameter craters on competent crystalline impact melt compared to the relatively unconsolidated, brecciated ejecta for the same assumed impactor parameters (Melosh 1989; Dundas et al., 2010; Wünnemann et al., 2010; Trey et al., 2011; van der Bogert et al., 2010, 2013). Other factors that may affect the measured CSFD on ejecta blankets may be the small range in diameters recorded when counting small areas, topographic or slope effects leading to enhanced erosion, image
resolution (Young et al., 1974; Ostrach et al., 2012), subjective effects of different counters (Xiao and Strom, 2012; Robbins et al., 2014) or external secondary cratering (McEwen and Bierhaus, 2006). An alternative hypothesis for the CSFD discrepancy between melt units and surrounding ejecta is that late-arriving ejecta fragments from the parent crater formation form small craters on the continuous ejecta blanket, immediately following the emplacement of the ejecta curtain. In a process first suggested by Shoemaker et al. (1968), the late-arriving fragments create auto-secondary (or “self-secondary”) craters that are measured in the SFDs of craters on the ejecta blanket, but cannot be distinguished as obvious external-secondary craters. Thus, auto-secondary craters are a population of craters on the continuous ejecta that are not primary craters, but are included in CSFD measurements, resulting in an overestimation of the flux of primary impacts (Plescia and Robinson, 2011; Williams et al., 2014; Zanetti et al., 2013, 2014).

Our objective is to better constrain the ages for several large Copernican craters (e.g., Tycho and Aristarchus) and to examine their CSFDs for small scale geologic processes. As the impact melt and ejecta are deposited nearly simultaneously, they should accumulate the same size-frequency of craters and aside from some statistical variation the ejecta and the melt should be uniformly cratered. In the current study we counted craters and compiled statistics for craters >50 m in diameter on the ejecta blankets of Tycho and Aristarchus Craters, as well as for smaller diameter craters in smaller count areas on impact melt and ejecta. We present evidence from the crater counts and from crater density maps that ejecta blanket units are not uniformly cratered and we favor the hypothesis that ejecta blanket crater populations are overestimated due to the presence of auto-secondary craters. The implications of inflated crater populations and possible formation mechanisms of auto-secondary cratering are discussed.
3.2 Methods

Crater counting is considered a reliable method to estimate the relative age of geologic units on planetary surfaces based on the idea that old surfaces have accumulated more impact craters than more recent ones (Baldwin, 1949; Shoemaker et al., 1962). Crater size-frequency distribution (CSFD) measurements done on mare surfaces and ejecta blankets of impact craters on the Moon have been calibrated on the basis of radiometric or exposure ages of lunar samples. The lunar chronology of Neukum et al. (2001) exhibits an approximately linear relationship over the past ~3 Ga (Neukum et al., 1976, Arvidson et al., 1979; Neukum, 1983, Neukum and Ivanov, 1994; Hartmann et al., 1981; Neukum et al., 2001; Hartmann et al., 2001; Stöffler and Ryder, 2001; Hiesinger et al., 2012). This approximation is based on CSFD measurements of homogenous ejecta blanket units at four geologically young craters (Copernicus, Tycho, North Ray, and Cone) with radiometric or exposure ages from Apollo samples (Hartmann et al., 1969; Neukum et al., 1976; Hartmann et al., 1981; Neukum and Ivanov, 1994; Hiesinger et al., 2012).

Techniques for measuring crater size-frequency distributions on planetary surfaces have been well established (Crater Analysis Working Group, 1979; Hartmann et al., 1981; Neukum and Ivanov, 1994; Stöffler and Ryder, 2000; Hiesinger et al., 2000; Neukum et al., 2001; Hartmann and Neukum, 2001; and references therein). In general, in order to obtain the CSFD we (1) measure the surface area of the unit, and (2) measure the diameters of each primary impact crater within the mapped unit. Ideally, one would only map primary craters; however, as we argue in this work, a significant number of these craters are likely not primary, but morphologic evidence is inadequate to distinguish them from degraded primaries. For the detailed procedure for proper selection of count areas and measuring CSFDs, see Hartmann et al., 1981 and Neukum, 1983.
Our study exclusively uses the Neukum Production Function (NPF) (Neukum, 1983; Neukum and Ivanov, 1994; Neukum et al., 2001) for absolute model age analyses, although the relative crater density results and counting statistics will exist regardless of which system is applied (e.g., Hartmann et al., 1981; Marchi et al., 2006; Robbins, 2014). The NPF uses an 11th order polynomial to estimate the production of craters on the lunar surface from the Nectarian epoch to the present, where the cumulative number of craters accumulated is represented by N, per km\(^2\) with diameters larger than a given value, D. Calibration of the lunar chronology and the derivation of the Neukum Production and Chronology functions can be found in Neukum (1983), Neukum and Ivanov (1994), and Neukum et al. (2001). In this study, we present absolute model ages (AMAs) calculated from N(1) values in the Neukum chronology system, as this is typically chosen as the standard diameter for reporting model ages in the literature, and as a convenient numerical way to show the large variations in crater populations between ejecta blanket units. We are not suggesting that there are millions of years between the emplacement of ejecta blankets and melt ponds. All units on the ejecta blanket of a parent crater should report the same age, and the purpose of this study it to determine why the measured AMAs are different. Craters were counted with the ArcGIS plug-in CraterTools (Kneissl et al., 2010). Crater statistics and absolute model ages (AMA) were compiled in CraterStats 2 (Michael et al., 2010) using production and chronology functions from Neukum et al. (2001).

The continuous ejecta blanket at Aristarchus and Tycho Craters were defined as 1 crater radius from the parent crater rim (Aristarchus crater radius: 21 km; Tycho crater radius: 41 km) (Moore et al., 1974) (Figure 3-1). The count area at Aristarchus Crater consists of the entire continuous ejecta blanket (~5,000 km\(^2\)). Owing to its larger size, the count area at Tycho Crater was subdivided into eastern and western regions of nearly identical size (~3,900 km\(^2\)). All craters
>50 m in diameter within the count areas were counted and included in the counting statistics. Craters were counted on sinusoidal map-projected 0.5 m/pixel LROC-NAC and 100 m/pixel LROC-WAC (Wide angle Camera) images (Robinson et al., 2010) and on 7.4 m/pixel Kaguya Terrain Camera images (Haruyama et al., 2007). High-resolution CSFDs were also measured in equal-sized (2.24 km²) adjacent count areas on impact melt and ejecta at Tycho using NAC images (M1151907706L, R). This method was done to determine if CSFD variation seen in the >50 m diameter range persists to the limit of image resolution. The distribution of impact melt morphologies (melt ponds and flows) at both craters was mapped using LRO-WAC, NAC, and Kaguya TC images overlain on the WAC_GLD100 DTM (Scholten et al., 2011). Smooth, dark regions found in depressions in the ejecta that resemble ponds and smooth and banded terrain resembling lava flows with flow fronts were mapped as impact melt (Zanetti et al., 2012; Krüger et al., 2015).

Areal crater density maps were created using the point density feature in ArcMap 10.1 (Silverman, 1986) in order to examine the distribution of craters irrespective of crater diameter. Center point location was determined from crater rim polygons and used to calculate the point density for a given area. Point density is reported as the magnitude per unit area from point features that fall within a neighborhood around each cell of a raster image. A neighborhood is defined around each cell center of a raster image, and the number of points (i.e., crater centers) that fall within the neighborhood is totaled and divided by the area of the neighborhood (in km²), resulting in the number of craters per km². Figure 1 has a 100 m cell size and 5 km search radius, which has the finest pixel resolution and a search radius that allows for interpretation of broad trends in crater density. The boundaries of the point density map results were buffered by the
size of the neighborhood search radius to remove edge effects at the slight expense of a reduction in study area size (Aristarchus: 3 km edge buffer; Tycho: 5 km edge buffer).

Using the density map to select subareas, CSFDs were measured for four different relative densities of craters on the ejecta blanket. The count areas were set based on the number of craters per unit area in the density data (at Tycho: 0-1, 1-2, 2-3, >3 craters/km$^2$; at Aristarchus 0-2, 2-4, 4-6, and >6 craters/km$^2$), and represent comparable regions of the ejecta blanket, and comparable numbers of craters owing to the scaling of the color bar. All point-density count areas contain a large range in crater diameters.

3.3 Results

3.3.1 Crater Density Mapping At Tycho

The point density data illustrate the spatial variation in the occurrence of craters, irrespective of crater diameter. The density of craters (>50 m diameter) on the ejecta blanket of Tycho Crater varies from 0.3 craters/km$^2$ to 3.7 craters/km$^2$ (Figure 1a, b). In the western ejecta blanket count area, low crater density regions are found in areas closest to the crater rim (within ~5 km) and crater density generally increases with increasing distance. Slightly lower density regions (1.5 craters/km$^2$) in the western count area are associated with a secondary crater chain in the NW (extending from Tycho), and a large post-Tycho primary crater in the SW. In the eastern ejecta blanket count area, the low crater density region is much larger, extending to 20 km from the crater rim. The low density regions throughout the ejecta blanket are associated with areas containing large amounts of ponded impact melt (Fig 1b, areas mapped in black). The highest density of craters on the ejecta blanket occurs in the SE. A lower crater density region is seen in the distal part of the continuous blanket in the SE corner of the count area associated with a large
30 km pre-existing primary crater wall. Topographic slopes in the count areas are generally less than 10º, although slopes >45º are seen associated with crater walls.

### 3.3.2 Crater Density Variation at Aristarchus Crater:

Crater density in the continuous ejecta blanket around Aristarchus displays a similar distribution to that exhibited by Tycho. Crater density varies from 0.5 craters/km² to 9.4 craters/km². Crater density is relatively higher around Aristarchus compared with Tycho owing to the fact that Aristarchus is relatively older than Tycho, based on stratigraphic relationships of distal crater rays (e.g. Wilhelms et al., 1987). The same pattern of increasing crater density with increasing distance from the crater rim is also repeated at Aristarchus. The lowest crater density regions are seen in the east, and are associated with impact melt ponds. An area of low density in the south occurs where the morphology of the ejecta has hillocks and mounds associated with the transition from the continuous to discontinuous ejecta blanket. A low density region in the SW is associated with a 1 km high scarp along the edge of the Aristarchus Plateau. The highest density of craters is found in the NW, on top of the plateau.

### 3.3.3 Ejecta Blanket CSFDs

CSFDs measured in this study are summarized in Table 1. The CSFDs and derived AMAs are shown in Fig 2a and 2b, for the whole ejecta blanket (black isochrons), impact melt ponds (red isochrons), craters on the ejecta blanket > 300 m diameter (blue isochron) and craters >100 m on impact melt ponds (green isochron, Fig 2a). The whole ejecta blanket count area at Aristarchus Crater is 5,000 km² and contains 13,792 craters > 50 m in diameter (diameter range of 50 m to 704 m). If the ejecta blanket is treated as a single unit, an isochron fit over the whole range of craters on the ejecta blanket yields an $N(1) = 1.38 \times 10^{4}$, resulting in an AMA of $164 \pm 1.4$ Ma (Fig 2b). At Tycho Crater the two halves of the ejecta blanket total 7780 km² in area.
(West: 3,910 km²; East: 3870 km²) and contain 15,007 craters (West: 7967 craters; East: 7040 craters) > 50 m in diameter. Summing the two count areas as a single unit, the AMA for Tycho is 69.1 ± 0.55 Ma with the isochron fit between 50 m and 530 m in diameter (Fig 2a).

AMAs were determined for areas of ponded impact melt (black units in Fig 1a, b). Impact melt ponds at Aristarchus cover an area of 79.7 km², containing a total of 165 craters >50 m in diameter, with an N(1) of 7.66x10⁻⁵ and a calculated AMA of 91.4 ± 7 Ma. Impact melt ponds at Tycho cover an area of 304 km², containing 355 craters >50 m in diameter, with an N(1) of 3.46x10⁻⁵ and a calculated AMA of 40.7 ± 2.1 Ma. Craters >300 m diameter on the ejecta blankets of both Tycho and Aristarchus fall below the predicted SFD for the whole ejecta blanket, and were fit with a separate isochron (blue isochrons in Fig. 2a, b). At Aristarchus, seven craters >300 m diameter were fit with an N(1) of 7.0x10⁻⁵ and a calculated AMA of 83.5 ± 31. At Tycho, 22 craters >300 m diameter were fit with an N(1) of 3.46x10⁻⁵ and an AMA of 41.3 ± 8 Ma. Craters >100 m diameter on the melt ponds of Tycho also fell below the predicted SFD based on the production function of Neukum et al. (2001), and were fit with a separate isochron (green isochron, Fig 2a). Craters >100 m in diameter on the melt ponds of Aristarchus also fall slightly below the predicted production isochron but can still be fit by the main melt-pond isochron.

3.3.4 CSFD Based on Density

The density map of craters on the continuous ejecta was used to determine relative AMAs based on crater density. Count areas are shown in Fig 1c and Fig 1d, and the CSFDs and their corresponding AMAs are shown in Fig 3a and 3b and recorded in Table 1. At both parent craters, the lowest density regions (purple units) have lower AMAs compared with highest density regions in the density map. The difference in calculated AMA is approximately a factor of 4 at
both Aristarchus and Tycho. The AMA of the lowest density regions are within error of the AMAs calculated for impact melt ponds at both craters. Whole ejecta blanket AMAs are close to the average value for the density subdivisions, and would plot in the middle of the range of AMAs.

### 3.3.5 High-resolution Melt-Ejecta CSFD comparison:

In order to determine if crater density variations are observed at the smallest scales, CSFDs of craters >3 m diameter were measured in equal-sized (2.24 km$^2$) count areas on the ejecta blanket and an adjacent impact melt pond at Tycho crater (Fig. 4). The Neukum Production Function is only valid for craters >10 m in diameter, so AMAs were calculated with this limitation. On the melt pond, 6795 craters > 3 m diameter were counted, with 189 craters >10 m in diameter, an N(1) of 1.8x10$^{-5}$ and an AMA of 21.5 ± 1.5 Ma. On the adjacent ejecta blanket, 10220 craters >3 m diameter were counted, with 608 craters >10 m diameter, an N(1) of 7.28 x 10$^{-5}$, and an AMA of 86.9 ± 3.5 Ma. Similar to the >50 m diameter results, the difference in AMA is about a factor of 4 between the melt unit and the ejecta blanket.

### 3.4 Discussion

#### 3.4.1 Crater Density Variation on Copernican Ejecta Blankets:

Both the size-frequency distribution and density of craters (irrespective of crater diameter) within the continuous ejecta of Aristarchus and Tycho varies depending on location and lithology of the count area. The following sections compare our results to previous efforts to determine the ages of Aristarchus and Tycho using CSFDs, discuss the leading hypotheses for the source of crater density variation, and the implications variable ejecta blanket CSFDs may have on the utility of small-crater SFDs for dating small surface units on the Moon.
Tycho Crater is used as an anchor point on the lunar cratering chronology based on a cosmic ray exposure age of $109 \pm 1.5$ (Arvidson and Guinness, 1976; Drozd et al., 1977) from Apollo 17 samples. Lucchitta (1972, 1977), Scott and Carr (1972), and others have suggested that distal ejecta from Tycho struck the side of the South Massif of the Taurus-Littrow Valley (2200 km away), triggering the landslide creating the light mantle terrain sampled by Apollo 17 (Wolfe et al., 1981). Recent CSFD measurements from Hiesinger et al. (2012) yielded model ages of $85 \pm 15$ Ma using selected small areas of Tycho ejecta in LRO-NAC frames and $125 \pm 12$ Ma for a large count area in the western continuous ejecta blanket using LRO-WAC. The Hiesinger et al. (2012) crater counts are in good agreement with $N(1)$ estimates from Neukum and König (1976) [$N(1) = 6.0 \pm 1.7 \times 10^{-5}$] and Neukum (1983) [$N(1) = 9.0 \pm 1.8 \times 10^{-5}$].

Our measurements for the whole ejecta blanket (70 Ma; Figure 2a) of Tycho crater give model ages that are less than both the small area LRO-NAC counts and large area LRO-WAC count of Hiesinger et al. (2012). However, our measurements of the highest crater density regions (125 Ma; Figure 3a) match well their LRO-WAC measurements and our mid-high crater density region measurements (87 Ma; Figure 3a), as well as the high resolution ejecta blanket measurement (87 Ma; Figure 4b), match well their LRO-NAC conclusions. The discrepancy between our results and those of Hiesinger et al. (2012) may be due to poor statistics derived from the WAC images (although Hiesinger et al., 2012 report 266 craters measureable at LRO-WAC resolution of 100m/pixel) and the small count area sizes on NAC images, or the areas they selected simply had high crater densities (in a statistical sense). We find approximately 35 craters larger than 300 m diameter that would be visible at WAC resolution in our Tycho ejecta blanket count areas, which correspond to an AMA of $41.3 \pm 8$ Ma. Our Tycho impact melt-pond counts
(26-41 Ma, Figure 2a; 22 Ma, Figure 4b) are similar to Hiesinger et al. (2012) melt pond counts (their TM1-5 count areas; avg: 32 ± 2 Ma) and Tycho floor counts (33 ± 5 Ma).

Aristarchus Crater does not have a sample age estimate and has only been dated using CSFD measurements. Aristarchus is relatively older than Tycho owing to bright ray superposition relationships and the relative crater populations (Wilhelms et al., 1987). CSFD measurements for Aristarchus generally place its age at less than 200 Ma (130 – 180 Ma, König and Neukum, 1976; 150 Ma, Young, 1975; 189 Ma, Zanetti et al., 2013). Our current age estimate, using the same techniques and production function, is comparable to these previous estimates. Strom and Fielder (1969) estimated an age of less than 1 Ga prior to the development of a well-calibrated lunar chronology, and were the first to document a discrepancy in crater counts on ejecta deposits and flow units near the rim.

Our CSFD measurements at Aristarchus and Tycho are some of the largest, most complete counts done on lunar crater ejecta blankets, and cover the widest range of crater diameters of any previous study. Although age estimates for the formation of the parent craters are “model ages” and production-function dependent, the count areas and measured diameter data are an accurate catalog of the size-frequency distribution of craters on the ejecta blankets.

3.4.2 **Crater Density and Apparent Age**

Crater density maps show that the density of craters on the ejecta is locally highly variable, and that AMAs derived from different areas around the ejecta can give ages that vary by a factor of 4. Areas near the parent crater rim exhibit the lowest crater density, and density gradually increases with increasing distance from the rim (Figure 1). Topographic slopes play a minor role in the density variation of craters within the continuous ejecta blanket at Tycho and
Aristarchus, which may be due to the removal of small craters through down-slope erosion. At Aristarchus Crater, the radially oriented, low-density region in the SW could be due to down-slope erosion along the 1 km high scarp of the Aristarchus Plateau. However, we do not suspect that topography at the scale of the counted craters is responsible for erasing craters at areas near the crater rim. Typical slopes within the continuous ejecta blanket measured on LRO-WAC stereo digital elevation models (100m/pixel) <8°, and are not suspected to completely erode craters >50m diameter on timescales estimated for the formation of Aristarchus and Tycho (i.e. <300Ma). The lifetimes of craters based on crater diffusion rates from Fassett et al. (2014) suggest that a 50 m diameter crater would erode to 1% of its initial depth with ~400Ma, and craters 100 m diameter erode in ~1.7 Ga. Although some 50 m craters may be lost to erosion, on both the melt and ejecta, the relative density difference in larger crater diameters (>100m) persists (Figure 2).

The calculated age of a surface unit depends on the size-frequency distribution of craters, so it is not surprising that subdividing the count areas based on crater density yields relatively older ages for high-density areas, and relatively younger ages for low-density units. However, the factor of 4 differences between the lowest and highest density regions at both parent craters merits consideration. Areas in high density regions have either received 4 times more impacts than areas close to the crater rim, or areas close to the rim have received (or have recorded) 4 times fewer impacts. Evidence for the enhanced production of craters on the ejecta blankets compared to the melt ponds is also seen in the high resolution counts on NAC images (Fig 4). Again, the ejecta blanket has a factor of 4 apparent increase in crater production compared to the adjacent impact melt surface.
Previous studies into the effects of rocket exhaust on lunar soil have attempted to measure the spatial extent of the disturbed area as well as determine what physical changes occurred in the regolith to contribute to changes in reflectance (see Kreslavsky and Shkuratov, 2001; Kreslavsky and Shkuratov, 2003; Kaydash et al., 2011; Kaydash and Shkuratov, 2012; Clegg et al., 2014a. In the following sections we compare the Chang'e-3 blast zone with those of historic sites and discuss similarities and differences in reflectance changes and the processes that led to these changes.

There are a number of possible explanations for the observed variation including: secondary cratering from distal sources, auto-secondary cratering, and/or target properties. Any hypothesis for the source of CSFD variation on ejecta must provide an explanation for the following observations: 1) CSFDs and corresponding AMAs on impact melt ponds are significantly lower (by a factor of 4) than ejecta blanket units (at both >50 m diameter, and >3 m diameter), and 2) the population of craters, irrespective of crater diameter, is higher on ejecta blankets than on impact melt ponds (and surfaces near the crater rim).

3.4.3 **Over-producing Small Craters or Under-representing Large Craters?**

The whole ejecta blanket CSFDs at both Aristarchus and Tycho have atypical curves for cumulative crater plots (Fig. 3.2). Craters >300 m diameter in the ejecta do not fall on the same isochron as the rest of the data and in both study regions can be fit by an isochron that is approximately a factor of 2 less the fit for the whole data range. One interpretation of these data is that craters in this size range are being under-produced, or are under-represented relative to the expected production by the NPF. Either large craters never formed, or some process has preferentially removed larger craters but not smaller craters. We consider this possibility to be unlikely.
Alternatively, smaller craters (50 m – 250 m diameter) could be over-produced in the ejecta blanket compared to >300 m diameter craters. The isochrons fit to the >300 m diameter data correlate very well with the fit to impact melt ponds. The emplacement of impact melt ponds and flows occurs during the last stages of ejecta emplacement and crater modification (Melosh, 1989); therefore it may be safe to assume that nearly all craters forming on impact melt ponds are primary craters. If melt pond crater production is the primary production rate, then craters >300 m also provide evidence for primary impactor cratering rates. Craters in the 50 m – 250 m diameter range would therefore be over-produced relative to the primary impact cratering rate observed on impact melt ponds and inferred for >300 m diameter craters.

3.4.4 Target Material Properties versus Auto-secondary Cratering:

3.4.4.1 Target Material Property Effects

Target material property differences between impact melt ponds and ejecta blanket units are important to consider when determining for the formation of small craters (<300m) on the Moon (Schultz and Spencer, 1979; Melosh, 1989, Wünnemann et al., 2010; van der Bogert et al., 2010, 2013; Hiesinger et al., 2012; Dundas et al., 2012). Because target properties result in smaller crater diameters on melt surfaces relative to ejecta surfaces, it is possible that many craters on the melt surfaces were not recorded in the counts because they fell below the 50 m cut-off diameter for counting. If this were the case, we would expect that the high-resolution count on NAC images to show a similar number of craters (albeit with smaller diameters) in the melt unit compared to the ejecta blanket. However, we measured nearly 3,500 more craters on the ejecta blanket at NAC scale compared to the melt unit (Fig 4) and reproduced the approximate factor of 4 difference between melt and ejecta units, suggesting that if there is a counting bias in
>50 m diameter counts it is negligible. Therefore, it is unlikely that target properties can explain the population differences (irrespective of crater diameter) observed in the crater density maps and high-resolution NAC counts. Although the crater population discrepancy between the melt and ejecta is difficult to explain solely with target property differences, it may be possible that the apparent difference in AMA can be resolved by correcting for the change in crater diameter. The competency of crystalline impact melt can result in craters on melt that may be 7% - 20% smaller in diameter compared to less competent ejecta units (van der Bogert et al., 2012; Dundas et al., 2012), which would make melt surfaces appear apparently younger than ejecta. Measurements of CSFDs on impact melt units and ejecta blankets at Jackson Crater (van der Bogert et al., 2010) suggest that increasing the diameter of craters on impact melt units by 20% can reproduce the AMA measured on the ejecta blankets, and that a correction factor may be enough to explain the observed discrepancies. In the case of our measurements from Aristarchus and Tycho adding 20% to the diameter of craters on the melt ponds does not overcome the apparent AMA variation. For example, adding 20% to the melt pond crater diameters measured in the high-resolution measurements at Tycho (Fig 4) would only increase the apparent AMA to 36.6 ± 2.1 Ma. Target property differences no doubt influence CSFDs on melt units by affecting final crater diameter, but may not be the primary explanation for the crater density variation seen on the ejecta blankets of Aristarchus and Tycho. For example, craters >100 m diameter on the impact melt ponds at Tycho (Fig 2a, green isochron) fall slightly below the predicted production, indicating that smaller crater diameters are being produced than expected. While not as pronounced, craters >100 m on melt also fall below the main melt isochron at Aristarchus (Fig. 2b, red isochron).
3.4.4.2 Auto-Secondary Cratering on Continuous Ejecta

The radial and circumferential variability of crater density on the ejecta blanket suggests that the variability in the CSFDs could be related to the formation of the parent crater. Auto-secondary cratering as part of the ejecta emplacement process provides a testable model that can account for the apparent differences in CSFDs and crater population differences. Continuous ejecta blankets have been shown in experiments and numerical simulations of simple bowl-shaped craters to be emplaced as a contiguous curtain of material, i.e., the ejecta curtain (Gault, 1970, Oberbeck, 1979, Shoemaker, 1963, Melosh, 1989). As the ejected material lands it transfers kinetic energy into the target surface, which rips up and entrains the target in a process known as ballistic sedimentation (Oberbeck, 1979), which is the general explanation as to why the ejecta blanket is smoothed and commonly exhibits radial grooves. Ejecta blanket thickness decays approximately exponentially with distance from the parent crater rim (e.g. McGetchin et al., 1973), and ballistic sedimentation is shown to entrain more target materials with greater distance from the parent rim (e.g., Hörz et al., 1983). Collectively both these processes act to resurface the continuous ejecta blanket and remove pre-existing small impact craters.

Impact melt flows and ponds are common morphologies observed near the rims of Copernican-aged craters (e.g., Howard and Wilshire, 1968; Hawke and Head, 1979; Bray et al., 2010; Krüger, 2015). The stratigraphic position of melt flows on top of ejecta deposits suggests that melt flows and ponds are emplaced at some (short) time after the emplacement of the ejecta curtain and ballistic sedimentation have taken place, although the duration of melt emplacement is not well constrained. Flows originating at crater rims can extend tens of kilometers radially away from the parent crater, and are observed to cascade from pond to pond down the topography of the ejecta blanket (e.g., eastern melt ponds at Tycho). Melt can also be
volumetrically significant requiring time to solidify. The arrival of melt flows and ponds following the emplacement of the ejecta curtain may take several minutes to days to come to its final resting position. The emplacement of melt is the final stage in the ejecta emplacement and modification process.

Shoemaker et al. (1968) observed and described the crater density variations between melt and ejecta at Tycho in Surveyor 7 images. He proposed that fragments of ejecta from the parent crater could land within the continuous ejecta blanket and form craters, and outlined a scenario similar to what is described below. However, little consideration has been given to auto-secondary cratering as part of the ejecta emplacement process since the late 1960’s. The work of Shoemaker et al. (1968), observations of crater distributions around Giordano Bruno (e.g., Plescia and Robinson, 2011; Williams et al., 2014), and our previous work (Zanetti et al., 2013; 2014) suggest a scenario in which auto-secondary fragments impact the ejecta blanket after the ejecta curtain and ballistic sedimentation has occurred and the bulk of the continuous ejecta blanket has been emplaced providing a hypothetical mechanism for producing craters on the continuous ejecta of Copernican craters that are not primary impacts. However, the formation of auto-secondary craters must occur prior to the arrival of melt ponds to explain the observed discrepancy between ejecta and melt-unit crater populations. As impact melt flows over the ejecta blanket, it will flood areas that have experienced auto-secondary cratering and resurface these areas of the ejecta blanket. When the impact melt comes to rest and crystallizes, it will then begin to record the primary flux of impacts.

3.4.5 Evidence for Auto-Secondary Cratering:

The addition of an auto-secondary crater population on the ejecta blanket and subsequent obscuring of auto-secondary craters near the rim by late-arriving melt can explain the correlation
of impact melt ponds with low crater density regions near the crater rim. Additionally, the factor-of-4 difference in cratering rate between the ejecta and melt may be explained by auto-secondary cratering and provides an estimate of the amount of auto-secondary cratering on the ejecta. Because impact melt ponds are the last emplaced units on the ejecta blanket and are most likely free from auto-secondary craters, the cratering rate on melt is the best approximation of the primary cratering rate since the formation of the parent crater. The good agreement between the cratering rates for melt ponds and >300 m craters on the ejecta (Fig 2) suggests that auto-secondary craters may be abundant in the <250 m diameter range, if we assume that craters >300 m diameter are primary impacts.

The possible embayment of small craters on the ejecta blankets by impact melt has been noted at Giordano Bruno (Plescia and Robinson, 2011; Williams et al., 2014), Necho, Aristarchus, and Tycho (Zanetti et al., 2013; 2014). Additionally, we observe putative “ghost” craters in large impact melt ponds at Tycho Crater (Fig. 5). The circular features are small (<50 m diameter) with slightly raised rims that appear like a crater that has been completely filled in. In order for these craters to be infilled by melt, they must have existed on the ejecta prior to melt arrival. The ghost craters we have identified are large enough in diameter and have raised rims high enough to persist after being flooded by meters thick impact melt, thus they are rare and we have observed only 6 putative ghost craters within the melt ponds in our count area at Tycho. If smaller auto-secondary craters existed, they would have been completely covered by melt without leaving any remnant rims. Despite their rarity, the ghost craters provide morphologic evidence for craters formed on the ejecta blanket prior to the arrival of impact melt flows.
3.4.6 **Models for the Formation of Auto-Secondary Craters**

The formation of auto-secondary craters is conceptually simple. Fragments of ejecta are lofted during the excavation and ejection stages of the impact process and impact the planetary surface after the ejecta curtain is emplaced. However, the timing of events and physical constraints for the origin of the ejecta fragments and emplacement of auto-secondary craters are problematic. The simplest manner for impacts to occur within the continuous ejecta blanket requires fragments to be launched at very high-angles and below the planet’s escape velocity (e.g., for the Moon ~2.4 km/sec), be aloft long enough for the ejecta blanket to be emplaced, and impact the surface with enough velocity to create a crater. Using a simple ballistic range equation, fragments from near the center of the parent crater at just below the escape velocity would need to be launched at angles >85º in order to land within one crater radius of the parent crater rim, and would allow for tens of minutes for the ejecta blanket to be emplaced. Unfortunately, such high angle ejecta are not expected from hydrocode simulations of complex crater formation (e.g., Melosh, 1989; Collins et al., 2007). High-angle ejecta fragments may arise from collisions between particles within the ejecta curtain, and evidence for collision and fragments overtaking one another are seen at Meteor Crater (e.g. Shoemaker, 1963; Shoemaker et al., 1968), but it is not known if the arrival of fragments can be delayed long enough for the ejecta curtain and ballistic sedimentation to resurface the target to produce a distinct crater upon impact.

Impact cratering experiments have produced ejecta fragments with ejection angles of ~70º under certain conditions (e.g., Schultz et al., 2007), but the extremely high angles required to re-impact within the continuous ejecta blanket have not been observed. Ejection of fragments at lower velocities (as slow ~300 m/s to still be able to produce craters on the ejecta blanket)
allows for a reduction in ejection angle, but slower moving fragments are generally considered to be ejected at lower angles from areas closer to the crater rim, reducing the amount of time the fragments can be aloft for the ejecta blanket to be emplaced. Some laboratory impact experiments have been reported to produce a column of material ejected at a late-stage during the opening of a small impact craters nearly normal to the target surface (Charters and Summers, 1959; Shoemaker et al., 1968; Gault et al., 1970), but the precise mechanism is not understood.

Alternatively, fragments may be ejected from the crater at some point after the main phase of ejecta, either related to the formation of the central peak or the emplacement of impact melt on the crater rims. Shoemaker et al. (1968) suggested that fragments might be ejected at a fairly late stage and at high or nearly vertical ejection angles from the region that becomes the central peak. This remains a possibility to explain auto-secondary cratering around Aristarchus and Tycho (both central peak craters) but cannot explain the variable crater density around simple craters (e.g. observations at G. Bruno Crater by Plescia and Robinson, 2011; Williams et al., 2014; and Cone Crater, Hiesinger et al., 2015). A multi-stage ejection hypothesis described by Osinski et al. (2011) suggests that melt and debris from the floor of the crater can be forced up the walls and out of the crater by the rapid rise of the central peak, but it is unclear at what velocity the melt and any accompanying fragments might be travelling, or if the process even occurs. Another possibility may be that the fragments originated as part of the parent-crater-forming projectile. When the rarefaction wave passes through the projectile during the contact stage of the impact process it may be possible to spall off material that is ejected at high angles. Projectile spall is thought to be the source of small fragments of meteoritic material in suevite fallback deposits at the Ries Crater (Hörz et al., 1983; Osinski et al., 2011), but there is no evidence that fragments large enough to create 50 m – 250 m diameter craters can be created by
this process. Determining if auto-secondary fragments can be formed as part of the excavation and ejection process warrants further study.

3.4.7 **Implications of Auto-Secondary Cratering on Ejecta Blankets:**

Auto-secondary craters appear as normal, small craters on the ejecta blanket and do not appear in chains and cluster like traditional secondary craters, making it impossible to distinguish them from true primaries based on morphometry or context. Therefore, CSFDs measured on the ejecta blankets of Copernican-aged craters have included the population of auto-secondary craters, and in turn, over-estimate the true production of craters since the parent crater formation. It has been reported that all craters that have been used to calibrate the lunar chronology curve (e.g., Copernicus, Tycho, Cone, North Ray) exhibit some radial variation in crater frequency on their respective ejecta blankets (e.g., Plescia and Robinson, 2011; Hiesinger et al., 2012; Robbins, 2014; Hiesinger et al., 2015), suggesting that all suffer from auto-secondary contamination. Consequently, the lunar cratering rate of small craters (<250 m) on the ejecta blankets of Copernican craters could be over-estimated by a factor of 4 based on our measurements at Tycho and Aristarchus. (the difference in cratering between impact melt ponds, which are recording mostly primary impacts and craters <250 m diameter on ejecta blankets). The cratering rate in the very recent history of Mars as observed by HiRise also suggests that primary small-crater production is over-estimated in the Neukum and other production functions (also by a factor of ~4), although estimating current cratering rates has a number of uncertainties (Dauber et al., 2013). Taken together, it may be the case that the lunar chronology curve should be re-calculated based on the production of craters on melt surfaces (despite probable target property influences) and the cratering rate of >300 m diameter craters on ejecta blankets. A
change in this manner would result in surfaces with few craters being relatively older than the current chronology would estimate. For example, if we assume the cratering rate of Tycho melt ponds represents 109 ± 1.5 Ma of exposure to primary impacts (based on cosmic ray exposure ages of Arvidson and Guinness, 1977; Drodz et al., 1977), then the age of Aristarchus would be ~280 Ma, significantly older than current estimates (e.g. 130-180 Ma; Koenig and Neukum, 1976; 164 ± 1.4 Ma, this work). Similarly, absolute model ages for small volcanic surfaces (e.g., IMPs; Braden et al., 2014) would be older by a factor of 2 – 4.

3.5 Conclusions

We measured the distribution of small craters on the continuous ejecta of large Copernican craters Tycho and Aristarchus and found that crater size-frequency distributions and crater density (irrespective of crater diameter) varies with location in the ejecta blanket and target lithology (impact melt versus ejecta blankets). We interpret these differences to result from auto-secondary craters that formed on the continuous ejecta blankets of impact craters by late-arriving ejecta fragments from the formation of the parent crater. Observations of impact craters on the ejecta blanket embayed by melt and “ghost” craters in impact melt support the existence of auto-secondary craters. N(1) values for impact-melt-pond crater populations are very similar to those of >300 m diameter craters, and we infer that impact melt ponds and >300 m diameter craters likely record the flux of primary craters. If this is the case, a new estimate for the primary flux of craters in the last ~110 Ma is: N(1) = 3.4x10^{-5} (craters with diameter >1 km/km^2/yr).

The formation mechanism of putative auto-secondary craters is not at all understood. Theoretical and numerical models do not account for these small features and more work should be done to evaluate possible modes of origin of the fragments and their emplacement. Regardless, it is likely that auto-secondary craters have been included in lunar chronology
calibration counts on the continuous ejecta blanket of Tycho and other craters, resulting in absolute model ages that currently under-estimate the ages of features by a factor of 2 – 4.

Acknowledgements

This study was funded by NASA PGG grant NNX13AM88G (BLJ), LRO contract NNG07EK00C via GSFC and ASU (BLJ), and a McDonnell Center for the Space Sciences Graduate Fellowship (MZ). The map of impact melt distribution at Tycho was supplied by Tim Krüger (Krüger et al., 2015).
References

48. Strom, R. G., & Fielder, G. 1970. Multiphase eruptions associated with the craters Tycho and Aristarchus. PLI communication #150
Figure 3-1: Crater population density maps of the continuous ejecta blankets of A) Aristarchus, and B) Tycho. Purple regions correspond to low crater density, red areas are high crater density. Impact melt pond distribution is mapped in black, and corresponds to areas of low crater density. C) and D), subdivided crater count areas from the density results. CSFD and AMA results from these areas are given in Fig 3.3.
Figure 3-2: Absolute model ages derived from whole-area counts at Tycho (a) and Aristarchus (b). Black line in both a and b represents the whole ejecta blanket count AMA isochron. Red isochrons represent AMA for impact melt ponds only. Green isochron in 3.2a represents AMA for craters >100 m diameter on impact melt ponds, which fall slightly below the total melt isochron (red line). Blue isochrons represent a separate fit to >300 m diameter craters in the continuous ejecta blanket at each crater.
Figure 3-3: Crater size-frequency distribution plots for density subdivision count areas in Figure 1c, d. Colors correspond to count area and crater density in Fig 3.1a, b (red = high, yellow = mid-high, blue = mid-low, purple = low). Low density regions have low absolute model ages, and AMA values for low density regions correlate well with impact melt pond only AMA results (Fig 2a, b).
Figure 3-4: A) LROC-NAC resolution count areas on impact melt and ejecta at Tycho Crater (NAC image pair: M1151907706). B) Crater Size-Frequency Distributions for impact melt (red) and ejecta blanket (black). Count areas are each 2.24 km$^2$. C) Location and numbers of individual craters counted to the limit of resolution (0.5 m/pixel) within their respective count areas. The impact melt contains 6,795 counted craters; the ejecta blanket contains 10,220 counted craters.
Figure 3-5: a) Impact melt pond on the continuous ejecta blanket of Tycho Crater containing ghost craters. B) 45 m diameter ghost crater (center) with radial fractures. Also visible a 100 m diameter fresh crater that post-dates pond formation. C) 50 m ghost crater found in same impact melt pond. (LROC-NAC M150578086RE)
### Tables

Table 3-1: Results of CSFD measurements for Tycho and Aristarchus. N(1) and AMA calculated using production and chronology functions of Neukum et al. (2001).

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<th>Parent Crater</th>
<th>Region</th>
<th>Area (km²)</th>
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<th>N(1)</th>
<th>Absolute Model Age (Ma)</th>
<th>AMA error</th>
<th>Isochron Fit Range (m)</th>
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<td></td>
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<tr>
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<td>7967</td>
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<td>72.4</td>
<td>1.5</td>
<td>50-550</td>
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<td></td>
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Chapter 4: Decomposition of Zircon in Mistastin Lake Impact Melt Glass: Results from an Integrated Multi-Instrument Study

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Abstract

Using an integrated multi-instrument approach we investigated two exceptionally well-preserved zircon grains in an obsidian-like (holohyaline) impact melt glass from the 28 km diameter Mistastin Lake Impact Structure (Labrador, Canada). The zircon grains contain a relict zircon core and a 20–50 µm thick, quenched decomposition rim, formed by the dissolution reaction of zircon to ZrO₂ plus SiO₂ phases when the grains were entrained in impact melt during the impact event ~36 Ma ±4 Myr ago. Using electron-probe microanalysis (EPMA), laser Raman spectroscopy (LRS), hyperspectral imaging cathod-luminescence (CL), secondary ion mass spectrometry (SIMS), and electron backscatter diffraction (EBSD), we determined the composition and phases present in the zircon core, decomposition rim, and surrounding glass, and we investigated the rim-core dissolution interface, correlating REE zoning with hyperspectral CL in the zircon core. We find that within the Mistastin zircon grains studied here, the decomposition of ZrSiO₄ → ZrO₂ + SiO₂ was induced by entrainment within a superheated (>1687°C) impact melt, and have not experienced high levels of shock. Within the decomposition rim we find no crystalline phases of SiO₂, but do find evidence of preserved metastable tetragonal ZrO₂. Additionally, there has been a diffusive exchange of material within the decomposition rim, with impact melt glass infiltrating behind the decomposition front, and Zr
diffusing into the surrounding glass. The cores of the zircons have remained essentially unmodified, and show relict magmatic zoning in CL, and have essentially undisturbed U-Pb ages (although a small amount of Pb may have occurred). A distinct band of “blue” CL is observed at the decomposition interface, and along melt filled fractures in the grains; probably related to incipient breakdown of the zircon core. Trace element analyses along a traverse of the zircon show an inverse correlation of REEs and Y with CL intensity. We suggest that CL intensity may be controlled by the ratio of Y to other activators (e.g. Er and Dy), rather than absolute concentrations of activators.

4.1 Introduction

The Mistastin Lake impact structure is a ~28 km diameter structure located in northern Labrador, Canada (55°53’N; 63°18’W). The crater formed in the late Eocene, likely by the impact of a ~1.5 km stony meteorite [meteoritic material has yet to be identified (Marion, 2009)]. The impact has been dated to ~36 Ma [36 ± 4 Ma (Ar-Ar: Mak et al. (1976); recalculated by Grieve (2006)); 35.8 ± 1.0 Ma (U-Th-He LA-ICPMS: Young et al. (2015))]. The structure contains a large, 16 km diameter lake with two small, centrally located islands (Horseshoe Island and Bullseye Island), interpreted as remnants of the crater’s central uplift. The target rocks are part of a series of intrusive bodies related to the Mesoproterozoic Mistastin Batholith, and are predominantly granodiorite, with bands of anorthosite and mangerite that extend northwest-southeast through the impact site (Fig. 1) (e.g., Currie, 1971; Grieve 1975; Marion and Sylvester, 2010). The mangerite rocks at Mistastin are a pyroxene-rich, quartz-bearing monzonite, characterized by deep green colored quartz, and remarkable rapakivi textured potassium feldspar (Marion, 2010). The anorthosite rocks are principally composed of labradorite to andesine compositions, with up to 10% pyroxene (Marion, 2010). The labradorite has mostly lost its
characteristic labraorescence luster, likely due to fracturing and shock effects (Marion, 2010). The granodiorite rocks are coarse-grained hornblende-biotite granodiorite, consisting of 20 – 25 vol% K-spar, 30 vol% plag, 20 vol % qtz, 5-15 vol% hornblende, and 5 vol% biotite; and also display rapakivi textures (Marion, 2010). Currie (1971) estimated that the area within the crater rim is ~77% granodiorite, 12% anorthosite, and 11% mangerite.

U-Pb dating of zircons in the major target rocks from outside the impact area was done by LA-ICPMS by Marion and Sylvester (2010), who reported ages of 1451 ± 12 Ma (Mangerite), 1440 ± 13 Ma (Granodiorite Gneiss), 1438 ± 9 Ma (Anorthosite), and 1429 ± 10 Ma (Granodiorite). Three impact melt locations also contained zircons that were dated by Marion et al. (2010) at 1435 ± 13 Ma (CM023, Cote Creek); 1431 ± 21 Ma (CM005, Steep Creek); and 1413 ± 13 Ma (CM065; South Ridge). Marchland and Crocket (1977) measured the age of the mangerite at 1409 ± 57 Ma, and 1318 ± 17 Ma for the granodiorite by Rb/Sr dating.

Impact melt outcrops are found circumferentially around the lake, and can have heterogeneous compositions depending on sampling location (Marion and Sylvester, 2010). The largest impact melt unit is Discovery Hill, a prominent butte located within the western crater wall, which features a ~60 – 80 m thick crystalline melt deposit that displays two tiers of large, vertical, columnar jointing. The melt unit at Discovery Hill contains meter to decameter-sized boulders of various target rocks (most commonly mangerite), around which curved columnar jointing occurs. Initially proposed as an erosional remnant of the crater floor melt sheet, recent detailed mapping has identified listric faults bounding the melt deposit. Tornabene et al. (2012) proposed that Discovery Hill is a remnant terrace melt pond and that its morphology and stratigraphic position are consistent with lunar impact melt deposits, such as those observed at Aristarchus Crater (e.g. Chapter 2, this thesis).
During an expedition to the Mistastin Lake impact structure in August of 2011, a fist-sized sample of glassy, quenched impact melt was collected at the top of Discovery Hill. The sample, shown in Figure 2, is an impact melt rock with a completely glassy matrix, containing sub-rounded mineral clasts. In hand-sample the rock resembles obsidian glass, dark brown to black in color, and has conchoidal fractures. The glass is solid, and contains no macroscopically visible vesicles. The sample was recovered as a float sample from the upper surface of Discovery Hill, adjacent to an outcrop face containing very large boulders of mangerite target rocks embedded in the crystalline impact melt. No other similar samples were found in the immediate vicinity. However, numerous reports of similar stones have been reportedly sourced from Discovery Hill, and the native Innu people have used them for stone tools. It is claimed they are used in spiritual practices (Marion, 2009).

Upon examination of thin sections made from the sample, two exceptionally well-preserved zircon grains containing vermicular rim textures were found in the impact glass. These reaction rims are indicative of zircon decomposition due to high temperature and result from the breakdown of zircon to a zirconium-bearing phase (El-Goresy; 1965). The zircon grains, labeled MZRN-1 and MZRN-2, are shown in cross-polarized light images in Figure 3, and in backscattered electron images in Figure 4. MZRN-1 is an oval shaped grain (~400 μm x ~200 μm) and contains a ~40 μm thick decomposition rim. MZRN-2 has a circular shape (~100 μm diameter), with a ~20 μm reaction rim.

4.1.1 Zircon Decomposition

The transition from ZrSiO$_4$ (zircon) to ZrO$_2$ (zirconium dioxide) has a characteristic reaction texture, which has a vermicular or wormy appearance, and the occurrence of this texture can be used as a temperature constraint for impact melts (e.g., El Goresy, 1965; Kleinmann
Breakdown of ZrSiO$_4$ (zircon) to monoclinic ZrO$_2$ (baddeleyite) and tetragonal ZrO$_2$ and amorphous SiO$_2$ can occur under ambient pressure conditions (e.g. Butterman and Foster, 1967). The breakdown reaction is $\text{ZrSiO}_4 \rightarrow \text{ZrO}_2 + \text{SiO}_2$. Depending on how rapidly the zircon-ZrO$_2$ assemblage is quenched in the surrounding impact melt, ZrO$_2$ can be preserved or it reacts with SiO$_2$ from the impact melt to form granular-textured zircon grains (Bohor et al., 1993; Wittmann et al., 2006).

The phase diagram for the ZrO$_2$–SiO$_2$ system is shown in Fig. 5 (modified from Kaiser et al. (2008)). The breakdown reaction ostensibly begins when zircon grains are heated above 1687°C, although impurities in the grains can lower the reaction temperature (Kaiser et al., 2008). Nucleation of ZrO$_2$ precipitates were shown by Kaiser et al. (2008) to begin as low as 1660°C, but a liquid phase containing SiO$_2$ and growth of dendritic ZrO$_2$ are not seen until >1680°C. The solid-state reaction proceeds through the formation of various metastable phases with a gradual increase in SiO$_2$. At temperatures higher than 1687°C, Kaiser et al. (2008) described the dissociation and breakdown of the zircon lattice structure as follows:

“Because of the smaller ionic radius of Si$^{4+}$ (0.26Å) the silicon shows higher diffusion velocities compared to the larger Zr$^{4+}$ (0.72Å) which prefers to maintain its eightfold coordination from the ZrSiO$_4$- structure in the tetragonal ZrO$_2$-structure, which is the stable ZrO$_2$ modification in the temperature range of 1650–1700°C. As a consequence, the ZrO$_8$-coordination polyhedra in the ZrSiO$_4$-structure, which are interconnected by SiO$_4$-tetraedra along the c-axis, become separated by the outward diffusion of the Si$^{4+}$ (as [SiO$_5$]$^{5-}$) while they start forming zirconia-units in b-direction. As interpreted before, the ZrO$_8$-units try to expand with increasing temperature and are thus under compressive stresses if there are still SiO$_4$-tetrahedra in the chain. Accordingly, the release of SiO$_2$ results in a relaxation of the structure,
In short, $i^{4+}$ ions can move faster through the zircon lattice compared with larger $Zr^{4+}$ ions. As the $Si^{4+}$ ions break free, $ZrO_8$ coordination polyhedra that make up zircon lattice are separated. As temperature increases the $ZrO_8$ units tend to expand, but they are held under compression by any remaining $SiO_4$ tetrahedra. As a result, the $ZrO_8$ polyhedra release $SiO_2$ to relax the structure, and $ZrO_2$ is a product of the rearrangement of the lattice. The product of this breakdown can be viewed as chains of interconnected $ZrO_2$ molecules and the release of amorphous $SiO_2$. The released $SiO_2$ can be reabsorbed if cooling is slow, resulting in the re-formation of $ZrSiO_4$, or in the case of fast quenching, preserved as amorphous silica glass in areas interstitial to the $ZrO_2$ dendrites.

The temperature at which this reaction takes place is sensitive to impurities and grain size and surface area effects. As a larger surface area is created through the release of $SiO_2$, lattice defects are generated, resulting in shorter and easier paths for the release of silica. Any local impurities, such as small inclusions of minerals (e.g., rutile, magnetite, spinels) in the zircon lattice, or ions of $Hf^{4+}$ or $Th^{4+}$ will lower the decomposition temperature (Kaiser et al., 2008). As the reaction progresses and more impurities are released from the zircon (or introduced from the melt), this can have the effect of further lowering the decomposition temperature and accelerating the dissolution rate.

4.1.2 Objectives

The exceptional preservation state of the zircon grains allow us to characterize the decomposition of a natural sample of zircon in an impact melt that was exposed to high heat and quickly quenched. We want to address the following questions:

1) Which mineral phases are present in the grains; including the core, decomposition rim, and surrounding glass?
2) What are the compositions of the glass phases in the decomposition rim and surrounding material, and what is the fate of SiO$_2$ released during decomposition?

3) What pressures and temperatures have the grains experienced, and can the temperature and viscosity of the impact melt be estimated?

4) What is the age and provenance of the zircons, and can an independent age for the Mistastin Lake impact structure formation be determined from the decomposition rim?

5) What is the cathodoluminescence (CL) signature of the decomposed zircons, and can CL be related to decomposition features?

6) Is compositional zoning preserved in the relict core, and what is the relationship between trace element concentration and CL intensity?

We used state-of-the-art analytical techniques to address these questions. Laser Raman spectroscopy (LRS) and electron back-scatter diffraction (EBSD; results reported in Timms et al., 2015, in preparation) were used to identify mineral phases and their spatial distributions, and to characterize crystallographic orientations. Compositions of the impact melt glass, interstitial decomposition rim material, and trace element analyses were done by electron probe microanalysis (EPMA), which also provides petrological context. U-Pb age dating of the zircon core and decomposition rim was done by secondary ion mass spectrometry (SIMS). Hyperspectral-cathodoluminescence (CL) spectroscopy was used to investigate CL activation mechanisms related to the decomposition process.

4.2 Methods

4.2.1 Sample recovery and Microscopic Petrography

The fist-sized sample of holohyaline impact melt glass was recovered from the top of the Discovery Hill outcrop at Mistastin Lake during an expedition to the impact structure in August, 2011. Eleven polished thin sections were prepared from the sample, of which two contained the zircon grains analyzed in this study. The zircon grains were cut from the thin-sections, still
mounted on glass slides with epoxy, and re-mounted in a 1”-round section. Optical analysis was done before re-mounting, and included polarized light and reflected light microscopy at Washington University in St Louis. The sample was carbon-coated for EPMA and Hyperspectral CL analyses, after which the carbon coat was removed and the sample was coated with gold for SIMS analysis. Following SIMS analyses, the sample was re-polished for EBSD analyses.

4.2.2 Laser Raman Spectroscopy (LRS)

Laser Raman Spectroscopy (LRS) was done on both zircon grain assemblages to identify the mineral phases present in the core, rim, and surrounding glass. Analyses were done using a Renishaw inVia Raman System at Washington University in St Louis. The samples were mapped with a 632 nm He-Ne laser at 0.6 μm step sizes (pixel resolution) and 1 second dwell times. Phases were mapped based on the Raman peak intensity for the 357 nm and 1006 nm positions in zircon spectra, the 187 nm, 385 nm, and 478 nm positions in monoclinic ZrO$_2$ spectra (Nasdala et al., 2003), and the occurrence of a prominent peak at 262 nm for tetragonal ZrO$_2$ (Naumenko et al., 2008) (Fig. 7). The grains were also investigated for reidite, the high-pressure polymorph of ZrSiO$_4$, by searching for spectra with the characteristic Raman bands at ~840 and ~880 cm$^{-1}$ (Gucsik et al., 2004), but was not identified for the two zircon assemblages. Three ~70x70 μm regions were mapped on MZRN-1, and the entire MZRN-2 grain was mapped. Figure 8 displays color-coded distribution maps of mineral phases.

4.2.3 Hyperspectral CL

Hyperspectral Cathodoluminescence data was acquired for both zircon assemblages using a Gatan MonoCL4 Elite system attached to a FEI Quanta 200F field emission scanning electron
microscope (FE-SEM) at the Smithsonian Museum Conservation Institute, Suitland, Maryland. High sensitivity photomultiplier tube (PMT) false-color RGB composite imagery was collected with R-G-B channels corresponding to broad band filters at 600 nm (red), 400 nm (green), and 350 nm (blue). PMT imaging shows distinct zoning in the zircon core, as well as distinct zoning at the core-decomposition rim interface (e.g. blue regions) (Fig. 11). Spectroscopic data was collected using a Princeton Instrument PIXIS charged coupled device (CCD) array, with 1340 channels (unbinned) and a spectral range of the detector ~ 250 nm – 800 nm. Spectra were correlated the CSIRO luminescence spectral database (http://www.csiro.au/luminescence/). CL deconvolution was done using Chimage™ software.

4.2.4 Electron Probe Micro-Analysis (EPMA)

The composition of zircon, ZrO₂ rims, and surrounding glass were analyzed using a 5-spectrometer JEOL JXA-8200 electron microprobe equipped with 5 wavelength-dispersive spectrometers, an e2v silicon-drift energy-dispersive X-ray spectrometer, and a Gatan MonoCL cathodoluminescence detector at the Department of Earth and Planetary Sciences of Washington University in St Louis. Analyses included backscattered-electron (BSE) imaging, X-ray intensity beam-raster and stage mapping, and qualitative analysis by energy-dispersive (EDS) coupled with quantitative wavelength-dispersive (WDS) analyses to characterize the major- and minor-element mineral chemistry, and to provide context for the SIMS analyses. Wavelength scans were done to determine all analyzable elements in the zircon grains and selected major and minor elements were analyzed by wavelength-dispersive spectrometry (WDS). Element intensities mapped by EDS were: Ca, Fe, Mg, K, Si, and Ti; element intensities mapped by WDS.

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3 Hyperspectral CL data was collected by E. Vicenzi for incorporation in a subsequent publication. M. Zanetti is a contributing co-author.
were: Al, Hf, Na, Pb, Zr. WDS spot analyses made during grain traverses included Ca, Fe, P, Zr, Si, Hf, Ti, Dy, Er, Th, Y. Quantitative spot analysis of the surrounding impact melt glass was done using WDS, with 15 kV accelerating voltage and 25 nA probe current with a 40 μm beam size. A core traverse using high-probe current WDS spot analysis was done to quantify trace element concentration for comparison with CL hyperspectral data. The core traverse spots were obtained at 15 kV and 150 nA, with a 2 μm beam size.

4.2.5 Secondary Ion Mass Spectrometry (SIMS)

U-Pb spot analyses of the zircon grains were done using the Cameca IMS 1280 ion-microprobe at the Swedish Museum of Natural History in Stockholm, Sweden (NordSIM Facility). The samples were gold coated prior to U-Pb SIMS analyses and the measurement methodology closely followed the analytical protocol for zircon (Whitehouse et al., 1999, Whitehouse and Kamber, 2005). The mass-filtered $^{16}$O$_2^-$ primary ion beam was reduced through Köhler apertures of 20 and 100 μm in order to obtain a spot size of 2 and 10 μm, achieving intensities of the primary beam of about 0.2 and 2.5 nA. Oxygen flooding techniques were used to enhance the secondary Pb$^+$ ion yields (e.g., Li et al., 2009). Before each analysis the analytical area was pre-sputtered for 80 seconds, with 2.5 nA primary beam to remove gold and possible surface contamination. This procedure was followed by automatic centering of the secondary ion beam in the 4000 μm field aperture and centering of the magnetic field. The secondary ions were measured using a “peak-hopping” routine with a single low-noise electron multiplier. The mass spectrometer was operated with a mass resolution of 5394 (M/ΔM), sufficient to separate Pb peaks from molecular interferences. The obtained ages are shown in Figure 10 with 2σ errors and average ages are calculated at the 95% confidence limit.
4.2.6 EBSD

Although most of the specific electron backscatter diffraction (EBSD) results are not reported in this manuscript, analyses of the zircon grains were done by EBSD (Timms et al. 2015; in preparation)\(^4\). Thin sections containing the zircon grains were polished with a 1 μm diamond paste prior to mounting regions of interest that include the zircon grains into an epoxy disk. The sample was given a further chemical-mechanical polish prior to EBSD analysis using colloidal silica in NaOH for three hours on a Buehler Vibromet II polisher. A thin coat of carbon was applied to mitigate charging during electron microscopy, yet permit EBSD patterns to be acquired. Scanning electron microscopy (SEM) was done by N. Timms at Curtin University using a Tescan MIRA3 field emission (FE-) SEM. Atomic number contrast imaging was done using a pole piece backscatter detector, accelerating voltage of 5 kV, beam current of 22 nA at a working distance of 15 mm. Orientation mapping was done by EBSD, and was acquired using a Tescan MIRA3 FE-SEM with an Oxford Instruments AZTEC EDS/EBSD acquisition system at Curtin University. Operation conditions were optimized for EBSD and include a stage tilt of 70° around a horizontal axis and beam acceleration voltage of 20 kV (Prior et al., 1999). EBSD patterns were processed using 4x4 pixel binning and indexed using match units for zircon, cubic, orthorhombic, tetragonal and monoclinic ZrO\(_2\). However, only zircon and baddeleyite were successfully indexed. The indexing tolerance for fit of the theoretical solutions to the EBSD patterns was 1.2°, and the mean angular deviation for zircon and baddeleyite indexing solutions was 0.42° and 0.49°, respectively.

\(^4\) EBSD work was done entirely by N. Timms and colleagues at Curtin University, with M. Zanetti as contributing co-author.
4.3 Results

4.3.1 Petrography

Macroscopically and in thin-sections, the impact melt rock sample appears holohyaline and contains no liquidus crystals. The impact melt rock contains ~5% clasts that are up to a few millimeters in size. Clasts are predominantly ballen-textured silica grains (e.g., Ferriere et al., 2010). Ballen-textured silica consists of interpenetrating spheroids (similar in appearance to grape clusters) of α-quartz or α-crystobalite (Ferriere et al., 2010). Ballen-texture silica may indicate a shock pressures as high as 30 GPa (Ferriere et al., 2010) or that the grains have experienced thermal shock, whereby the grains are heated by an initial shockwave, and then rapidly quenched (Chanou et al., 2015). The rare maskelynite clasts (i.e., diaplectic feldspar) are present as transparent minerals in plain polarized light that are observed to be isotropic (fully extinct) under cross-polarization. The maskelynite grains suggest peak pressures of 20-35 GPa (e.g., William and Jeanloz, 1990; Singleton et al., 2011). All clasts show signs of thermal erosion, and have rounded edges and embayments that suggest dissolution in the impact melt.

4.3.2 Laser Raman Spectroscopy (LRS)

LRS imaging and spot analyses show that the decomposition rim is predominantly a vermicular intergrowth of baddeleyite and glass, and confirms the presence of a zircon core. Representative LRS spectra (Fig. 7) of three crystalline phases are identified in LRS mapping are color-coded and overlain on a BSE context image of MZRN-1 (Fig. 8). Within Fig. 8: blue areas are crystalline zircon, and have the spectral characteristics shown in Fig 7a; green areas are crystalline monoclinic ZrO₂ (baddeleyite) with spectra shown in Fig 7b; and red blebs indicate tetragonal ZrO₂ with spectra similar to Fig 7c. We did not observe the high pressure polymorph of zircon (reidite) among the LRS spectra from MZRN-1 and MZRN-2. The cores of both grains
are unaltered zircon. The crystalline phases in the decomposition rim are monoclinic and tetragonal polymorphs of ZrO$_2$. Within the decomposition rim, no other phases are observed, and the interstitial areas are filled with amorphous glass. No crystalline phases are observed exterior to the zircon assemblages in the impact melt glass.

4.3.3 **Cathodoluminescence**

Within the zircon core, magmatic zoning related to original crystallization is preserved and appears in photomultiplier tube RGB images as dark, low-CL intensity zones, intermediate CL-intensity pink/salmon colored zones, and white, high CL-intensity zones (Figs. 11 and 12). The baddeleyite decomposition rim is displayed in green in the false color image. A distinctive “blue” CL is observed at the interface between the decomposition rim and zircon core. The same blue CL is seen along some fracture boundaries.

4.3.4 **Electron Probe Micro-Analysis (EPMA)**

4.3.4.1 **Glass composition**

The composition of the impact melt is given in Table 1 and is an average of two time-dependent-intensity (TDI) 40 µm EPMA spot analyses done well away from the zircon grains outside of the Zr halo. Three target rock types were considered as possible components, anorthosite, mangerite, and granodiorite. Compositions of these rocks have been reported by Marion and Sylvester (2010), Marchant and Crockett (1977), and Currie (1971). We used an average composition for each of the three rocks, computed as a weighted average of the reported compositions, in a mixing model to determine proportions of the target rocks that have contributed to the impact melt (Table 1). Using a least-squares minimization technique (Korotev et al., 1995), we found that the impact melt is approximately a 50:50 mixture of the anorthosite
and mangerite compositions (Table 1), consistent with results obtained by Marion and Sylvester (2010) for crystalline impact melt at Discovery Hill.

4.3.4.2 Trace element concentrations

Selected major- and trace-element concentrations for areas of the zircon core, decomposition rim, and surrounding glass are shown in Table 2. These analyses were done to investigate elemental concentration of known and suspected CL quenchers (e.g., Fe) and activators (Ti, Th, Y, Dy, Er), and are measured in spot analyses of specific areas of interest within the photomultiplier tube RGB composite – CL image (Fig. 11). The spot analyses reported in Table 2 support the WDS spot traverse analyses shown in Fig 12. Blue CL areas have elevated concentrations of Ti, Fe, and P compared to areas of bright, dark, and salmon colored areas of CL intensity in Fig 11. Areas within the core show variation in trace element abundances that are inversely proportional to the intensity of CL (Fig. 12). Th, Y, Dy, Er were measured at or above detection limits along the core traverse and show increased concentration within areas of dark CL compared to areas of bright CL.

4.3.5 Secondary Ion Mass Spectrometry (SIMS)

4.3.5.1 Age of the grain cores

Concordia plots for MZRN-1 and MZRN-2 are shown in Fig. 10. Sample MZRN-1 was large enough to measure three 10-micron spots and six 2-micron spots. A Concordia age of 1404 ± 12 Ma (MSDW of concordance: 0.081; probability of concordance: 0.78) was measured for the 10 micron spot analyses on MZRN-1. Including all spot analyses on sample MZRN-1, the Concordia age is 1403 ± 10 Ma (MSDW of concordance: 0.85; probability of concordance: 0.36). Error bars are 2σ with decay-constant errors included. The smaller size of sample MZRN-
only permitted five 2-micron spot analyses. A Concordia age of 1392 ± 17 Ma (MSDW of concordance: 5.6; probability of concordance: 0.018) was measured for MZRN-2. Error bars are 2σ with decay-constant errors included. Despite using ultra high resolution and precision instrumentation, an age for the baddeleyite rim could not be determined. The small surface area of the largest ZrO₂ region (<5 microns) and the proximity to the glass did not allow us to focus the beam to obtain an accurate measurement.

4.4 Discussion

The following sections discuss the results of the multiple analytical techniques in order to bring observations and measurements into context with the geologic history of the zircon grains. We first discuss the provenance and petrogenesis of the zircon grains in terms of original crystallization ages and textures, and in terms of subsequent decomposition as a result of heating induced by impact melt entrainment. We then discuss the provenance and rheology of the impact melt glass in which the samples occur. The general implications of our observations, in terms of decomposed zircon grains found in tektites and lunar impact melt samples, is also discussed.

4.4.1 Petrogenesis of Decomposed Zircons

Although both zircon grains are fractured, there are no diagnostic shock metamorphic features in the core of the grains, such as planar micro-features (glassy lamellae, micro-twins, or high-pressure polymorphism). The two zircon grains MZRN-1 and MZRN-2 show evidence of short-lived presence in the melt before the melt quenched. The 40 µm and 20 µm thick decomposition rims around the grains indicate incomplete dissociation of the grains within the area of the decomposition rim during heating. Temperature conditions above 1700 °C are commonly inferred for impact melts from the existence of the vermicular decomposition texture
(Butterman and Foster, 1967; El Goresy, 1968; Wittmann et al., 2006). On the basis of tetragonal ZrO$_2$ present in the rim of MZRN-1, as seen in the LRS data (Fig. 8), we can infer that the temperature of the melt was at least 1687°C (Kaiser et al., 2008). Because the SiO$_2$ released during the decomposition reaction was immediately mixed with infiltrating impact melt, a more accurate estimate of the melt temperature using the ZrO$_2$ – SiO$_2$ phase diagram is not appropriate. Evaporation experiments inducing the decomposition of zircon in a vacuum indicate that zircon starts to break down at ~1500 °C in a vacuum (Chapman and Roddick, 1994). Kinetically, this reaction depends on the temperature, compositional, and structural conditions of the zircon crystals. From their experiments under vacuum conditions, Chapman and Roddick (1994) suggested that the rate of decomposition can proceed at ~0.46 µm/min at 1670°C, meaning the ~40 µm-thick MZRN-1 decomposition rim would develop in ~90 min, and the ~20 µm-thick MZRN-2 decomposition rim in ~55 min. However, in order to produce a 10-µm-thick decomposition rim in a high-purity zircon, Chapman and Roddick (1994) reported ~30 min at 1600 °C under vacuum conditions. According to these authors, this reaction is six times faster at 1700 °C and would therefore require ~5 min to form the ~10-µm-thick decomposition rim of the zircon grain in their experiment. The zircon grains in the Mistastin impact melt glass are not high-purity, and the reaction did not take place under vacuum conditions, however the results of Chapman and Roddick (1994) show that the reaction at the temperatures the MZRN grains experienced took place within perhaps a few hours.

Samples MZRN-1 and MZRN-2 have similar ages (1403 ± 10 and 1392 ± 17 Ma; respectively), the same, within uncertainty. However, both MZRN zircon ages are less than those of zircons in the target rock. U-Pb dating of zircons in the major target rocks from outside the impact area were measured by LA-ICPMS by Marion et al. (2010) who reported ages of 1451 ±
12 Ma (Mangerite), 1440 ± 13 Ma (Granodiorite Gneiss), 1438 ± 9 Ma (Anorthosite), and 1429 ±
10 Ma (Granodiorite). Three impact melt locations also contained zircons that were dated by
Marion et al. (2010) at 1435 ± 13 Ma (CM023, Cote Creek), 1431 ± 21 (CM005, Steep Creek),
and 1413 ± 13 Ma (CM065; South Ridge). Marchand and Crocket (1977) measured the age of
the mangerite at 1409 ± 57 Ma, and 1318 ± 17 Ma for the granodiorite by Rb/Sr dating. Ages of
the MZRN samples fall below the error bars of all major target rocks, but are closest in age to the
granodiorite. Minor resetting of grains owing to Pb-loss through thermal diffusion may have
occurred. Although the ages of the two MZRN samples are close in absolute age, it is possible
that they came from different target rocks, as the Pb-loss of the individual grains may have been
different. The possibility of minor resetting prevents us for unambiguously determining the
source rock of the zircon grains on the basis of zircon ages, but the melt mixing composition
~50-50% anorthosite and mangerite would suggest that either of these target rocks contributed
the zircons to the melt. The Discovery Hill outcrop, where the sample was obtained, contains
enormous meter to decameter-sized mangerite boulders, and the bulk impact melt rock
compositions reflect a greater mangerite component (Marion et al., 2010). Our melt mixing
calculations (Section 3.4.1) suggest that the impact glass sample represents a ~50:50 mixture of
anorthosite and mangerite. The mangerite target rocks contain a much higher concentration of
zircon grains compared to the other target rocks, and the anorthosite is poor in zircons (Marion et
al., 2010). On these grounds, we infer that the MZRN samples may have been sourced from the
mangerite target rocks.

With respect to dating the time of the impact, an absolute age could not be determined
from the baddeleyite decomposition rim. The measurement is complicated by a number of
factors, including crystal orientation effects and generally low U contents in the baddeleyite
grains (e.g., Wingate and Compston, 2000). Regions of the baddeleyite decomposition rim are rarely more than 2 microns across, and only one region on MZRN-1 was potentially large enough (~5 µm) to attempt a measurement with the 2 µm ion beam. In order to locate this area the beam was rastered over the area, but no strong signal was observed. Another factor contributing to the difficulty of the measurement were low count values for Pb owing to the young 36 ± 4 Ma age of the impact. However, although we were not able to measure a reliable age, we can infer that the U concentration in the decomposition rim baddeleyite is very low. During the phase conversion from ZrSiO$_4$ to ZrO$_2$, incompatible elements were purged from zircon and released into the melt.

4.4.2 Provenance of the Impact Melt Glass

Although the Discovery Hill outcrop is largely crystalline impact melt, the impact melt glass sample studied here is chemically indistinguishable in composition from the crystalline melt from this general location (see values in Marion and Sylvester, 2010). If the Discovery Hill outcrop is a terrace melt-pond of the Mistastin Lake impact structure, as suggested by Tornabene et al. (2013), it is possible that the specimen was formed as a lava fragment or large droplet of melt from a cascading flow on the wall of the crater. Alternatively, the impact melt glass sample may have quenched when it came into contact with one of the large (>5 m) mangerite blocks within the Discovery Hill melt. Thin sections of samples taken radially away from these large boulders show increased crystal content and longer duration crystal growth with increasing distance from the block, from wispy, feathery shapes of plagioclase phenocrysts with a glassy matrix, in areas close to the boulders, to coarser grained (mm-sized) lath-shaped plagioclase phenocrysts in the massive, columnar jointed melt.
The morphology, texture, and varying levels of shock seen in different clasts within the impact glass sample suggest that clasts were not entrained for a long period of time before the melt was quenched. The clasts within the impact melt, including the zircon grains, were likely introduced to the melt as clasts. Our melt-mixing calculations suggest that the melt glass is derived from both anorthosite and mangerite, indicating that two liquids were formed and well mixed. Although clasts can survive this complete melting, we infer that the zircon grains were introduced into the melt, possibly as part of an air-fall deposit. The various levels of shock and grain textures (seen in incomplete conversion to maskelynite in rare plagioclase grains, and in ballen texture quartz), suggest grains may have come from different areas of the crater. Although speculative, the flow textures in the melt and the observation of zircon movement (Figs. 14, 15) could also imply settling after air-fall deposition. The difference in thickness of the decomposition rims suggest that the two zircon grains may have spent different lengths of time in the melt, or that the decomposition preceded faster in MZRN-1.

4.4.3 Rheology of Impact Melt Glass

The LRS observation of small blebs of remnant tetragonal ZrO$_2$ is useful as a temperature constraint for the decomposition of the zircon grains. Despite the lack of agreement with EBSD analyses (shown in Fig 8), which may not have indexed the tetrag-ZrO$_2$ owing to poor crystallinity or through removal of the blebs during re-polishing, we can infer that the temperature of the impact melt was $>1687^\circ$C (Kaiser et al., 2008). Further evidence for this temperature can be inferred by the lack of crystalline SiO$_2$ phases, such as cristobalite, the high-temp SiO$_2$ polymorph that should be present if, that would be expected if the grains experienced lower temperatures. We also observe that the zircon grains were mobile in the melt prior to the quenching of the impact melt glass (Fig. 15), as the impact melt flowed around the grains. We
use the model of Girodano et al. (2008), which predicts the non-Arrhenian Newtonain viscosity of silicate melts as a function of T and melt composition, to approximate the viscosity of the impact melt for a range of plausible melt temperatures (>1600°C; 1000°C and 1200°C), and the glass composition measured with EPMA (Fig. 17). The model results suggest that the impact melt was very low viscosity (<1 Pa·s), approximately the viscosity of SAE 40 motor oil, and was likely highly mobile prior to quenching.

4.4.4 Material exchange within Decomposition Rim

EPMA x-ray maps (Fig. 9) show the concentration of elements within the grains and surrounding glass. Three selected element maps show the relative abundances of the major elements in the impact melt glass surrounding assemblages MZRN-1 and MZRN-2 (Al, Na in Fig. 9) and from the relative abundance of Zr in the vicinity of assemblage MZRN-2 (Zr in Fig. 15b). These maps also show that there was an exchange of material both into and out of the decomposition rim. Impact melt infiltrated interstitial spaces between ZrO\textsubscript{2} dendrites (Figs. 9 A and B), while ZrO\textsubscript{2} diffused into the impact melt (Figs. 15 A and B). The impact melt surrounding MZRN-2 exhibits a ~10 µm-wide Zr-halo around the zircon assemblage that is drawn out into a ~70 µm-wide halo of Zr-enrichment that can be traced for several mm away from MZRN-2 and shows a wavy, ribbon-like shape where the melt rock exhibits a flow texture (Fig. 15). Interestingly, the Zr-enriched halo in the impact melt surrounding MZRN-2 also extends ca. 0.3 mm "ahead" of the long ribbon trace that presumably marks a relative movement of the zircon assemblage in the melt, or a differential movement of the melt relative to MZRN-2 in a turbulent flow pattern. A similar, but narrower halo of enhanced Zr content is observed around MZRN-1 (not pictured).
**4.4.4.1 Implications for Zircon Decomposition Studies**

U-Pb ages of granular-textured zircon grain separates from impactites have been used to approximate impact ages for several terrestrial impact structures (e.g. Sudbury: Krogh et al., 1996; Chicxulub: Krogh et al., 1993; Vredefort: Kamo et al., 1996; Gardnos: Kalleson et al., 2009; Acraman: Schmieder et al., 2013). These data were mostly obtained by whole-grain thermal ionization mass spectrometry of statistically significant subsets of samples, and assumed close to complete resetting of the U-Pb radioisotopes during impact. This assumption that may not be justified in most cases, as indicated by the frequent occurrence of relict zircon cores in impact-decomposed ZrO$_2$-ZrSiO$_4$ crystals (e.g., Wittmann et al., 2006). Nonetheless, a lack of petrographic context for these granular zircon grains raised concerns about the veracity of these impact ages (cf., Krogh et al. 1993; Kalleson et al. 2009). Recently, decomposition-textured zircon grains were studied in-situ with the sensitive high resolution ion microprobe (SHRIMP) in lunar meteorite Dhofar 458 (Zhang et al., 2011) and in Apollo sample 15405 (Pidgeon et al., 2013) to determine impact ages. However, these rare lunar zircon grains tend to be very small, and the parent lunar craters for these impact melts and the precisions of these analyses remain poorly constrained. Our investigation of the Lake Mistastin sample, where the age is well known, and the zircon history is not complicated by multiple events can provide a baseline to help to better understand the effects of impact-related thermal events on zircon geochronology.

**4.4.5 Cathodoluminescence Dependence**

As with the LRS data, we do not observe CL features in the zircon cores that indicate structural lattice changes due to shock processes, suggesting that the margins of the grains were only affected by the high heat of impact melt immersion, and not shock effects. The lack of
major modification is also evident in the preserved magmatic zoning in the cores observed by CL.

The hyperspectral data set collected for both zircon assemblages allows for a precise examination of areas of CL features, and the spectrum can be related to specific elements involved in activation of CL. The crystallization zoning patterns in CL display broad spectral features, in addition to high-frequency sharp peaks, interpreted as resulting from HREEs based on correlation with the CSIRO luminescence spectral database (http://www.csiro.au/luminescence/). Figure 14 shows the level of detail acquired by hyperspectral - CL analyses of MZRN-2. The left column show the CL for a given wavelength of spectra (325 nm, 400 nm, 575 nm). CL spectra in energy space that were extracted from 3 spots of the zircon grain; the blue CL rim, the bright CL at the center of the grain, and the adjacent salmon CL (Fig. 14 center column). The same spectra are also shown as a function of wavelength in the right-hand column in Fig 14. The overall broad shape of the spectrum appears to be intrinsic to the region of interest. The sharp peaks can be assigned to specific rare earth elements (REEs) (Tm$^{3+}$, Er$^{3+}$, Dy$^{3+}$). Our EPMA spot traverses across the zircon assemblage MZR-2 (Figs. 16) were done in part to quantify the concentration of REEs that could be activators of the CL spectra.

It is known that REEs (including Sc and Y) are important activator elements, and that Dy$^{3+}$ is the main CL activator in zircon by virtue of more efficient excitation in the zircon lattice (Kempe, 2000). Our high resolution WDS trace-element analyses (traverse seen in Fig. 12a; 2 µm beam, 15 kV, 150 nA) indicate an inverse correlation between Y concentrations and CL intensity (Fig 12b). Dark areas in CL within the core contain low concentrations (measureable, but at detection limits) of Er (20-30 ppm) and Dy (~10 ppm), and relatively high concentrations
of Y (~1500 ppm). Measurable concentrations of Th\(^{4+}\) and Er\(^{3+}\) are also observed in areas of high CL intensity. It may be that Y\(^{3+}\) quenching dominates the CL signal in dark areas, and activators (Dy\(^{3+}\), Er\(^{3+}\), and Th\(^{4+}\)) are sufficient to cause the bright CL, even in very low concentrations (< 100 ppm). An alternate possibility may be that elemental ratios (e.g., Y/Dy or Y/Er), rather than absolute concentrations are an important factor in determining the CL intensity.

Blue CL observed in zircon at the dissociation boundary is probably related to the dissociation of zircon to ZrO\(_2\) + SiO\(_2\), and possibly related to lattice structure changes, rather than the result of elemental activation. Although not well understood, blue CL in zircon has been attributed to a delocalized electron on the [SiO\(_4\)] group (Kempe, 2000), suggesting the interface records the initial stage of breakdown. Blue CL along fractures in the grain may be related to the infiltration of melt, which imparted heat energy into the lattice, delocalizing electrons in the zircon. If left undisturbed it is likely that zircon along these fractures would have begun decomposing, but the process was halted by the quenching of the impact melt.

4.5 Conclusions

The decomposition of the zircon grains was induced entirely as a result of their entrainment in very high temperature, low-viscosity impact melt. No evidence of shock features are seen in the zircon cores, although the fracturing of the grains is likely to have occurred as a result of the impact. The relict cores of the grains are ZrSiO\(_4\), and magmatic zoning related to initial crystallization ~1430 ± 50 Ma is preserved. The provenance of the zircons is uncertain, as the high temperatures of the melt may have allowed some resetting of the core ages by Pb loss; however, we postulate that they originated in the mangerite target rocks. Within the decomposition rim, both monoclinic ZrO\(_2\) and tetragonal ZrO\(_2\) phases were identified by LRS.
EBSD did not discern tetragonal ZrO$_2$, possibly owing to poor crystallinity or removal during re-polishing. No other mineral phases, other than glass, were identified in the interstitial areas of the decomposition zone. An independent age of the Mistastin Lake impact structure through dating of the decomposition rim was not able to be determined, due in part to a young impact age (36 ± 4 Ma), small sampling area, and low counting statistics of U and Pb.

No crystalline phases were observed in the impact melt glass, and its composition reflects a roughly 50:50 mixture of anorthosite and mangerite target rocks. The compositions of the interstitial rim material and surrounding glass indicate there was an exchange of elements both into and out of the decomposition rim, with zircon elements (e.g., Zr and Pb) expelled into the surrounding melt, and impact melt infiltrating the decomposition rim. Composition of the interstitial material is similar to the surrounding impact melt glass, albeit with a higher Zr concentration. No areas of pure SiO$_2$ were identified, likely due to rapidly infiltrating impact melt, however a slight increase in SiO$_2$ can be inferred in areas closest to the core-decomposition rim interface. Based on occurrence of preserved, metastable tetragonal ZrO$_2$, we infer that the temperature of the impact melt was in excess of 1687°C, with an estimated viscosity of ~0.5 Pa*s (e.g. motor oil). The low viscosity of the impact melt likely facilitated the infiltration of melt within the decomposition rim.

The core-decomposition rim interface displays a characteristic blue-colored zoning, not previously reported in similar occurrences, which is also observed along melt filled fractures within the core. The blue CL is possibly related to emission from electron defects localized at SiO$_4$ groups (e.g. Kempe, 2000) as silicon ions diffused from the zircon lattice. Blue CL areas within fractures may indicate areas that were about to breakdown prior to quenching. Trace-
element compositions determined along a traverse of the magmatic zoned cores show an inverse relationship of REEs with CL intensity, and suggest that elemental ratios (e.g. Y/Dy and Y/Er) rather than absolute concentrations of REEs could be an important factor in interpreting the CL data.

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Figure 4-1: Mistassin Geologic Map: A) Simplified geologic map of the Mistassin Lake Impact Structure, located in northern Labrador, Canada (55º53’N; 63º18’W; modified after Marion and Sylvester, 2010). The target rocks consist of anorthosite, mangerite, and granodiorite. Horseshoe Island is a remnant central peak structure. B) Discovery Hill (black arrow in A), is an ~80 m thick, perched crystalline impact melt pond. Red star marks area where the hand sample in Fig. 2 was found. Listric faults (red lines) related to terrace development separate a mangerite unit (eastern most block) from a granodiorite unit (middle block).
Figure 4-2: Mistastin Impact Melt Glass: A) The hand-sample collected from the top of Discovery Hill. Note the conchoidal fracture and resemblance to a clast-laden obsidian glass. B) The impact melt glass as seen in thin-section. The sample has <5% clasts (predominantly ballen-texture silica), and very few vesicles.
Figure 4-3: X-Polarized Photomicrographs: Cross polarized light photomicrographs of Discovery Hill impact melt glass zircons MZRN-1 (A) and MZRN-2 (B). Brown halos are zones of decomposition. The surrounding impact melt is isotropic quenched impact melt glass.
Figure 4-4: Backscattered electron images (BSE) of decomposed zircon grains found within Discovery Hill impact melt glass. A) MZRN-1; Grain is ~400 µm long, ~175 µm wide, and is exposed parallel to the C-axis. Decomposition rim is ~40 µm thick. B) MZRN-2; Grain is ~100 µm in diameter, and is exposed parallel to the A-axis. The decomposition rim is ~20 µm thick.
Figure 4-5: A) inset from Fig. 4 showing detail of the decomposition rim and core-rim interface. Black areas are amorphous impact melt glass and interstitial glass (nearly compositionally identical, except for Zr content), white areas are monoclinic ZrO$_2$ (Baddeleyite), and grey areas are ZrSiO$_4$ (Zircon). B) Close-up image of the decomposition interface. Zircon “fingers” cored by baddeleyite are surrounded by amorphous silica-rich regions.
Small blebs of Tetragonal ZrO$_2$ and amorphous SiO$_2$ phases identified by LRS in MZRN-1 suggest that the grains experienced temperatures in excess of 1687$^\circ$C.
Figure 4-7: Example LRS Raman spectra for mineral phases identified in Fig. 8. A) Typical zircon spectrum B) Typical monoclinic ZrO$_2$ (baddeleyite) spectrum. C) Example tetragonal ZrO$_2$ spectrum identified in regions of red blebs in Fig. 7. Note low counts, but bands are above noisy background, indicating poor crystallinity.
Figure 4-8: A) Three LRS maps over a BSE image. Mineral phases are color coded based on LRS indexing to specific peaks or spectra, with ZrSiO$_4$ colored blue, monoclinic ZrO$_2$ colored green, and blebs of tetragonal ZrO$_2$ in red. B) Phase map based on EBSD mapping (modified from Timms et al., 2015, in preparation). Zircon is coded blue, baddeleyite is coded red. No tetragonal ZrO$_2$ was identified by EBSD analysis. 3 large spots within core are from 10 μm SIMS analyses.
Figure 4-9: X-Ray intensity maps. A) Aluminum content. B) Sodium. Note the areas of high aluminum and sodium concentration that have infiltrated the interstitial areas of the decomposition rim.
Figure 4-10: Concordia plots of SIMS spot analyses of the MZRN zircon cores: A) MZRN-1 age of 1403 ±10 Ma (thick red oval) based on three – 10 µm spot (black ovals in center) and six – 2 µm spot (large red ovals) analyses. B) MZRN-2 age of 1392 ± 17 Ma (thick red oval) based on five – 2 µm spot analyses. Note low probability of concordance, likely due to a small amount of resetting.
Figure 4-11: False color, photo-multiplier tube CL image over BSE image. Green CL indicates baddeleyite phases, white/salmon-pink/dark regions of the zircon core are remnant magmatic zoning. Blue CL around the core-rim interface and along fractures is a distinctive CL color indicating the position of the decomposition front. Blue CL along fractures likely indicates the onset of Si dissociation.
Figure 4-12: A) CL – PMT image from Fig. 11. Yellow dashed line (A-A’) marks the location of the high-probe current WDS trace element spot analyses. B) Wt % element vs position along cross-section (A-A’) for elements Th, Y, Dy, Er. Note the inverse correlation of CL intensity with Y and other trace element concentrations.
Figure 4-13: Hyperspectral CL data collected from various spot analyses. (Left column) BSE images with spot analyses (yellow dots). Top row: decomposition interface (blue rim in Fig 10); middle row: bright CL (white in Fig 10); bottom row: intermediate CL (salmon color in Fig 10). (Center column) de-convolved spectra for spot analyses as a function of energy (eV). Broad humps may represent “intrinsic” CL, and sharp peaks are related to specific REE activation. (Right column): same spot spectra as a function of wavelength.
Figure 4-14: A) flow banding in the impact melt glass as seen in high contrast BSE image of MZRN-2. A bright trail of high atomic number elements extends more than 2 mm from the zircon grain. B) Elemental x-ray map of MZRN showing zirconium concentration (red = high, blue = low). Note a distinct zirconium halo around the grain (white dashed lines mark the margins), indicating Zr diffused into the surrounding melt during decomposition and transport in the melt.
Figure 4-15: A) Wt % Zr as a function of distance from MZRN-2 extending from the grain into the surrounding melt glass. Red square represents a measurement made within the interstitial glass in the decomposition rim. Concentration remains steady through the Zr-halo (Fig 12), and tapers off to very low concentrations in the impact melt glass. B) Location of south-to-north traverse.
Figure 4-16: Graph of calculated viscosity versus temperature using the composition of the impact melt glass. Based on the occurrence of a tetragonal-ZrO$_2$ phase and no crystalline SiO$_2$ phases preserved in the decomposition rim, temperature of the melt was is excess of 1687$^\circ$C. This implies a viscosity of $\sim$0.6 Pa*s, similar to SAE 40 motor oil.
### Table 4.1: Impact Melt Glass Composition

<table>
<thead>
<tr>
<th>Mixture Model</th>
<th>Anorthosite avg wt’%</th>
<th>Mangerite avg wt’%</th>
<th>Granodiorite avg wt’%</th>
<th>Glass away fr Zirc TDI**</th>
<th>Glass Avg</th>
<th>Deviation from Model Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>n=2</td>
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<tr>
<td>SiO2</td>
<td>54.45857</td>
<td>61.01977</td>
<td>67.27333</td>
<td>57.60795</td>
<td>54.64425</td>
<td>2.963696</td>
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<tr>
<td>TiO2</td>
<td>0.365714</td>
<td>1.251628</td>
<td>0.51</td>
<td>0.790953</td>
<td>0.977463</td>
<td>-0.18651</td>
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<tr>
<td>FeO</td>
<td>3.130714</td>
<td>8.803256</td>
<td>3.873333</td>
<td>5.853534</td>
<td>6.172415</td>
<td>-0.31888</td>
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<tr>
<td>MnO</td>
<td>0.067143</td>
<td>0.115814</td>
<td>0.066667</td>
<td>0.090505</td>
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<tr>
<td>MgO</td>
<td>1.890714</td>
<td>0.944186</td>
<td>0.49</td>
<td>1.436381</td>
<td>1.28689</td>
<td>0.149491</td>
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<td>CaO</td>
<td>9.576429</td>
<td>4.546744</td>
<td>1.933333</td>
<td>7.16218</td>
<td>7.45251</td>
<td>-0.29033</td>
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<tr>
<td>Na2O</td>
<td>4.443571</td>
<td>3.354419</td>
<td>3.703333</td>
<td>3.920778</td>
<td>4.193605</td>
<td>-0.27283</td>
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<tr>
<td>K2O</td>
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<td>3.893953</td>
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<tr>
<td>P2O5</td>
<td>0.146429</td>
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<td>-0.09553</td>
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<tr>
<td>LOI</td>
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<td>0.883721</td>
<td>0.46</td>
<td>0.844272</td>
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<tr>
<td>Total</td>
<td>99.94</td>
<td>99.53791</td>
<td>98.79333</td>
<td>99.71542</td>
<td>97.38025</td>
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</tr>
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</table>

* Average weighted values taken from tables in Marion and Sylvester (2010); Marchand and Crocket (1977); and Currie (1971)

** Values from our study using time dependent intensity (TDI) EPMA measurements (15 kV; 25 nA; 40 μm spot)
### Selected Major and Trace Element Concentrations: Areas of Zircon Core, Rim, and Impact Melt Glass

All values in ppm, otherwise in wt% as indicated.

<table>
<thead>
<tr>
<th>Phase</th>
<th>CL*</th>
<th>Ti</th>
<th>Fe</th>
<th>Th</th>
<th>Zr</th>
<th>Y</th>
<th>Hf</th>
<th>Dy</th>
<th>Er</th>
<th>P</th>
<th>Hf/Zr</th>
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</thead>
<tbody>
<tr>
<td>Zircon - Blue CL rim (50nA)</td>
<td>Blue</td>
<td>200-400</td>
<td>850-1000</td>
<td>207</td>
<td>-</td>
<td>670</td>
<td>-</td>
<td>10-220</td>
<td>≤ 150</td>
<td>-</td>
<td>-</td>
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<tr>
<td>Zircon - Blue CL rim (150nA)</td>
<td>Blue</td>
<td>830</td>
<td>2970</td>
<td>407</td>
<td>-</td>
<td>0</td>
<td>-</td>
<td>0</td>
<td>130</td>
<td>500</td>
<td>0.024</td>
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<tr>
<td>Zircon - Blue CL rim, first point of traverse (150nA)</td>
<td>Blue</td>
<td>900</td>
<td>4300</td>
<td>407</td>
<td>-</td>
<td>457</td>
<td>-</td>
<td>0</td>
<td>0</td>
<td>500</td>
<td>0.022</td>
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<tr>
<td>Zircon - discrete points (150nA)</td>
<td>Bright</td>
<td>150-250</td>
<td>250-475</td>
<td>~35</td>
<td>-</td>
<td>0</td>
<td>-</td>
<td>0</td>
<td>0</td>
<td>280</td>
<td>-</td>
</tr>
<tr>
<td>Zircon - discrete points (100nA)</td>
<td>Dark</td>
<td>170-330</td>
<td>360-800</td>
<td>200-340</td>
<td>-</td>
<td>1000-1780</td>
<td>-</td>
<td>60?</td>
<td>0</td>
<td>300-400</td>
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<td>180-260</td>
<td>300-600</td>
<td>0-200</td>
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<td>600</td>
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<td>0</td>
<td>300-400</td>
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<tr>
<td>Zircon - core traverse (150 nA)</td>
<td>Bright</td>
<td>100-150</td>
<td>350-400</td>
<td>&lt;100</td>
<td>-</td>
<td>0</td>
<td>-</td>
<td>0</td>
<td>~135?</td>
<td>240-340</td>
<td>-</td>
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<tr>
<td>Zircon - core traverse (150 nA)</td>
<td>Dark</td>
<td>100-300</td>
<td>300-590</td>
<td>150-300</td>
<td>-</td>
<td>1200-1700</td>
<td>60-150</td>
<td>70-325</td>
<td>350-430</td>
<td>-</td>
<td></td>
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<tr>
<td>Zircon - core traverse (150 nA)</td>
<td></td>
<td>200-300</td>
<td>380-945</td>
<td>60-120</td>
<td>-</td>
<td>280-630</td>
<td>&lt;50</td>
<td>150-300</td>
<td>270-420</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Baddelyite (50nA)</td>
<td></td>
<td>-</td>
<td>6900</td>
<td>4.57%</td>
<td>420</td>
<td>6.33%</td>
<td>120</td>
<td>910</td>
<td>200</td>
<td>0</td>
<td>0.022</td>
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<tr>
<td>Interstitial Rim Glass (adjacent to Baddelyite and Zircon)</td>
<td></td>
<td>-</td>
<td>6400</td>
<td>4.70%</td>
<td>0</td>
<td>4.90%</td>
<td>0</td>
<td>500</td>
<td>260</td>
<td>0</td>
<td>1400</td>
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<tr>
<td>Interstitial Rim Glass (average)</td>
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<td>-</td>
<td>5900</td>
<td>4.80%</td>
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<td>190</td>
<td>0</td>
<td>0</td>
<td>70</td>
<td>0</td>
<td>1800</td>
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*Color in CL – PMT image (Figure 11)