The History of Water at Lyot Crater, Mars: Possible Surface Manifestations of Ancient Groundwater and/or Recent Climate Change

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The History of Water at Lyot Crater, Mars: Possible Surface Manifestations of Ancient Groundwater and/or Recent Climate Change

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MGeol

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Abstract

This thesis explores the history of water in and around Lyot crater (50°N, 30°E), a ~215 km impact crater on Mars. The primary research objective is to understand whether the landscape records the actions of both ancient subsurface water and recent atmospherically-derived water, to help determine whether the study area contains material of astrobiological interest.

I first mapped the distribution of landforms and surface types potentially indicative of water and/or ice across the interior and ejecta blanket of Lyot crater. I identified landforms of particular interest to this study – fluvial channels, fan deposits and clastic polygonal networks – and recognised that the majority of fluvial features are within the crater interior. Using geomorphological mapping, I constructed a stratigraphic history of the units and landforms in the crater interior, providing context for the fluvial activity.

Morphological study of the small channels and fan deposits suggests they have a subaerial origin, with water sourced from the melting of atmospherically-deposited ice-rich material, and reveals the presence of potential standing bodies of water within closed basins during the mid- to late-Amazonian.

Clastic polygonal networks in the outer ejecta blanket were analysed: their spatial distribution indicates a genetic link to the impact event. A mechanism is proposed whereby thermal contraction cracks in ice-rich ground are modified to form angular clastic margins.

Impact models, using iSALE hydrocode, show a maximum exhumation depth of 10-12 km for the Lyot-forming event, indicating that the cryosphere was penetrated and that outflow channels to the north could be groundwater related. Furthermore, pore ice/water in the inner ejecta might have been stable as liquid water, and later flowed into the region where the polygons formed. These results indicate a history of both ancient subsurface water and recent atmospheric water, and that Lyot Crater is a site of significant astrobiological interest.
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Chapter 1: Introduction

1.1 Introduction

Lyot crater (50°N, 30°E) is a ~215 km impact crater located in the northern latitudes of Mars, immediately north of the dichotomy boundary and Deuteronilus Mensae (Russell and Head, 2002; Dickson et al., 2009; Harrison et al., 2010; Hobley et al., 2014; Weiss et al., 2017; Pan et al., 2018). It is the most recent large impact crater on the surface of Mars, being dated to the late-Hesperian/early-Amazonian, and the deepest point in the northern hemisphere with an elevation of ~7000 m (Figure 1.1; Russell and Head, 2002; Dickson et al., 2009; Harrison et al., 2010; Weiss et al., 2017). The crater has a central peak, within an inner peak-ring, and a double layer ejecta blanket extending ~2.5 crater radii, with a distinct inner ejecta scarp at ~1 crater radii. The ejecta blanket is not well preserved in the south/southwest, either due to superposition of the later deposits of Deuteronilus Mensae (Russell and Head, 2002; Harrison et al., 2010; Robbins and Hynek, 2011), or a lack of ejecta deposition in this region as a result of an oblique impact angle (Russell and Head, 2002; Weiss et al., 2017; Pan et al., 2018).

Lyot crater is a fascinating location that contains evidence of some of the most recent fluvial channels on Mars (Dickson et al., 2009, Fassett et al., 2010; Weiss et al., 2017), alongside geomorphic features, such as viscous flow features and potential sorted stone circles, indicative of periglacial and/or glacial activity (Dickson et al., 2009; Hobley et al., 2014). Further to this, Russell and Head (2002) argued via cratering mechanics that the impact event should have penetrated the cryosphere (a region of crust theorised to remain below the freezing point of water), and thereby released ancient groundwater onto the martian surface. Thus, Lyot crater is a site where both ancient water sourced from the subsurface, and more recent water sourced from the atmosphere could be present. It might also have the youngest sample of cryospheric material present at the martian surface.
Figure 1.1: A) MOLA topographic map displaying the location of Lyot crater on a global map of Mars. B) MOLA topographic map showing the Lyot crater study area with the approximate locations of the outer ejecta extent, inner ejecta extent, crater rim, inner peak-ring and central peak marked by black lines.

The extensive theorised subsurface ice on Mars could contain material that dates back to the end of the Noachian (Smith and McKay, 2005). Any life that might have originated during this period could theoretically have been preserved in such ice for substantial
periods of time (Bada and McDonald, 1995; Smith and McKay, 2005; McKay, 2010). Furthermore, the subsurface environment on Mars is far more hospitable to life than the surface – in the subsurface life would be protected from harmful surface radiation and large sterilising impact events (Westall et al., 2000; Dartnell et al., 2007). Below the cryosphere, temperatures could be above the freezing point of water due to radiogenic heating, providing a potential habitable environment (Clifford, 1993; Clifford et al., 2010). Therefore, the excavation of such subsurface material and its deposition in a location that is easier to study from orbit, or reach via the use of rovers, gives an opportunity to study material of potential astrobiological importance, a major focus of planetary exploration programs worldwide (e.g. Bada et al., 2005, Vago et al., 2006, Grotzinger et al., 2012; Mustard et al., 2013).

The presence of landforms potentially indicative of melt in and around Lyot crater also has significance when assessing the astrobiological potential of Lyot crater. The melting of ice-rich material during a time period that is largely inferred to be hyperarid reveals information about the martian climate during the Amazonian, and has important implications for the habitability of Mars during this time period.

Thus, Lyot crater has a long history related to water and ice, as well as potentially providing a location where subsurface ancient ice and/or groundwater can be studied. This means that Lyot crater could preserve material of astrobiological importance, and so be a potential high priority future landing site. It is for these reasons that Lyot crater is chosen as the focus of this thesis.

1.2 The Main Research Questions

To study the history of water within and around Lyot crater, and reveal whether it could contain material of astrobiological importance, a number of research questions were posed. These questions guided my research and enabled me to uncover new information about the Lyot Crater region, including the environmental conditions and processes this location might have experienced. The main research questions and objectives linked to these questions are presented below.

1.2.1 Where are the key landforms and landscapes located?

Previous work on Lyot crater has revealed that there is evidence for landforms and landscapes that appear to have been formed from fluvial (Dickson et al., 2009; Harrison et al., 2010; Hobley et al., 2014; Weiss et al., 2017), and possibly glacial and periglacial
processes (Dickson et al., 2009; Hobley et al., 2014). However, it is uncertain as to why and how these landforms formed in this area. Thus, this project aimed first to identify the different geomorphic landforms and particular landscapes associated with water and/or ice, and second to map where these landforms occur, to see if there is a link with particular surface materials. The approach was to survey the landscape around Lyot crater to document the features of interest, before producing a more formal map of their occurrence across the area. The mapping served to identify patterns in the spatial distribution of the landforms so that, following this analysis, key landforms were identified as subjects for in-depth study.

1.2.2 What is the stratigraphic history of Lyot crater?

Once the spatial distribution of the landforms was mapped out, a more in-depth map of a specific region of importance was created. Units were described from observations and mapped throughout the crater interior. The stratigraphic relations of these units were used to identify relative timings, which are checked against absolute ages taken from previous studies of Lyot crater. This was used to identify the stratigraphic history of Lyot crater. This history was necessary to study the timing of the water-related activity in and around Lyot crater, as well as to provide context for in-depth study of landforms of importance in the mapped region.

1.2.3 Are the actions of both atmospheric and ancient water/ice recorded at Lyot crater?

One of the key questions in this thesis was whether the Lyot study area contains evidence of both ancient subsurface and more recent atmospherically-derived water/ice. Key landforms were chosen based upon mapping and observations of the landscapes in and around Lyot crater, selecting those that are most likely to have formed via recent atmospheric water, and those that are most likely to have been formed by more ancient subsurface water. Once these landforms were selected, they were studied in more depth using morphological analysis and comparison to both terrestrial and martian analogues. These studies were then used to propose how these landforms might have formed and discuss what this means for the history of water in and around Lyot crater.

1.2.4 Was water and/or water ice released onto the surface as a result of the impact event?

To understand if water and/or water ice was released onto the surface during the impact event, the formation of Lyot crater and the material exhumed as a result of this impact needed to be studied. Through gaining a better understanding of the impact process that formed the crater, the location of deposited subsurface material, and in
particular cryospheric material, can be estimated. Based on these results, it can be predicted if water and/or ice was released and/or exhumed.

Although remote observation provides a lot of information as to the morphology of the crater and ejecta emplacement, it cannot reveal the dynamics of the impact process (Pierazzo and Collins, 2004). There are also issues if an experimental approach were used, although useful for providing information about impact cratering dynamics, experimental techniques are limited in the reproduction of large impacts due to both the influence of gravity and the extreme pressures and temperatures that cannot be replicated in a lab (Pierazzo and Collins, 2004). Fortunately, computer simulations afford the opportunity to simulate the formation of a large impact structure such as Lyot crater (Pierazzo and Collins, 2004). Thus, to develop a better understanding of the impact process that formed the crater, hydrocode impact modelling was used.

Hydrocode simulations were conducted for a variety of different scenarios, and these results were compared to observations from Lyot crater to identify similarities and differences. The scenarios were chosen based upon identification of the most likely parameters from the literature, measurements of Lyot crater’s final topography and morphology, and preliminary mapping and modelling results.

1.2.5 Could Lyot crater preserve material of astrobiological interest?

The overriding aim of this project was to identify whether Lyot crater might preserve material of astrobiological interest, and therefore if it could be a high priority future landing site. To this end, the research outlined by the prior questions was combined to see if there is a link between the landforms and landscapes around Lyot crater and the impact event that formed the crater. If there is, then it is possible that certain landforms are indicative of where subsurface water and/or water ice was deposited. Furthermore, this provides key information as to if material of astrobiological interest might be found at Lyot crater and where it might be located.

1.3 Thesis Structure

The thesis is structured around the research questions presented above. Chapter 1 provides a brief introduction to the reasons why Lyot crater is an important location worthy of further study, and outlines the main research questions. Chapter 2 gives a more in-depth background to the importance of studying Lyot crater, and details previous studies in this region. The research is then placed into the context of the climate, and landforms and
landscapes that formed, during the Amazonian on Mars, by presenting an overview of previous work pertinent to this thesis. **Chapter 3** details the data used in this thesis and the rationale by which the research was approached (i.e. the methodology). Each research chapter within the thesis contains pertinent background information within it, and details the methods applied to answer the research questions.

The research chapters comprise Chapters 4, 5, 6, 7 and 8. **Chapter 4** describes the results of grid mapping of the Lyot crater study area. This method allowed me to locate key landforms and landscapes across the entire study region. In this manner, the spatial distribution of these features could be analysed to see if they correlate with particular locations in and around the crater. Using this work, particular features of interest were chosen to be studied in more detail. **Chapter 5** presents Map Sheet 1, which is a detailed geomorphological map of the interior of Lyot crater. Using this map, and previous research on Lyot crater, I constructed a stratigraphic history of the crater, which enabled me to study the potential timings of water-related activity. Chapters 6 and 7 present work that focusses on particular landforms, and landform assemblages, potentially representative of atmospheric water and more ancient water/water ice respectively. These studies involve the quantitative morphological analysis of these landforms, qualitative observations, and comparison to terrestrial and martian analogues. Using this work, the possible origins of these features are discussed. In **Chapter 6**, landform and landform assemblages potentially indicative of more recent water sourced from the atmosphere are studied in more detail. These landforms are small fluvial channels and associated fan deposits located in the interior and inner ejecta blanket of Lyot crater. **Chapter 7** presents work on enigmatic clastic polygonal landforms located within the outer ejecta blanket of Lyot crater. These landforms are located at a specific radial distance from the crater, indicating a genetic link with the impact and potentially related to the material deposited in this region. This work was published as an article in the peer-reviewed journal ‘Icarus’. **Chapter 8** presents impact models run using iSALE-Dellen (Collins et al., 2016), a code based upon the SALE hydrocode (Amsden et al., 1980). These models are used to test whether the cryosphere could have been penetrated during the Lyot-forming impact event, and where cryospheric material and ancient groundwater could have been deposited. Preliminary 3D models are also presented that test the potential impact angle of the event.

**Chapter 9** combines the results from the five research chapters to present a possible timeline of the history of water within and around Lyot crater, and discusses their
implications for whether this location could contain material of astrobiological interest. **Chapter 10** summarises the research presented in this thesis with emphasis placed on the key conclusions. Avenues of further investigation are also discussed that would build upon the work here contained.
Chapter 2: Background

2.1 Introduction

Many studies of the martian surface indicate that surficial fluvial activity was more prevalent early in martian history (e.g. Hartmann and Neukum, 2001; Carr and Head, 2010; Ehlmann et al., 2011). This activity has since declined and, during the most recent period, known as the Amazonian (~3 billion years ago to present, Hartmann and Neukum, 2001), conditions have remained relatively cold and arid, with abundant evidence of glacial and ground-ice activity (e.g. Head et al., 2006a; Dickson et al., 2008; Head et al., 2009; Soare et al., 2008, 2016; Barrett, 2014), and some evidence of short-duration bursts of fluvial activity (e.g. Berman and Hartmann, 2002; Basilevsky et al., 2006; Fassett and Head, 2008; Dickson et al., 2009; Morgan and Head, 2009; Carr and Head 2010; Fassett and Head, 2010). Figure 2.1 summarises this, showing that whilst fluvial landform formation and activity decline after the Noachian and Hesperian, ice-related activity, such as the development of the polar layered terrains and glaciers, starts to become more prevalent.

Figure 2.1: Modified from Figure 1, Carr and Head, 2010. The diagram shows geological activity as a function of time on Mars. The approximate boundaries of the major martian time periods are

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taken from Hartmann and Neukum, 2001. The Amazonian period is indicated by the dark blue box and the impact event that formed Lyot crater is marked by the yellow and red symbol.

The lack of substantial liquid water on the surface of Mars is problematic in the search for habitable environments, a major focus of many current exploration programmes (e.g. Bada et al., 2005, Vago et al., 2006, Grotzinger et al., 2012; Mustard et al., 2013). This is because one of the fundamental requirements for Earth-like life is the presence of sustained liquid water (Franks, 2000; Westall et al., 2000; Rothschild and Mancinelli, 2001). If life, or evidence of past life, does exist on present day Mars it would likely be in the subsurface within protected environments (McKay and Stoker, 1989; Boston et al., 1992; Weiss et al., 2000; Westall et al., 2000). This is for a number of reasons discussed below.

First, Mars lacks a significant atmosphere (Hess et al., 1980) and, due to the lower amount of ozone present in this atmosphere, UV levels are between 55 and 84% higher than those on Earth (Westall et al., 2000). Therefore, the surface is exposed to high levels of ionizing radiation and, although UV radiation can only penetrate the top few millimetres of the surface, oxidation effects go deeper (Westall et al., 2000; Dartnell et al., 2007). There are also other types of radiation which could have an impact to greater depths, such as solar energetic protons and galactic cosmic rays (Dartnell et al., 2007). Dartnell et al. (2007) show that, in general, the greater the depth, the longer microbial life survives. Hence, life has a better chance of survival at greater depth.

Second, Mars has extensive buried ice that could date as far back as the end of the Noachian (Smith and McKay, 2005). Microorganisms originating from this period would be preserved within ice for substantial periods of time, if sheltered from radiation, due to lower temperatures that prevent thermal decay (Bada and McDonald, 1995; Smith and McKay, 2005; McKay, 2010). Such an ice deposit would be found deep (~1 km) beneath shallow overlaying ice (Smith and McKay, 2005) and therefore within the subsurface.

Finally, life might have originated and been entirely confined to the subsurface where it would have been afforded protection from large impact events (Maher and Stevenson, 1988). As described above, Mars has extensive buried ice deposits. Nearly all water on Mars is theorised to reside within buried or pore-filling ice in a region of crust where temperature remains below the freezing point of water (known as the cryosphere), or as groundwater even deeper in the crust where radiogenic heating leads to temperatures above the freezing point of water (Clifford, 1993; Clifford et al., 2010; Figure 2.2). Such a system is theorised to be global in extent (Clifford, 1993), and the cryosphere is expected to lie at a
depth of 0-9 km at the equator to ~10-22 km at the poles (Clifford et al., 2010). The interaction of impact events with the cryosphere might lead to the generation of hydrothermal systems and the stability of liquid water for up to 10 Ma (Abramov and Kring, 2005). On Earth non-photosynthetic microbial ecosystems have been found to exist near to hydrothermal systems (Boston et al., 1992; Zierenberg et al., 2000) and life might even have originated near to these systems (Baross and Hoffman, 1985; Holm, 1992; Shock, 1996; Martin et al., 2008). This indicates the potential for such subsurface areas on Mars to support life after surface conditions became unsuitable.

**Figure 2.2:** Taken from Figure 1, Clifford et al., 2010. A hypothetical pole-to-pole cross section of the present day martian crust showing the relationship between surface topography (taken from the MOLA Gridded Data Record with a 2° smoothing function), ground ice and groundwater. Where the cryosphere is in contact with the water table, dissolved salts depress the freezing point of the groundwater, and thereby reduces the thickness of frozen ground (Clifford et al., 2010). Where groundwater and the cryosphere are not in direct contact, the leaching of soluble salts by low-temperature hydrothermal convection results in higher basal melting temperatures closer to ~273 K (Clifford et al., 2010).

Based upon these factors, the subsurface environment on Mars appears to be the best location to look for evidence of extant and past life. This presents difficulties for exploration, as the depth to which life could be present exceeds the depth to which current rover designs can drill (Smith and McKay, 2005; Dartnell et al., 2007). Fortunately, a solution to this issue is presented in part by large impact events. Large impact events could penetrate the cryosphere and, as well as providing energy for hydrothermal systems that
could support microorganisms, they could reveal them on the surface within the ejecta blanket (Cockell and Barlow, 2002), and/or flush these microorganisms to the surface during subsequent groundwater release (Westall et al., 2000). The released water will pond in depressions and the organisms could potentially survive until the water freezes leaving them dormant or extinct, yet preserved (Westall et al., 2000). Therefore, large impact craters might preserve material of astrobiological interest on the surface (or near to the surface), in a location where rovers could study it.

It is for this reason that Lyot crater is a location of particular interest, as it is a large impact crater, that cratering mechanics have demonstrated should have penetrated the cryosphere (Russell and Head, 2002). Extensive outflow channels are also present around the crater, which might have formed through groundwater release during the impact event (Harrison et al., 2010; Weiss et al., 2017). In particular, Lyot is the most recent of the large martian impact craters, dating from the Early Amazonian/Late Hesperian (Greeley and Guest, 1987; Werner, 2008; Dickson et al., 2009). Thus it might provide an opportunity to study the youngest sample of cryospheric material present at the martian surface. Finally, Lyot Crater also contains some of the most recent evidence of fluvial activity on the surface of Mars (Dickson et al., 2009; Fassett et al., 2010; Harrison et al., 2010; Hobley et al., 2014), and possible periglacial and/or glacial activity (Dickson et al., 2009; Hobley et al., 2014). The formation of glaciers on Mars is consistent with the accumulation of snow and ice as a result of atmospheric precipitation, probably during periods of high obliquity (Forget et al., 2006; Head et al., 2006a). Therefore, Lyot crater probably records more recent water sourced from the atmosphere, as well as ancient water sourced from the subsurface.

2.2 Lyot Crater

Lyot crater (50°N, 30°E) is a ~215 km impact crater located in the northern latitudes of Mars, north of Deuteronilus Mensae and immediately north of the dichotomy boundary (Russell and Head, 2002; Dickson et al., 2009; Harrison et al., 2010; Hobley et al., 2014; Weiss et al., 2017; Pan et al., 2018). It has a central peak within an inner peak-ring and contains the point of lowest elevation in the northern hemisphere, at ~7000 m (Figure 2.3; Russell and Head, 2002; Dickson et al., 2009; Harrison et al., 2010; Weiss et al., 2017).
Figure 2.3: MOLA topography map showing Lyot crater with outer ejecta extent, inner ejecta extent, crater rim, inner peak-ring and central peak marked.

Lyot has an ejecta blanket consisting of an inner continuous ejecta sheet extending to ~1 crater radii, and an outer more hummocky ejecta sheet extending to ~2.5 crater radii, which includes pristine secondary craters distributed predominantly towards the north (Figure 2.4; Russell and Head, 2002; Harrison et al., 2010; Robbins and Hynek, 2011; Weiss et al., 2017). To the southeast, ejecta is patchy (Russell and Head, 2002); and to the south/southwest, the ejecta blanket is not well preserved (Russell and Head, 2002). This is either a result of the superposition of the later deposits of Deuteronilus Mensae (Russell and Head, 2002; Harrison et al., 2010; Robbins and Hynek, 2011), and/or that ejecta was not deposited in this region due to the impact event being oblique (Russell and Head, 2002; Weiss et al., 2017; Pan et al., 2018).
Figure 2.4: Taken from Figure 1, Robbins and Hynek (2011). THEMIS daytime mosaic showing the distribution of near-field secondary craters around Lyot crater (white circles). The black outline marks the extent of the continuous ejecta blanket.

That Lyot impact material overlies late-Hesperian and early-Amazonian age deposits, indicates that the crater formed within the late-Hesperian/early-Amazonian (Russell and Head, 2002; Dickson et al., 2009; Harrison et al., 2010). Crater counting by Dickson et al. (2009) yields a formation age of ~3.3-1.6 Ga. Lyot Crater includes many geomorphic features which indicate prior fluvial activity (Dickson et al., 2009; Harrison et al., 2010; Hobley et al., 2014), and possible periglacial and/or glacial activity (Dickson et al., 2009; Hobley et al., 2014). Dickson et al. (2009) calculate a Middle Amazonian age for fluvial networks in the area; whereas Hobley et al. (2014) calculate a slightly later age of Late Amazonian for the fluvial systems. In either scenario, Lyot preserves evidence of some of the most recent fluvial channel formation on Mars (Dickson et al., 2009, Fassett et al., 2010).

2.2.1 Mantling Deposits and Possible Bedrock

Lyot crater ejecta material overlies the Hesperian-aged Vastitas Borealis Formation that covers the majority of the northern lowlands (Greeley and Guest, 1987; Russell and Head, 2002). Hobley et al. (2014) interpret stratigraphically-low material, which is rough in
texture, as potential bedrock which has been modified by dust infilling and impact gardening (Figure 2.5). Hydrated mineral signatures including Fe/Mg smectite and chlorite have been found spatially-associated with the bedrock units exposed in the central peak, peak ring and crater rim (Carter et al., 2010; Pan et al., 2017; Pan et al., 2018). Of particular interest is the detection of prehnite close to the crater rim which requires temperatures of 200-350°C and low pressures to form (Pan et al., 2017, 2018). This indicates either that the pre-impact target contained these hydrated minerals as a result of burial diagenesis or hydrothermal activity prior to the formation of Lyot crater, or that the Lyot-forming impact event initiated hydrothermal activity leading to alteration of the bedrock material (Pan et al., 2018). Hobley et al. (2014) suggest that rather than bedrock this material could be an icy regolith, though the topography indicates that true bedrock is at least near the surface.

Figure 2.5: From Figure 6a, Hobley et al. (2014). Partially dissected, flat-lying terrain (blue) infilling topographic lows, representative of an older, degraded mantle. The hills are interpreted as being composed of ‘bedrock’.

Mantling (i.e. deposited so as to ‘drape’ topography) deposits are common in the Lyot area with varying textures and superposition relationships signifying multiple generations of emplacement (Hobley et al., 2014). These textures range from smooth, intact areas pervasive in topographic lows and around breaks of slope, to more heavily-dissected, ‘stippled’ terrains that are probably older, degraded mantling deposits (Dickson et al.,
Hobley et al. (2014) interpret the mantles as ice-rich due to accompanying surface features such as “brain terrain”, a glacial surface texture composed of complex ridge patterns that resemble the patterns on brain surfaces (Levy et al., 2009a), and polygonal ground textures, suggestive of periglacial/ground-ice landforms indicative of the freezing and thawing of ice (Mangold, 2005; Kostama, 2006; Levy et al., 2009b). Degradation is likely the result of ice loss due to sublimation from the ice-rich mantle (Hobley et al., 2014). Dickson et al., (2009) calculate a mid-Amazonian age for the stippled, older mantling unit (~1.5-0.78 Ga) using crater size-frequency distribution measurements. Thus, there is a minimum time lag of 0.8 Ga between the formation of Lyot crater and the deposition of the stippled mantle unit (Dickson et al., 2009). Smoother mantle units are less degraded and are likely representative of the youngest deposits (Hobley et al., 2014; Pan et al., 2018). Pan et al. (2018) interpret these smoother mantle units as mass wasting products derived from uplifted bedrock, as there is evidence of elongate areas of hydrous mineral detections within this unit that are typically representative of the mineralogical signature associated with the possible bedrock units (i.e. crater rim, peak ring and central peak). This suggests that the bedrock units are degrading and debris is falling down slope.

2.2.2 Important Landforms in and around Lyot Crater

Lyot Crater has a variety of geomorphological features of interest. These features provide a window into past fluvial, glacial and/or ground-ice processes active throughout the history of this crater. Below, the key landforms present in the Lyot region are described.

2.2.2.1 Large Outflow Channels

Large outflow channels cover an area of ~300,000 km² and extend >300 km beyond the ejecta margins to the north, west and east of Lyot crater (Figure 2.6; Harrison et al., 2010). These channels originate from the ejecta margins and are not observed incising the ejecta material (Harrison et al., 2010; Weiss et al., 2017). Channel widths range from ~0.1-1 km with relatively shallow gradients of ~0.07° (Harrison et al., 2010). There are large areas in which the channels are not well defined that span >30 km (Harrison et al., 2010; Weiss et al., 2017). The channels are shallow (metres in depth), follow regional slope, and divert around local topography (Harrison et al., 2010; Weiss et al., 2017). Harrison et al. (2010) suggest that the channels are unlikely to predate the impact event due to their close geographic relationship to Lyot crater, and a lack of morphologically similar channels elsewhere in the northern plains.
Figure 2.6: From Figure 1, Harrison et al. (2010). **A**) Lyot Crater and surrounding terrain (MOLA gridded topography (128 px/deg) over MOLA shaded relief map). The Lyot primary ejecta blanket is outlined in white. Thick black lines denote channels and areas of channel scour, while dotted lines delineate candidate areas of channel scour (mapped using MRO CTX and THEMIS images). Yellow boxes mark the locations of Figures b–d (included) and Figures 2a–2b from Harrison et al., 2010. Contour lines denote 250 m intervals. **B**) Braided channels north/northwest of Lyot. **C**) Detail of channels. **D**) CTX image showing an area of channel scour with superposed pedestalized craters northeast of Lyot.
Fluvial Channels, Ridges and Fans

Smaller fluvial channels are commonly located in the interior of the crater (Figure 2.7; Dickson et al., 2009; Hobley et al., 2014), though some are located within the inner ejecta blanket (Hobley et al., 2014). Most originate within the crater rim or central peak ring (Balme et al., 2013a). Dickson et al. (2009) interpret channels as incising the stippled mantling unit, which has a mid-Amazonian age (~1.5 – 0.78 Ga) calculated from crater counting. Dating of the channels by Hobley et al. (2014) indicate that they formed in the late-Amazonian (116 – 29 Ma), though they postulate that channels formed under a thick ice cover (discussed in section 2.2.4), and thus this age could represent the deglaciation age, not the channel formation age. Either way, there is a significant time lag between the formation of Lyot crater and the onset of the fluvial activity which formed the channels.

Channels are commonly small, with lengths ranging from 2-50 kilometres and typical widths of ~250 metres (Dickson et al., 2009). They are usually simple and non-branching, though some contributory networks exist (Balme et al., 2013a). Dickson et al. (2009) measured the slopes associated with twenty observed channel systems, finding slopes ranging from 0.4-6°, with a median of 2°. However, Hobley et al. (2014) found channel slopes of 5-6° for Lyot crater, indicating that average channel slopes might be steeper. Dickson et al. (2009) interpret channels as following local slopes, whereas Hobley et al. (2014) found that some channels go against local slope gradients and cross topographic divides.

Several channels have fan deposits at their termini, which have been interpreted as alluvial fans (Dickson et al., 2009). These deposits vary from broad smooth surfaces, to smaller dissected fans (Dickson et al., 2009). Some channel terminations are associated with dune fields (Balme et al., 2013a); this indicates the presence of sand-grade material. Sand is often transported via rolling and sliding along the bottom of fluvial channels or saltation within a fluid (Pettijohn et al., 2012), so is an indication of fluvial deposition in certain environments.
Figure 2.7: HiRISE image (ESP_016339_2295) showing a rare contributory channel from the south-east interior of Lyot crater incised into the unit identified as stippled mantle by Dickson et al. (2009). The red box within the inset map indicates the extent of the main image.

A few sinuous ridges are observed within Lyot Crater (Hobley et al., 2014), commonly terminating within closed basins or inverted deposits, and which could be interpreted as inverted channels (Balme et al., 2013a). One example described by Balme et al. (2013a), shows branching and potential braiding, and terminates in a broad, possibly delta-like deposit, associated with and superposed by dunes. Hobley et al. (2014) describe an assemblage of channels connecting to ridges which terminate in mounds (i.e. fans). They interpret the assemblage as a sub-ice sheet drainage network in which ridges represent eskers, and fan deposits form from the unconstrained flow of water onto the surface of Mars freezing in layers (Hobley et al., 2014) – similar to the mechanism through which terrestrial aufeis form (e.g. Kane, 1981; Hodgkins et al., 2004).

Dickson et al. (2009) interpret these channels as being formed by fluvial processes as they display meandering, and they measured typically low slopes, consistent with fluvial incision. They also noted a direct association between viscous flow features and the channels (Dickson et al., 2009), though this association has not been observed in other studies of the channels (Balme et al., 2013a; Hobley et al., 2014). Hobley et al. (2014) also assume a fluvial origin for the channels. It is important to note that little to no evidence for
lateral migration of the channels has been observed (Hobley et al., 2014), potentially indicating that channel formation was short-lived, or that the channel banks were erosion resistant.

2.2.2.3 Viscous Flow Features (VFFs)

Lobate features with convex-outward ridges and convex-upward profiles can be seen predominantly in the south of Lyot Crater along the rim and peak ring (Figure 2.8; Dickson et al., 2009). VFF display parallel or transverse surface lineations and are associated with compressional or extensional troughs (Hobley et al., 2014). These features are interpreted as debris-covered glacier deposits, similar to those mapped in Deuteronilus Mensae (Head et al., 2006a, 2006b; Morgan et al., 2009; Dickson et al., 2009). Some channels have been seen associated with them (Dickson et al., 2009), but this association is not strong enough to definitively indicate proglacial fluvial activity (Balme et al., 2013a). A lack of fluvial channels associated with such features is expected to be due to low ice temperatures and low heat available for melting (Fassett et al., 2010).

Figure 2.8: CTX mosaic image showing a VFF, interpreted by Dickson et al. (2009) as a debris-covered glacier, from the southern rim of Lyot crater. The red box within the inset map indicates the extent of the main image.
2.2.2.4 Polygonal patterned ground

Clastic polygonal networks are common within the innermost area of the outer ejecta blanket (Figure 2.9; Balme et al., 2013b). They are limited to the north and south east regions of the outer ejecta and appear to occur on “hills” rather than within lower topography depressions (Balme et al., 2013a). The polygons are generally 4 or 5 sided and 100 – 200 metres in diameter, with clasts ranging up to several meters in size (Balme et al., 2013a). The process by which these structures formed is unknown. Such features are commonly associated with periglacial processes, indicating the freezing and thawing of ground ice (Lachenbruch, 1962; Kessler and Werner, 2003). Their clastic form might be the result of freeze-thaw cryoturbation processes (MacKay, 1984; French, 2007) or boulder clustering due to a “ratcheting” process, which is the result of boulders becoming locked within a CO$_2$ frost layer (Orloff et al., 2013). Alternatively, the organised clasts could be the result of gravitational slumping whereby boulders fall into thermal contraction cracks (Levy et al., 2010). The mechanism by which they form is uncertain. The location of the features is strongly associated with ejecta material, indicating that the conditions for their formation are only met within this area/material (Balme et al., 2013a), suggesting a genetic link with the crater ejecta.

Figure 2.9: HiRISE image (ESP_016985_2315) showing a clastic polygonal network from the east of the outer ejecta blanket of Lyot crater. The red box within the inset map indicates the extent of the main image.
2.2.2.5 Boulder Fields

Boulder fields can be found within the outer ejecta blanket as low albedo, rough-textured areas in CTX and HiRISE images (Figure 2.10; Balme et al., 2013a). They occasionally show texture that could be organised patterns, perhaps indicating periglacial reworking of the clasts (Washburn, 1956; Feuillet et al., 2012; Balme et al., 2013a). These features are not controlled by topography and are associated with low thermal-inertia units (Balme et al., 2013a).

Figure 2.10: HiRISE image (ESP_017842_2350) showing a boulder field from the north-east of the outer ejecta blanket of Lyot crater. The red box within the inset map indicates the extent of the main image.

2.2.2.6 Other landforms

Other landforms found within Lyot Crater include dune fields (Figure 2.11a; Balme et al., 2013a; Hobley et al., 2014) gullies (Figure 2.11b; Hart et al., 2009) and a possible thermokarst paleolake assemblage (Figure 2.11c; Glines and Gulick, 2018).

Dunes indicate the presence of sand-grade material, which is potentially indicative of deposition within certain fluvial or aeolian environments, followed by aeolian reworking (Pettijohn et al., 2012). They are located mainly within the centre and southern half of the crater (Hobley et al., 2014). Dune fields have been observed at the termini of a number of channels (Balme et al., 2013a), in agreement with the idea that the source for the aeolian
dunes are degrading fluvial deposits. Dune gullies, potentially formed as a result of seasonal CO$_2$ frost (e.g. Diniega et al., 2010) or dry granular flows (e.g. Horgan and Bell, 2012), have been observed in the Lyot dune fields (Diniega, 2010; Widmer and Diniega, 2018).

Hart et al. (2009) studied two major gullies on the central peak of Lyot crater. Gully channel widths range from 5.5 – 30.5 metres, with depths of 0.7 – 2.7 metres and a meander wavelength of between 60 – 120 metres (Hart et al., 2009). Slopes for the gullies were calculated: 16.4 – 29.7° for the alcoves, 11.3 – 20.6° for the channels, and 8.4 – 16.9° for the debris aprons (Hart et al., 2009). This demonstrates that the larger channels on the floor of Lyot crater are far shallower than the gully features. Gullies might form as a result of water sourced from the melting of frost or snow (e.g. Head et al., 2008, Conway et al., 2011, de Haas et al., 2015), but other authors suggest the action of granular flow, triggered by CO$_2$ frost, is the erosive mechanism (e.g. Dundas et al., 2015, 2017; Sylvest et al., 2016, 2018).

Glines and Gulick (2018) describe circular depressions connected by possible channels located near to the central peak of Lyot crater. The circular depressions have an average diameter of ~200 metres, connecting channels appear to outline polygons of 500 – 1000 metres across which could form through thermal contraction (Glines and Gulick, 2018). This assemblage is interpreted by Glines and Gulick (2018) as beaded streams. Beaded streams are common thermokarst features in permafrost regions of the Arctic (Arp et al., 2015). Such landforms represent depressions formed as a result of the thaw of large ice masses which are connected by streams following ice-wedges (Hopkins et al., 1955).
Figure 2.11: The red boxes within the inset maps indicate the extent of the main images. (a) CTX mosaic image of a large dune field in the interior of Lyot crater. (b) HiRISE image (PSP_009245_2310) of gullies located on the central peak. The downslope direction is from the bottom to the top of the image. (c) CTX mosaic image of a possible thermokarst paleolake assemblage in the interior of Lyot crater. The downslope direction is from the top to the bottom of the image.
2.2.3 Origin of the Hesperian Fluvial Activity – Large Outflow Channels

The large outflow channels shown in Figure 2.6 might have formed as a direct or indirect result of the impact event (Russell and Head, 2002; Harrison et al., 2010; Weiss et al., 2017).

2.2.3.1 Was deep groundwater released by the impact?

Under current conditions, a zone of crust in which temperature is below the freezing point of water could extend from the surface to an unknown depth dependent upon megaregolith characteristics (Clifford, 1993; Russell and Head, 2002; Clifford et al., 2010). Below this depth, liquid water is stable (Clifford, 1993; Russell and Head, 2002), and it is inferred through analogy to terrestrial, large-scale permeability that global-scale communication of groundwater to depths of ~26 km could occur on Mars (Clifford and Parker, 2001; Russell and Head, 2002). Hence, Clifford (1993) infers the presence of a global, interconnected groundwater system, confined beneath an ice-rich cryosphere. If an event disrupts the cryosphere, water is predicted to flow to the surface under artesian-like conditions, i.e. the water flows to the surface naturally due to hydraulic pressure (Russell and Head, 2002).

Impacts can cause such a disruption as they provide heat and excavation of material to considerable depths (Russell and Head, 2002). The larger the diameter of the impact crater, the greater depth from which ejecta material is excavated (Melosh, 1989). It is also important to note that a theoretical cryosphere is predicted to mirror surface topography to the first order (Clifford, 1993; Russell and Head, 2002), it is probable that the ice-rich zone would be thinner in areas of low elevation (Russell and Head, 2002) and the depth to a water table in hydrostatic equilibrium would be less (Clifford and Parker, 2010). Furthermore, the cryosphere has evolved over time with the estimated volume of water assimilated by the cryosphere increasing as crustal heat flow declined, thus the Amazonian cryosphere likely contains the greatest volume of water in Mars’ evolutionary history (Clifford and Parker, 2001; Russell and Head, 2002). Lyot crater is the lowest point in the northern latitudes, dated as Amazonian in age, and has a diameter of ~215 kilometres (Dickson et al., 2009; Russell and Head, 2002; Harrison et al., 2010; Weiss et al., 2017; Pan et al., 2018). Russell and Head (2002) therefore used the crater to test the Clifford (1993) hydrological model.

They predicted that during the formation at Lyot, the zone of physical disruption would reach to a depth of ~33 km and convert about 35% of the projectile kinetic energy.
into heat, leading to impact melting (Russell and Head, 2002). This should be sufficient to penetrate or destroy a cryospheric layer of nominal thickness, leading to liberation of groundwater to the surface (Russell and Head, 2002). Ejecta is estimated to have been excavated from a depth of ~11-24 kilometres (Russell and Head, 2002; Pan et al., 2018). The crustal composition used for this study was based upon Clifford (1993), who summarises the average Martian megaregolith and upper crust as impact ejecta interbedded with lava flows, sediments, and weathered materials, overlying a breccia composed of fractured basement rock (Clifford, 1993; Russell and Head, 2002). This might not be a suitable representation within the northern latitudes, but Russell and Head (2002) suggest that it is reasonable due to evidence for underlying Noachian heavily cratered terrain.

Another important point to note is that Russell and Head (2002) did not find evidence of fluvial activity associated with Lyot Crater in the manner expected within this model. Evidence is not seen within the context of their study for fluvial erosion, flooding and sedimentation expected to accompany such a breach of the cryosphere (Russell and Head, 2002). More recent work by Harrison et al. (2010), however, describes large outflow channels (see section 2.2.2.1), which provides evidence for fluvial erosion, though not evidence for the ponding of water in the crater interior which would indicate significant artesian release after the impact. Weiss et al. (2017) also conclude that there is no evidence of groundwater inflow into the crater interior and thus a deep groundwater origin of the large channels within Lyot is not supported by these studies.

2.2.3.2 **Do the outflow channels form as a result of water sourced from underground?**

Harrison et al. (2010) invoke the mobilisation of groundwater through either the seismic energy of the impact event, or by dewatering of the ejecta blanket as possible mechanisms for the formation of the large outflow channels. This conclusion stemmed from their interpretation that the braided morphology of the outflow channels indicates sediment-laden water, and the location of the channels relative to Lyot crater shows that their formation is associated with the crater (Harrison et al., 2010). Weiss et al. (2017) assess a number of formation mechanisms for the outflow channels; these mechanisms are discussed below (Figure 2.12).
Figure 2.12: Modified from Figure 2, Weiss et al. (2017). Schematic representation of various possible formation mechanisms for the large outflow channels around Lyot crater. These mechanisms include: (A) deep groundwater release from penetration of the cryosphere; (A) shallow groundwater release due to the mobilisation of water from seismic energy; (B) rainfall precipitation due to condensation of the impact vapour plume; (C) dewatering of the Lyot crater ejecta blanket; (D) melting of ground ice by hot ejecta; and (D) melting of surficial snow and ice by hot ejecta (Weiss et al., 2017).

Impact events impart large amounts of energy which inevitably leads to seismic activity within the impacting site. Seismic activity can trigger the release of groundwater through either ejection of groundwater under pressure, or unconstrained upwelling from a shallow water table (Harrison et al., 2010). These scenarios produce different features due to differences in pressure. In the case of confined pressured fluids, more explosive features will be observed such as fissures, mud volcanoes and sedimentary dykes (Harrison et al., 2010). This is different for fluids which are unconstrained, as in these cases, water-sediment flows will probably be non-eruptive (Harrison et al., 2010). It has been inferred that due to a lack of fissures and cones, water ejection was non-explosive within Lyot crater, and therefore unconstrained upwelling is more likely (Harrison et al., 2010). Wang et al. (2005) showed that lithostatic pore pressures created during liquefaction by undrained compaction of soils during the impact event might be sufficient to cause significant enough groundwater release to form outflow channels, but Weiss et al. (2017) suggest that it is unlikely that unconfined groundwater would have been present and stable.
and that a depressed freezing point of \(~195\text{ K}\) would be required, which they found physically implausible. Hence, a seismically-driven discharge of shallow groundwater was deemed unlikely by Weiss et al. (2017), but this hypothesis was favoured by Harrison et al. (2010).

Another hypothesis considered by Harrison et al. (2010) is the dewatering of pore water within the ejecta blanket. This hypothesis is based on the dewatering of terrestrial large volcanic landslides (Harrison et al., 2010). In such events, residual water within a body of flow is either forced upwards as a result of compaction and pore pressure to form surficial flows, or travels along the basal interface to emerge as a hyperconcentrated flow (Harrison et al., 2010). Such dewatering streams are “load-rich, discharge-poor and prone to braiding” (Harrison et al., 2010), which matches observations of the large channels around Lyot crater. However, this hypothesis is not favoured by Harrison et al. (2010) due to a lack of evidence of undermining and collapse within the ejecta blanket, as well as the observation of fully formed channels which cover a significant area near to the ejecta margin. Weiss et al. (2017) disagree with this conclusion, stating that undermining and collapse features do not need to be present to indicate flow beneath ejecta. This is because in their study they found that virtually all ejecta pore ice would have been melted by shock heating, indicated by model ejecta temperatures of \(~400\text{ K}\) (Weiss et al., 2017). They then conclude that this meltwater would flow along the basal interface and flow from the ejecta margins. Thus they consider dewatering of the ejecta blanket possible (Weiss et al., 2017).

Weiss et al. (2017) infer that surface ice was present at the time of the Lyot impact event due to the location of Lyot near to known ice deposits, the association of the crater with young glacial features, and the paucity of secondary craters in certain locations around the crater (inferred to be a result of secondary impacts into later-removed ice rather than ‘permanent’ martian surface materials). Hot ejecta superposing these ice deposits could lead to contact melting of up to 330 metres depth of ice, producing a maximum meltwater volume of \(~7570\text{ km}^3\) (Weiss et al., 2017). Harrison et al. (2010), however, deemed this hypothesis unlikely, as outflow channels theorised to form in this way around Sinton crater (Morgan and Head, 2009) incise the ejecta blanket and originate from the rim of the crater. Neither of those characteristics are observed at Lyot crater. Weiss et al. (2017) argue that the thinner ejecta layer around Sinton crater makes it more likely that surficial channels are expressed and, as Sinton crater sourced ejecta material from a shallower depth, this ejecta material contains a higher proportion of pore ice
making channel formation more likely. Furthermore, they argue that the outflow channels are morphologically similar to the channels around Lyot crater (Morgan and Head, 2009; Weiss et al., 2017). Thus they infer that the lack of channels incising the ejecta blanket does not preclude the formation of channels at the ejecta margins (Weiss et al., 2017). They consider contact melting of surficial ice/snow deposits to have a great relative importance in the formation of the Lyot outflow channels (Weiss et al., 2017).

Impactors above 2 kilometres in diameter are expected to lead to impact-induced rain as a result of suspended rock vapour and volatiles vaporised during the impact event being injected into the atmosphere and, eventually, condensing and precipitating out (Segura et al., 2002, 2008; Harrison et al., 2010; Weiss et al., 2017). Harrison et al. (2010) suggest that the lack of tributaries and abrupt channel heads indicate this hypothesis is unlikely. Furthermore, Weiss et al. (2017) also deem impact-induced rainfall as unlikely to have occurred at Lyot crater due to the low atmospheric pressures in the Amazonian promoting vapour plume dispersal at regional/global scales. Thus impact-induced rainfall is an unlikely explanation (Harrison et al., 2010; Weiss et al., 2017).

Finally, Weiss et al. (2017) discount hypotheses involving atmospherically-derived rainfall, top down heating of snow and ice by the impact vapour plume, and contact melting of permafrost. This is because (i) atmospherically-derived rain is considered unlikely within the hyperarid Amazonian climate, (ii) a thin surface lag deposit is deemed likely to prevent top-down melting of an ice layer by a hot vapour plume, and (iii) a combination of insufficient melt volumes and there being a lack of mechanisms by which water could have been discharged to the surface. Hence, contact melting of permafrost is seen as unlikely (Weiss et al., 2017). Therefore, the favoured hypotheses of Weiss et al. (2017) are: dewatering of the ejecta blanket and the melting of surficial snow/ice deposits by the superposition of hot ejecta. Conversely, Harrison et al. (2010) favour channel formation as a result of the release of groundwater triggered by seismic activity.

2.2.4 Small Channel Formation – Open or Closed?

The simplest explanation for the smaller channels within Lyot Crater is that they formed sub-aerially by fluvial incision, and that the water was sourced from the melt of subsurface or surface ice (Dickson et al., 2009). This is inferred from observations that channels follow local gradients and appear to be directly associated with viscous flow features (Dickson et al., 2017). Hobley et al. (2014), however, suggest that the channels were not open to the atmosphere, but rather existed under a layer of ice and were
therefore pressurized. They came to this conclusion due to a regional slope analysis of channels, which indicated flow against local slope, and observations of drainage patterns compared with digital terrain models of varying resolutions, which showed tens of channels crossing topographic drainage divides (Hobley et al., 2014).

In this uphill flow scenario, hydrostatic pressure would need to be high, which cannot occur within open channels. Hobley et al. (2014) present several scenarios to achieve pressurised flow: (i) flow occurred within the subsurface, (ii) flow occurred within an ice carapace, or (iii) flow occurred beneath a regionally extensive ice sheet (Hobley et al., 2014).

Hobley et al. (2014) show that formation in the subsurface is unlikely due to a lack of evidence of landforms such as collapse pits and note that there is no sufficient explanation as to how channels in such a scenario could meander (Hobley et al., 2014). Flow beneath an ice carapace is also unlikely due to implications that a cap would need to be in place prior to fluvial activity to allow pressures to build up for the water to flow upslope. It would be physically implausible that high pressures could be reached in a scenario where a carapace forms after flow has begun (Hobley et al., 2014). Therefore, the regional ice sheet model is their favoured explanation. A benefit of this scenario is that a layer of ice would increase the partial pressure of water vapour, leading to higher stability of liquid water (Hobley et al., 2014).

Although this form of channel formation should not be ruled out, it is highly dependent upon the results of the slope analysis. Thus, more work needs to be done verify the slope trends. Finally, other evidence for a regional ice sheet is not provided within the study, and it remains to be tested whether similar channels occur in surrounding areas, as would be expected due to the size estimated for a regional ice sheet.

2.2.5 Origin of Amazonian Fluvial Activity – Small Channels

Hypotheses for the source of heat which generated the meltwater that formed the young, small channels within the Lyot region fall into one of two groups: (i) fluvial activity resulted from a global climate excursion (Hobley et al., 2014), or (ii) fluvial activity was more common in general in Lyot due to a local microenvironment, different to the general Mars climate, which was more conducive to ice melt (Dickson et al., 2009; Fassett et al., 2010).
2.2.5.1 Global Climate Change?

Hobley et al. (2014) consider melting of the theorised ice sheet as the source of flowing liquid water. They consider a variety of causes of heat for melting including: (i) orbitally driven insolation pattern changes, (ii) atmospheric change due to impacts, volcanism or insolation leading to warming, (iii) changes to geothermal flux caused by geologic changes such as movement of magma deep underground, (iv) additional geothermal heat flux to the surface as a result of an impactor, and, (v) thickening of a covering snowpack until average geothermal fluxes melt the base (Hobley et al., 2014).

Unlike in the previous studies discussed, impacts releasing groundwater are not considered a suitable mechanism for water release within the crater, as the channels are carved into terrains much younger than the Lyot impact. However, impact heating was considered, but ruled out by Hobley et al. (2014), due to the location of channels across the site; they consider impact driven heating to be limited purely to the interior and rim of the crater, which would be unlikely to cause the formation of channels observed up to 180 kilometres from the rim of Lyot. This, combined with the significant time lag between the impact event and channel formation, suggest formation of channel-forming melt water due to impact heating was unlikely (Hobley et al., 2014). Changes in geothermal flux due to the movement of magma is also ruled out due to a lack of evidence for more recent igneous activity that would be expected with such an intrusion of material (Hobley et al., 2014).

The mechanism of heating beneath a snowpack is considered a possible contributing factor to melting, but not a sole mechanism. This is due to the thicknesses of snow and ice required for significant melting being too large when compared to thickness estimates of the predicted regional ice sheet (Hobley et al., 2014). Hence, global climate change is the preferred source of heat in Hobley et al. (2014). Further differentiation as to whether this climate change was orbitally driven or due to a differing atmospheric environment is not conclusive. Hobley et al. (2014) infer that the morphology of channels indicates a short, single episode of fluvial activity; which might indicate an atypical warming event such as a particularly extreme orbital cycle or formation under a volcanically-forced Amazonian atmospheric excursion. In either case it would be expected that further examples of such channel systems within the Amazonian would also be observed.

2.2.5.2 Is Lyot crater’s climate unique?

Dickson et al. (2009) propose that glacial deposits in Lyot are the likely source of water, interpreting a direct relationship between glacial deposits and fluvial valley systems. In this
case, the source of energy for melting such deposits must be considered. Similarly to Hobley et al. (2014), Dickson et al. (2009) state that the time lag between the impact event and channel formation rules out the impact event as a potential source. Climatic conditions are instead invoked as a possible explanation.

Due to the low elevation of the crater, pressures above 6.1 mb – which is the triple point of water – have been predicted for current day conditions (Dickson et al., 2009). The low elevation, high pressure conditions are not enough as a sole mechanism for melting, as THEMIS daytime IR data shows a maximum temperature of ~260 K in present day conditions (Dickson et al., 2009). Therefore differences in obliquity are invoked to provide further solar insolation changes as an energy source for melting (Dickson et al., 2009).

Climate models indicate that, at higher obliquities, mean annual temperature would increase in the latitude of Lyot, with peak temperatures of above 273 K (Costard et al., 2002; Dickson et al., 2009). Further to this, precession of Mars orbital rotation has led to epochs when northern summers are warmer than southern summers (the opposite case to today) which would increase temperatures further (Dickson et al., 2009). This temperature increase, combined with the high-pressure conditions within Lyot, could have led to melting of surface and near-surface ice deposits, leading to channel formation (Dickson et al., 2009). Unlike a global climate change, this process would not lead to widespread or regional channel systems as melting requires a more unique set of parameters found only within low elevation, high pressure areas (Dickson et al., 2009).

Although this model for the origin of water within the Lyot arguably better explains the features present, and has little contrasting evidence, one unknown is whether glacier-like deposits correlate spatially with the source of channel systems. According to Balme et al. (2013a) there does not appear to be a significant correlation, as very few channels begin beneath or at glacier-like forms. This implies that the glaciers present today might not be the source of the water that carved the channels.

2.3 Mars During the Amazonian

Lyot crater formed during the Amazonian period on Mars, however the landforms and landscapes present within and around this crater are atypical of those found within the majority of the martian surface during this time period. It is therefore important to provide context for this thesis by reviewing our current understanding of Mars during the Amazonian.
The Amazonian period on Mars began around 3 billion years ago and continues to the present day. During this period, Mars has been characteristically arid and cold with conditions in which liquid water could be stable on the surface existing only in a few regions and for short periods of time (Haberle et al., 2001). As a result of these conditions, fluvial activity is rare and the majority of active processes on Mars are related to the actions of wind or ice. However, there are indications that the climate of Mars has varied significantly during the Amazonian as, though rare, we still find evidence of water-related landforms that require larger quantities of water to form, such as Gulick and Baker (1989, 1990), Fassett et al. (2010) and Balme et al. (2013b), as well as the deposition of extensive ice deposits in the recent past, such as Kreslavsky and Head (2002a) and Head et al. (2006a, 2006b). Such landforms and landscapes provide important windows into the processes that have occurred during the Amazonian and are key to our understanding of the martian climate. Below I will summarise the key activity that has occurred during the Amazonian period which is pertinent to this thesis.

2.3.1 Fluvial Processes

As fluvial channels within Lyot crater formed during the Amazonian, it is important to provide context for the potential origins of fluvial activity within an environment that is analogous to a hyperarid cold desert, such as can be found in the Antarctic Dry Valleys on Earth (Marchant and Head, 2007). Fluvial channel formation during this period of time is rare, with only 3% of mapped channels occurring on Amazonian terrain (Figure 2.13; Hynek et al., 2010). Fluvial channels (or valleys) are generally found in areas with favourable local environmental conditions (e.g. Head et al., 2006b; Dickson et al., 2009; Fassett et al., 2010; Howard and Moore, 2011; Morgan et al., 2011), water-sources mobilised by impact craters (e.g. Cabrol et al., 1999; Morgan and Head, 2009; Kite et al., 2011; Mangold, 2012; Hauber et al., 2013), or relatively recent volcanic activity (e.g. Gulick and Baker, 1989, 1990; Gulick, 2001; Berman and Hartmann, 2002; Hynek et al., 2010). The majority of Amazonian-aged channels are distributed around Hellas Basin and Alba Patera (40°S and 50°N respectively; Carr, 1995; Hynek et al., 2010).
Figure 2.13: Modified from Figure 1, Hynek et al. (2010). MOLA shaded relief map and topography from high (red) to low (blue) displaying the locations of mapped fluvial valleys. The colours of the mapped valleys represent the inferred ages, with Amazonian channels mapped as blue, Hesperian mapped as purple and Noachian mapped as red. AP = Alba Patera, AT = Arabia Terra, HB = Hellas Basin and NT = Noachis Terra.

Amazonian-aged fluvial channels have a lower overall drainage density when compared to overall drainage densities for the Hesperian and Noachian (Hynek et al., 2010). However, individual networks can have high drainage densities, comparable with Noachian values. Alba Patera valleys, for example, have measured drainage densities of 0.0196 km$^{-1}$ (Hynek et al., 2010) or up to 0.3-1.5 km$^{-1}$ (Gulick and Baker, 1989, 1990) and have been dated to the Early Amazonian (1.74-3.52 Ga) based on buffered crater counting (Fassett and Head, 2008). Such significant networks within the Amazonian are unusual and morphologically resemble terrestrial stream valleys such as those located on Hawaiian volcanoes (Gulick and Baker, 1989, 1990; Gulick, 2001). It might typically be expected that young fluvial channels would have had less time to form than the older Noachian valleys and would therefore be smaller and less well-developed (Gulick, 2001). Fluvial valleys not associated with volcanoes are typically far smaller than more ancient valleys, less well-integrated, and associated with cold-climate landforms (Morgan and Head, 2009; Fassett et al., 2010; Hobley et al., 2014; Wilson et al., 2016). The channels found within and around Lyot crater fall into the latter category.

The majority of potential fluvial activity during the Amazonian is related to the formation of gullies – though whether these very young features formed as a result of the actions of water is a point of contention (Musselwhite et al., 2001; Treiman, 2003; Pilorget and Forget, 2016; Sylvest et al., 2018). Gullies have an upper alcove which feeds a channel that terminates in a depositional fan, they are typically hundreds of metres to kilometres...
in length (Heldmann and Mellon, 2004; Heldmann et al., 2005; Balme et al., 2006), and tens of metres wide (Heldmann and Mellon, 2004; Heldmann et al., 2005). Gullies are preferentially located on poleward-facing slopes in the mid-latitudes (Balme et al., 2006; Dickson et al., 2007; Head et al., 2008; Dickson and Head, 2009), but have less preference in orientation, or are more commonly equatorward-facing, elsewhere (Heldmann and Mellon, 2004; Kneissl et al., 2010; Conway et al., 2017). The distribution of gullies provides potential evidence that their formation is linked to areas of preferential ice deposition (Balme et al., 2006; Dickson et al., 2007; Head et al., 2008; Dickson and Head, 2009).

2.3.2 Ice-Related Activity

Ice has had a significant role in shaping the martian landscape during the Amazonian. The vast majority of non-aeolian processes that have occurred within this period of time are ice-related. In fact, ice could be the source of melt for the formation of (potential) fluvial landforms discussed above and within this section (e.g. Dickson et al., 2007; Head et al., 2008; Fassett et al., 2010; Gallagher and Balme, 2015). For instance, the distribution of gullies in the mid-latitudes of Mars is often linked to a geologically recent accumulation of volatiles followed by melting driven by orbital variations (Dickson et al., 2007; Head et al., 2008; Dickson and Head, 2009; Kneissl et al., 2010; Conway et al., 2018).

There is also very rare evidence of wet-based glaciation occurring during the Amazonian (Gallagher and Balme, 2015; Butcher et al., 2017). One of the best candidates for wet-based glaciation is an assemblage described by Gallagher and Balme (2015). This assemblage consists of sinuous ridges in the southern Phlegra Montes region of Mars that are interpreted as eskers emerging from the terminus of a degraded glacier like form, alongside other surface textures and landforms indicative of glaciation (Figure 2.14; Gallagher and Balme, 2015). Such melt is interpreted as resulting from bottom-up geothermal heating as opposed to resulting from climate change (Gallagher and Balme, 2015).

Another example of potential wet-based glaciation during the Amazonian is a sinuous ridge, interpreted as an esker, that emerges from a possible debris-covered glacier located in Tempe Terra, north-east of the Tharsis volcanic province (Butcher et al., 2017). Similar to the other example, this assemblage occurs alongside other surface textures and landforms (such as possible moraines) indicative of glacial activity (Butcher et al., 2017). Another important similarity is that both features are located in graben-like rifts which could enhance geothermal heat flux, and thus both episodes of wet-based glaciation are attributed to enhanced geothermal heat flux (Gallagher and Balme, 2015; Butcher et al.,
Based on the sparse current evidence, wet-based glaciation within the Amazonian is therefore limited to locations where glaciers occur in association with a local environment that enhances geothermal heat flux, leading to bottom-up melting.

Figure 2.14: Taken from Figure 11, Gallagher and Balme (2015). CTX mosaic draped over 50 m HRSC topographic data showing the assemblage of landforms and surface textures indicative of a wet-based glacial system in southern Phlegra Montes, Mars.

The vast majority of landforms interpreted as glaciers, or indicative of glacial-flow, appear to be cold-based (Head et al., 2005, 2006a; Millkovich et al., 2006; Souness and Hubbard, 2012; Dickson et al., 2010, 2012). Subsurface radar sounding data from SHARAD confirm that these deposits are rich in water ice (Holt et al., 2008; Plaut et al., 2009). Such deposits are extensive within the mid-latitudes of Mars (Head et al., 2005, 2006a, 2006b, 2009; Holt et al., 2008, Plaut et al., 2009; Fastook and Head, 2014). Therefore, a large volume of ice deposition has occurred within a geologically recent time period, at latitudes below which ground ice is presently theoretically stable (Mellon and Jakosky, 1995; Forget et al., 2006). However, modelling indicates that at higher obliquity, such ice deposition can occur, thus mid-latitude glaciation provides another example of orbital variations driving climate change on Mars (e.g. Jakosky and Carr, 1985; Forget et al., 2006; Madeleine et al., 2009).

Another common type of ice-rich deposit thought to be deposited during the Amazonian, are the mantling deposits. Such deposits drape topography, contain textures indicative of the sublimation of ice, and are estimated to be metres-thick (Kreslavsky and Head, 2002a; Kostama et al., 2006; Schon et al., 2012). Mantling deposits have been observed covering at least 23% of surface within the middle to high northern and southern
latitudes of Mars (Kreslavsky and Head, 2002a). This mantle is interpreted as representing ice-rich material deposited due to changes in latitudinal insolation controlled in turn by differences in obliquity (Kostama et al., 2006; Schon et al., 2012; Dickson et al., 2015). It represents some of the youngest material deposited on Mars, as indicated by crater counting (Kostama, 2006; Schon et al., 2012).

Ice-rich material on Mars contains abundant landforms and surface textures indicative of the sublimation of volatiles, and some landforms potentially indicative of the actions of freeze-thaw processes. Studies of such landforms and surface textures are too numerous to discuss in-depth and features pertinent to the Lyot crater study area are summarised within Chapter 4. Some common examples of sublimation features include: sublimation pits (e.g. Head et al., 2005; Kadish et al., 2008) and scallops (e.g. Lefort et al., 2010; Ulrich et al., 2010). Landforms potentially indicative of thaw-related processes include: sorted patterned ground (e.g. Balme et al., 2009; Barrett, 2014), solifluction lobes (e.g. Gallagher et al., 2011; Johnsson et al., 2012), thermokarst terrain (e.g. Costard and Kargel, 1995; Soare et al., 2008), and pingos (e.g. Dundas et al., 2008; de Pablo and Komatsu, 2009).

2.3.3 Amazonian Climate

Landforms and surface textures found in Amazonian-aged landscapes include glacier-like landforms, mantling deposits, and gullies (as discussed in section 2.3.2). The distributions of these features are often attributed to variations of the martian climate affecting the locations where ice is deposited and, in a variety of cases, where melt has occurred. On Earth, climate change leading to ice ages is largely attributed to variations in the Earth’s orbital geometry (e.g. Hays et al., 1976; Berger, 1988). Mars has larger variations in its orbital parameters when compared to the Earth, largely due to a lack of stabilisation of obliquity – which for the Earth results from the presence of the Moon (Lasker et al., 1993). As such, it is important to understand how martian orbital parameters (e.g. location of perihelion, obliquity and eccentricity) have varied over time, and how these variations lead to changes in the locations where ice is deposited and melting occurs.

Changes in the eccentricity and obliquity of Mars result from gravitational perturbations as a result of the sun and planets, and oscillates on a time scale of \( \sim 10^5 \) years (Ward, 1973, 1974; Ward and Rudy, 1991). Present day obliquity is \( \sim 25.1^\circ \) but has been shown using orbital models as varying between values as significantly different as 11 -- 82° (Figure 2.15; Ward, 1973, 1974; Touma and Wisdom, 1993). Eccentricity values have been modelled to have been between 0.005 -- 0.141 (Ward, 1974). Averaged over 5 Ga Mars has
an eccentricity of 0.0690 (standard deviation: 0.0299) and obliquity of 37.62° (standard
deviation: 13.82°), with a probability of 89.3% that Mars’ obliquity has been greater than
60° in the past 3 Ga Laskar et al., 2004). Changes in these values result in significant
differences in solar insolation experienced on Mars, and thus the locations where ice is
deposited and melt occurs (Murray and Ward, 1973; Ward, 1973, 1974; Laskar et al., 2004).

During periods of high obliquity (>30°), the martian poles experience increased
insolation during summer, resulting in a greater release of water vapour (Head et al., 2003;
Forget et al., 2006) and higher mean annual temperature (Mellon and Jakosky, 1995). Due
to the increased insolation at the poles, less solar energy is delivered to the midlatitude
and equatorial regions resulting in lower annual mean temperatures there (Mellon and
Jakosky, 1995). Furthermore, the resulting higher humidity from the increase in water
vapour in the atmosphere leads to an increase in the stability of subsurface ice closer to
the equator and closer to the surface (Head et al., 2003). Models therefore predict that ice
from the poles is deposited onto the surface at lower latitudes at high obliquity (e.g. Mellon
and Jakosky, 1995). This transport of water is predicted to occur quickly and leads to
deposits of ice-rich material up to metres thick (Head et al., 2003). Such periods of high
obliquity represent martian “ice ages” and are estimated to occur on a timescale of ~40 kyr
(Figure 2.16, Head et al., 2003). Global circulation model (GCM) simulations by Haberle et
al. (2003) indicate that dust lifting potential also increases with increasing obliquity. Thus,
dust could act as nuclei for the condensation of water leading to the dust and ice being
rapidly removed from the atmosphere to the surface (Jakosky and Carr, 1985; Madeleine

![Figure 2.15: Variations in Mars' orbital obliquity over the past 10 Ma. Data taken from Laskar et al. (2004).](image)
et al., 2009). This leads to the formation of young blanketing mantle deposits that are ice-rich with incorporated dust (Head et al., 2003; Schon et al., 2012; Dickson et al., 2015).

![Figure 2.16: Taken from Figure 4, Head et al. (2003). Graphs show orbital forcing of the martian climate. **A**) Obliquity variations for the past 3 Ma with glacial (dark grey) and interglacial (light grey) periods marked. Small amplitude dotted line between 22 and 24° represents the obliquity on Earth during the same period of history. **B**) Maximal insolation of the north and south poles during the past 1 Ma. Before 300 kyr ago insolation was primarily driven by obliquity and is symmetric. As obliquity decreases, surface ice stability shrinks back toward the polar cap and humidity decreases, leading to water-ice remaining sequestered in the polar regions (Head et al., 2003). Ice sublimes from the upper layer of deposited ice, with calculations...](image-url)
suggesting poleward shrinking and desiccation of the ice-rich layer between 30 and 60° leaving behind a porous surface lag that protects against further sublimation (Mellon and Jakosky, 1995; Kreslavsky and Head, 2002a; Head et al., 2003; Kadish et al., 2010). Thus, Mars experiences cycles of net deposition and net desiccation (Figure 2.16; Kreslavsky and Head, 2002a; Head et al., 2003; Kadish et al., 2010; Dickson et al., 2015).

2.4 Summary

This background chapter provides important context for the work presented within this thesis and highlights what is already known about Lyot crater. Much of the previous research has been focused on both the small channels in the crater interior, and the large channels to the north and east of the crater. Using this as a starting point, the previous research was critically assessed and new research relating to the history of water in the Lyot crater study site was conducted. Through cultivating a better understanding of the history of water at this location it can be understood if the crater contains both recent atmospherically-derived water/ice and more ancient subsurface water/ice, which is indicated by previous research. This has important implications for both climate and habitability during the Amazonian on Mars, and whether material of astrobiological interest might be present at this location.
Chapter 3: Data and Methodology

3.1 Introduction

Within this chapter I detail the choice of methods used in this thesis, which led to the collection and interrogation of data and thereby a greater understanding of the history of water within and around Lyot crater. This chapter presents the methodology rather than the methods used throughout the thesis as a wide variety of techniques, software and analysis methods have been employed, such that it was more prudent to include the methods associated with each chapter of work within the appropriate thesis chapters. The first part of this chapter focusses on summarising the various key datasets used, as though techniques applied vary, the core datasets are the same.

3.2 Data

Current exploration of the planet Mars is limited to remote sensing. Before manned missions to Mars can occur infrastructure must be put in place and the various risks need to be mitigated and planned for (e.g. Huntsberger et al., 2000; Tavana, 2004). Until the advent of manned missions to Mars, the majority of the data about the planet is collected through the use of orbiters, rovers and landers. As at August 2018, there are six functioning satellites in orbit around Mars, two rovers collecting data on the surface, and a lander and two CubeSat’s en route (Table 3.1).
<table>
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<td>NASA</td>
<td>07/07/2003</td>
<td>25/01/2004</td>
<td>Rover</td>
</tr>
<tr>
<td><strong>Mars Reconnaissance Orbiter</strong></td>
<td>NASA</td>
<td>12/08/2005</td>
<td>10/03/2006</td>
<td>Orbiter</td>
</tr>
<tr>
<td><strong>MSL Curiosity</strong></td>
<td>NASA</td>
<td>26/11/2011</td>
<td>06/08/2012</td>
<td>Rover</td>
</tr>
<tr>
<td><strong>Mars Orbiter Mission</strong></td>
<td>ISRO</td>
<td>05/11/2013</td>
<td>24/09/2014</td>
<td>Orbiter</td>
</tr>
<tr>
<td><strong>MAVEN</strong></td>
<td>NASA</td>
<td>18/11/2013</td>
<td>22/09/2014</td>
<td>Orbiter</td>
</tr>
<tr>
<td><strong>ExoMars Trace Gas Orbiter</strong></td>
<td>ESA/Roscomos</td>
<td>14/03/2016</td>
<td>19/10/2016</td>
<td>Orbiter</td>
</tr>
<tr>
<td><strong>InSight</strong></td>
<td>NASA</td>
<td>05/05/2018</td>
<td>N/A</td>
<td>Lander</td>
</tr>
<tr>
<td><strong>Mars Cube One</strong></td>
<td>NASA</td>
<td>05/05/2018</td>
<td>N/A</td>
<td>2x CubeSat</td>
</tr>
</tbody>
</table>

**Table 3.1**: Summary of all operational missions (including missions that are en route) as at August 2018.

This thesis is primarily focussed on the geomorphological investigation of a number of landforms and surface types. Geomorphology is concerned with the evolution and origin of landforms and landscapes; seeking to understand how physical, chemical and biological processes have shaped the surfaces that are being studied. When conducting a geomorphological study it is therefore crucial to have imaging data sufficient to resolve the landforms and landscapes of interest – the higher the resolution, the greater the morphological information that can be drawn out from it. Within this study a combination of thermal, visible, topographic and spectral data have been used, which are summarised in Table 3.2, and will be discussed in the following sections.
<table>
<thead>
<tr>
<th>Orbiter</th>
<th>Instrument</th>
<th>Data Type</th>
<th>Dataset/Image Type Used</th>
<th>Resolution</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mars Global Surveyor</td>
<td>Mars Orbiter Laser Altimeter (MOLA)</td>
<td>Topographic</td>
<td>MOLA points</td>
<td>~300 m along track</td>
<td>PDS</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>MOLA (gridded elevation)</td>
<td>~168 m spot size</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>~128 pixels/degree</td>
<td>USGS</td>
</tr>
<tr>
<td>2001 Mars Odyssey</td>
<td>Thermal Emission Imaging System (THEMIS)</td>
<td>Thermal</td>
<td>THEMIS-IR Day Global Mosaic</td>
<td>100 m/pixel</td>
<td>USGS</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>THEMIS-IR Night Global Mosaic</td>
<td>100 m/pixel</td>
<td>USGS</td>
</tr>
<tr>
<td>Mars Express</td>
<td>High Resolution Stereo Camera (HRSC)</td>
<td>Visible</td>
<td>Nadir</td>
<td>~12.5 – 20 m/pixel</td>
<td>FUB</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Topographic</td>
<td>Up to 50 m/pixel</td>
<td>FUB</td>
</tr>
<tr>
<td>Mars Reconnaissance Orbiter</td>
<td>High Resolution Imaging Science Experiment (HiRISE)</td>
<td>Visible</td>
<td>RED</td>
<td>0.25 – 1.3 m/pixel</td>
<td>ASU²</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Visible</td>
<td>Anaglyph</td>
<td>ASU²</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Topographic</td>
<td>Digital Terrain Model (DTM)</td>
<td>ASU²/l made</td>
</tr>
<tr>
<td></td>
<td>Context Camera (CTX)</td>
<td>Visible</td>
<td>Panchromatic</td>
<td>5 – 6.5 m/pixel</td>
<td>ASU³</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Topographic</td>
<td>Digital Terrain Model (DTM)</td>
<td>ASU³/l made</td>
</tr>
<tr>
<td>Compact Reconnaissance Imaging Spectrometer (CRISM)</td>
<td>Spectral</td>
<td>IR Derived Product (Mafic Mineralogy)</td>
<td>15 – 19 m/pixel</td>
<td>JHUAPL</td>
<td></td>
</tr>
<tr>
<td>---------------------------------------------------</td>
<td>---------</td>
<td>-------------------------------------</td>
<td>-----------------</td>
<td>--------</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.2: Summary of the various datasets used within this thesis. Sources:

- Planetary Data Service (PDS) - [http://ode.rsl.wustl.edu/mars/dataPointSearch.aspx](http://ode.rsl.wustl.edu/mars/dataPointSearch.aspx),
- United States Geological Survey (USGS) - [https://astrogeology.usgs.gov/search](https://astrogeology.usgs.gov/search),
- Freie Universität Berlin (FUB) - [http://hrscview.fu-berlin.de/](http://hrscview.fu-berlin.de/),
- Arizona State University (ASU) - ASU¹: [http://viewer.mars.asu.edu/viewer/hiriseT=0](http://viewer.mars.asu.edu/viewer/hiriseT=0), ASU²: [http://viewer.mars.asu.edu/viewer/anaglyphT=0](http://viewer.mars.asu.edu/viewer/anaglyphT=0), ASU³: [http://viewer.mars.asu.edu/viewer/cbxT=0](http://viewer.mars.asu.edu/viewer/cbxT=0)
- I made – Indicates that DTM’s were generated by me, as opposed to being readily available in this format.
3.2.1 Thermal Data

Data products from the THEMIS camera (Christensen et al., 2004) are used predominantly as a tool for the distinction of the outer ejecta blanket around Lyot crater, and partially as a base layer for the geomorphological map created in Chapter 5. THEMIS uses thermal infrared (IR) multispectral imaging between 6.5 – 15 μm and visible to near-IR images from 450 – 850 nm (Christensen et al., 2003, 2004). Within this thesis, the global day- and night- time mosaics have been used (Table 3.2; Edwards et al., 2011). The daytime mosaic uses data from -60°N to 60°N and solar incidence angles of <85°, ensuring the sun was a minimum of 5° above the horizon, with only full-resolution data used (Edwards et al., 2011). The nighttime mosaic uses the same data constraints with one variation – a solar incidence angle of >95° was used, ensuring the sun was a minimum of 5° below the horizon (Edwards et al., 2011).

THEMIS data enable the spatial visualisation of differences in thermal properties of the martian surface, which are related to properties such as composition and topography (Christensen et al., 2003). The daytime mosaic displays the effects of surface emissivity, reflectance, thermal properties and slope; whereas the nighttime mosaic predominantly displays the effects of inherent thermal properties (Christensen et al., 2003). Contrasts between the two mosaics indicate differences in thermal inertia (Christensen et al., 2003) i.e. materials with a high thermal inertia will retain heat for longer and therefore appear brighter in the nighttime mosaic.

Within the THEMIS daytime mosaic the outer ejecta blanket is visible as a bright unit with a darker “halo” (Figure 3.1). This indicates that the ejecta material reflects electromagnetic radiation of the wavelength used by THEMIS. In particular, materials that are lighter or redder appear bright in this image. As a result of this apparent “brightness” the outer ejecta extent has been demarcated.
3.2.2 Visible Data

Visible wavelength imaging data has been of crucial importance for the work conducted within this thesis. It has been used to conduct morphometric measurements, map out morphological units and identify key features of interest to this body of work. Three sources have been used: HRSC (Neukum et al., 2004), CTX (Malin et al., 2007), and HiRISE (McEwen et al., 2007). These data are summarised in Table 3.2 and detailed in the following sections. Topographic data constructed using visible data has been included within this section.

3.2.2.1 High Resolution Stereo Camera (HRSC)

HRSC is a multi-sensor push-broom scanning instrument with nine CCD line detectors mounted in parallel in the focal plane; and consists of the HRSC colour scanner, Super-Resolution Channel (SRC) and digital unit (Neukum et al., 2004; Jaumann et al., 2007). HRSC is able to obtain triple to quintuple panchromatic stereo images with a vertical resolution of up to one pixel (Jaumann et al., 2007). Data used within this thesis are images from the panchromatic nadir channel (ND), and ready-constructed DTMs from HRSC stereo images (Gwinner et al., 2009; summarised in Table 3.2).
HRSC DTMs are the highest resolution topographic data available to provide complete topographic profiles of the channel landforms present within the interior of Lyot crater with a resolution of up to 50 m/pixel (Gwinner et al., 2009). These DTMs were used to create topographic profiles of a number of fluvial channels mapped across the interior (Chapter 6). A drawback of these data are that area matching employed in the creation of the DTMs results in noise as a result of low contrast of the images leading to poor matching (Albertz et al., 2005).

Nadir image products were used to fill in gaps within a CTX mosaic generated for the study area (Chapter 4). These products are not of sufficient resolution (~12.5 – 20 m/pixel) to be used for grid mapping (Chapter 4) or geomorphological mapping (Chapter 5). This mapping often requires the resolving of surface textures or small features (e.g. clastic polygons, high-centred polygons and boulders). Furthermore, higher resolution CTX data are available, which have a good level of coverage (near 100%), so it is unnecessary to rely on HRSC data.

### 3.2.2.2 Context Camera (CTX)

CTX obtains panchromatic images with a higher pixel resolution of 6 m/pixel (Malin et al., 2007; Table 3.2), when compared to HRSC. It provides context for targeted observations provided by HiRISE and CRISM, and is able to produce overlapping stereo images for use in the creation of DTMs (Malin et al., 2007). Data from this instrument are available for almost 100% of the study area and have been used extensively within this thesis in the form of a mosaic. CTX data are sufficient for the identification of surface texture and landforms of interest within this body of work (e.g. clastic polygons, boulder fields and rough texture). For instance, clastic polygonal features can be identified within CTX images, although smaller-scale features such as the demarcating clasts are not resolvable (Chapter 7). Therefore, a mosaic created from CTX data has been used as the primary basemap onto which the geomorphological and grid maps were created (Chapters 4 and 5). The only study for which CTX data were not suitable was the study of the clastic polygonal landforms in Chapter 7, for which it was necessary to have data which resolved the clasts that demarcate the polygon margins – HiRISE data was required. Some of the clastic polygon study sites did not have HiRISE stereo images available, so in these cases CTX DTMs were used. These DTMs were generated using the method of Kirk et al. (2008) which uses Integrated Software for Images and Spectrometers (ISIS) software to perform radiometric and instrument specific calibrations on the raw data, before processing within SOCET.
SET® (www.socetset.com), a commercially available photogrammetry suite. This process involves the identification of up to millions of matching points between the two images, and the calculation of the elevation for each matching point. As part of this process DTMs are controlled to MOLA track data within SOCET SET®. The generation of DTMs is a challenging and time-consuming process, with each DTM taking between two days and a week to make; it is for these reasons that the number of DTMs included in the thesis is limited.

3.2.2.3 High Resolution Imaging Science Experiment (HiRISE)

HiRISE data are the highest resolution data available across the study area, with a resolution of down to 0.25 m/pixel. Whilst there are a number of HiRISE images available across the study area (predominantly within the crater interior), they cover <1% of the mapping area, therefore coverage is not sufficient that HiRISE can be used extensively; rather these data are used as a higher resolution “check” of the surface textures and landforms observed in lower resolution CTX data. HiRISE data were predominantly used for the study of the clastic polygonal features, presented in Chapter 7. This is because CTX data are not sufficient to resolve the clastic margins of the polygons, and this is necessary to analyse the possible polygon formation mechanisms (as discussed in Chapter 7). Furthermore, a HiRISE DTM was generated for one polygon study site for which a stereopair was available using a similar method by which the CTX DTM’s were generated (Kirk et al., 2008). The generation of HiRISE DTMs involves extra pre-processing steps within ISIS to mosaic and colour balance 20 Experimental Data Records (EDR) for each HiRISE image, as such they take longer to generate. One readily available HiRISE DTM was also used, which corresponds with the location of a fan deposit studied in Chapter 6. This DTM allows important topographical information about the fan deposit to be distinguished.

Another HiRISE-derived product used within this thesis is the stereo colour-anaglyph systematically produced by the HiRISE team for stereo acquisitions. This product gives the perspective of a 3D view when viewed through glasses with the appropriate filters. This product is used to provide a 3D view of the clastic polygonal features to demonstrate that these polygons have topographically high margins (Chapter 7).

3.2.3 Topographic Data

The highest resolution topographic data used within this study are derived from digital terrain models created using HiRISE and CTX stereopairs (detailed in section 3.22). The other topographic data products that are used extensively are obtained from MOLA (Zuber
et al., 1992). MOLA measured the round-trip time of flight of infrared laser pulses which were transmitted from the Mars Global Surveyor orbiter to the martian surface (Zuber et al., 1992). This value can then be used to calculate the distance from the orbiter to the point measured on the surface, and this can be used to calculate the height of the surface topography with respect to the radius of the aeroid, which is determined from measurements of the martian gravitational field (Smith et al., 2001). Therefore, zero elevation from MOLA represents the equipotential surface (Smith et al., 2001). MOLA data are available as point elevations and as a gridded raster (Smith et al., 2001). These data are used to calibrate all other elevation datasets used in this thesis.

3.2.3.1 MOLA point data

There are ~4 x 10^6 MOLA points which cover the entirety of the Lyot crater study area. These points have an along-track spacing of ~300 metres (Table 3.2), where each point has a surface spot size of ~168 metres (Smith et al., 2001). These data have a vertical accuracy of ~1 metre (Smith et al., 2001). The orientation of the points is determined by the orbit of the craft with which the measurements were conducted – the Mars Global Surveyor (Smith et al., 2001). MOLA point data were primarily used to control the generated CTX and HiRISE DTMs within SOCET SET®. When the points were suitably aligned they were used to check the elevations measured for channels within Chapter 6.

3.2.3.2 MOLA gridded data

Predominantly, when elevation data were required, the MOLA gridded raster product was used (Smith et al., 2003). This product is created by interpolating the MOLA point data at a variety of resolutions; in this case the highest resolution of 128 points/degree was used (Som et al., 2008). The fact that these data are interpolated means that some grid cells are devoid of shots and therefore do not contain any “real” information (Som et al., 2008). This is an important caveat to be aware of when using these data products. They are however still a valuable source by which elevation values across Mars can be extracted, or at least estimated.

Within this thesis MOLA gridded data have been used in a variety of different ways. MOLA gridded data was used in the identification of Lyot crater’s inner ejecta scarp, crater rim, inner peak-ring and central peak. These data were also used to analyse slopes on a variety of scales including the creation of: contour maps for the entirety of the Lyot study area, topographic profiles across Lyot crater, topographic profiles of fluvial channels, and,
“flood” maps analysing the possible locations in which water could pool. These products were created using ArcGIS 10.1.

3.2.4 Data Manipulation

Data within this thesis were manipulated within a Geographic Information System (GIS). A GIS is a software framework in which various spatial data can be stored, displayed and analysed. It is essentially a powerful system by which multiple datasets are related by a particular spatial projection and can be visualised simultaneously, usually as a map. Within this thesis the GIS used is ArcGIS 10.1, which is a commercial software package developed by ESRI.

Two important considerations when using a GIS are the geographic coordinate system and the choice of projection coordinate system, as this will affect the way in which the data are displayed and whether measurements taken within the GIS are accurate. The geographic coordinate system is defined by a datum and a prime meridian and represents a “real-world” location defined by coordinates, typically on a sphere or an ellipse. A projected coordinate system defines how a 3D area is represented in 2D, and therefore has constant lengths, angles and areas. The projected coordinate system that is used has a profound effect on the accuracy of measurements taken using the GIS.

Within this thesis the Mars 2000 spherical datum has been used for the geographical coordinate system, which defines Mars as a sphere with a radius of 3396.19 km. The projected coordinate system that has been chosen is a Cassini projection centred on 30° (the centre of Lyot crater). The Cassini projection can be imagined as a transverse cylinder projected onto a globe that is tangent along the central meridian. Lyot crater is located high up in the midlatitudes of Mars, so a Cassini projection prevents “flattening” of the crater in the vertical direction. This projection has a line of zero distortion along the central meridian; however, north-south linear distortion increases eastwards and westwards from the central meridian (Maling, 1992). The Lyot study area is not wide enough for significant distortion to occur that could impact on the measurements taken.

3.3 Methodology

3.3.1 Key Landforms and Grid Mapping

Based upon the first research question (Chapter 1.2.1), it’s important to distinguish where the key landforms and landscapes indicative of water are located. This enables both recognition of the various landforms that could be used to analyse the history of water,
and study of the distribution of these landforms which could indicate where this water might have originated. As the focus of this work is the history of the water associated with Lyot crater, rather than the distribution of this water, this stage is predominantly a reconnaissance stage.

I decided that the applied method would need to quickly identify and map a number of landforms and surface types over a relatively large area (~600 km²). For these reasons the grid mapping method, as described by Ramsdale et al. (2017) and within Chapter 4, was chosen. This method is used to identify where a selected group of landforms and surface types are located quickly and efficiently over a large area, and outputs a useful visualisation of specific patterns in their distribution. The method employs a gridded “tick box” approach to record the presence or absence of particular landforms in a chosen mapping area (Ramsdale et al., 2017). Thus, a coarse resolution map displaying the distributions was created to identify key landforms. This preliminary map was then used to assess the distributions of the landforms and surface types, helping to provide a basis for further work.

Grid mapping identifies the landforms and surface types related to water and/or ice and where they are located. The distributions provide evidence as to how this water might have originated. One of the primary goals of this thesis is understanding if Lyot crater could preserve evidence of both ancient subsurface and more recent atmospherically-sourced water. The reconnaissance grid mapping enabled me to identify two avenues of future work which potentially provide evidence of such atmospheric and subsurface water. These avenues of work consist of a concentrated study of fluvial landforms within the interior of Lyot crater (Chapters 5 and 6); and, an in-depth study of unique clastic polygonal features within the outer ejecta blanket around Lyot crater (Chapters 7 and 8).

3.3.2 Investigating Landforms

3.3.2.1 Geomorphological Map

Grid mapping in Chapter 4 identified fluvial landforms within the interior and outer ejecta of Lyot crater that are morphologically similar to river channels with associated depositional fans. These features are predominantly located within the interior of the crater, and previous work has identified that the source of melt was either an ice sheet (Hobley et al., 2014) or glaciers (Dickson et al., 2009) which were likely atmospherically-derived (Forget et al., 2006; Head et al., 2006a), with the source of heat for melt related to
shifts in the martian climate (Kreslavsky and Head, 2002a; Head et al., 2003; Kadish et al., 2010; Dickson et al., 2015).

An important stage in the identification and distinction of the fluvial landforms is the development of a better understanding of the modification history of the crater. Through mapping the various morphological units, their stratigraphic relationships and associations, and how they have been modified, a history of the crater can be constructed which provides context for the fluvial activity that formed the channels and fans. It is for these reasons that a more detailed mapping project was undertaken. This map focussed on the interior of Lyot crater where the majority of the fluvial landforms are located and is presented in Chapter 5. It was constructed using the highest resolution data available for the entirety of the crater interior which is CTX data. Prior work by Dickson et al. (2009) and Hobley et al. (2014) did not involve the creation of a geomorphological map; thus, the map was used as context for both the collection of further data and to review the current hypotheses for their formation.

3.3.2.2 Analysis of landforms potentially indicative of atmospheric or subsurface water/ice

Prior mapping identified landforms which could provide important information about the history of water in and around Lyot crater. These include: the channels and fan deposits in the crater interior, and clastic polygonal landforms in the outer ejecta blanket. To analyse these features the same methodology was applied and is presented below.

First, the landforms are mapped in planview, and observations are noted of the landforms, and the relationship these features have with morphological units. In particular, channels and fan deposits were assessed to see if they could have been sourced from ice-rich deposits. The next stage of analysis involved detailed morphometry of the landforms which was used to identify different formation processes based upon analysis of competing hypotheses within the literature and comparison to terrestrial analogues. These data and observations were used to establish the closest possible morphological terrestrial analogue and then propose possible formation mechanisms based on these analogies.

3.3.3 Impact modelling

To build on work conducted on the clastic polygonal features (Chapter 7), 2D- and 3D-modelling of the impact event that formed Lyot crater was conducted using impact-SALE-Dellen (iSALE-Dellen) – a multi-material, multi-rheology shock physics code (Amsden et al, 1980; Collins et al., 2004; Wünnemann et al., 2006; Elbeshausen et al., 2009; Elbeshausen and Wünnemann, 2011; Collins et al., 2016). This work focusses on identifying the depth
from which the material within the ejecta blanket originated, to test whether it could be related to a zone of high ice content within the impact target.

The depth from which material within the ejecta blanket was exhumed can be estimated by examining the output of the 2D- and 3D-models. These values were either extracted directly from the model, by taking the start and end coordinates of tracer particles placed within the projectile and target to work out where they originated and where they were located at the end of the model run; or calculated manually using an initial ejection velocity extracted from the model – this prevented the model space from needing to be significantly large as to capture material deposited up to ~350 km away from the crater rim. The exact methods employed are discussed in more detail within Chapter 8.

Using the models, the depth from which the material in which the polygons have formed was exhumed can be estimated. These values were then compared to theoretical depths of the cryosphere and crustal composition in this region, at the time of the impact, to determine whether this exhumed material could have had a particularly high ice content. Furthermore, these models were used to further test if Lyot crater could have penetrated the cryosphere and thus released groundwater onto the martian surface.

3.4 Conclusions

To summarise, the majority of the research was conducted using topographic and visible data, with thermal data mostly used as a basemap. To answer the research questions presented in Chapter 1, landforms and surface types were identified and mapped across the Lyot crater study area. An in-depth morphological map was created of the crater interior which provides context for the fluvial landforms found in this location. Landforms of particular interest were selected for further study including: fluvial channels and fans, and clastic polygonal networks. These features were analysed via qualitative observations and quantitative measurements, which are used to test hypotheses taken from the literature and compare these observations and results to terrestrial analogues. Impact modelling was then conducted to analyse the depth from which material deposited in the ejecta blanket was exhumed. These results were then compared to theoretical depths of the cryosphere, and analysis is conducted as to whether water could have been released during or immediately after the impact event.
Chapter 4: Preliminary Mapping

4.1 Introduction

Grid mapping is an approach by which specific landforms of interest can be identified across a study area quickly and efficiently (Ramsdale et al., 2017; Voelker et al. 2017). To apply this method, a number of different landforms and surface types are first identified through observation of the mapping region. In a Geographical Information System (GIS) computing package, the site is then split into a “shapefile” grid of suitable size taking into consideration the resolution that the mapping shall be conducted at. An attribute table is then produced for this “shapefile” (e.g. Table 4.1), with a column for each landform and surface type (Ramsdale et al., 2017). Finally, each grid square is then examined in the appropriate imaging dataset to identify which of the chosen landforms/surface types are present and this information is then stored in the attribute table (Ramsdale et al., 2017). The result of this style of mapping is a series of “raster” datasets that display the distribution of the different landforms/surface types across the study area (Ramsdale et al., 2017, Voelker et al., 2017).

Table 4.1: Screen-grab from ArcGIS 10.1 of part of the attribute table storing grid mapping data from Lyot crater.

This method has been applied to Lyot crater to create a coarse resolution map showing where landforms and surface types of interest are present. Thus, any patterns within the results are observed, and a foundation for further work was been created.
Detailed below is the method applied to create the coarse resolution map, followed by the results of this reconnaissance mapping.

4.2 Method

4.2.1 Base Map

The first stage of grid mapping involved the production of a suitable base map. This map needed to be composed of images that have a suitable resolution to identify landforms and surface types of interest to the investigation. The coverage that the images provide also needed to be considered, so that as much of the Lyot crater study area as possible could be mapped. In this case, 6m/pixel context camera (CTX) images (Malin et al., 2007) were chosen as they have suitable resolution for the identification of all the features to be studied. The coverage across Lyot crater is also near 100%, and all the images can be displayed at once using ArcGIS 10.1 software. CTX images were downloaded from the Arizona State University Mars portal and manipulated within ArcMap 10.1 (Chapter 3).

To ensure the best possible base map was used, a “mosaic” dataset of CTX images was created. The mosaic covers an area of ~2223 x 1450 km centred on Lyot crater (50°N, 30°E). This size of mosaic was chosen to ensure that the entirety of Lyot crater and its ejecta blanket could be examined. A Cassini projection centred on the 30° East meridian was used.

4.2.2 Landforms and Surface Types

To decide upon the landforms and surface types to be mapped, a combination of preliminary observations using the CTX mosaic and a literature review was used. During this process, the map scale was decided upon by looking at the scale at which the key landforms/surface types could be identified. In this case 1:20,000 was found to be most suitable. The suite of landforms and surface types to be identified within the grid mapping was combined into a reference document, with the chosen “type examples” shown at a scale of 1:20,000 (except for the type example for viscous flow features, as these features are too large to be fully viewed at this scale). Accompanying each image was a short description to aid in the identification of these features. Descriptions of the various landforms and surface types are presented below with the corresponding “type example” image. The landforms and surface types are also summarised in Table 4.2.
<table>
<thead>
<tr>
<th>Landform/Surface Type</th>
<th>Type Example</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Dunes</strong></td>
<td><img src="Dunes.png" alt="Image" /></td>
<td>Large-scale features found within the crater interior. Often recognised as distinctive dark areas associated with channel terminations.</td>
</tr>
<tr>
<td><strong>Viscous Flow Features</strong></td>
<td><img src="ViscousFlow.png" alt="Image" /></td>
<td>Features that indicate the slow, plastic deformation or movement of ice and debris down slope. They are convex-outwards in plan-view and convex-upwards in long- and cross-profile.</td>
</tr>
<tr>
<td><strong>Channels/Inverted Channels</strong></td>
<td><img src="Channels.png" alt="Image" /></td>
<td>Commonly found throughout the crater interior. They are generally small (up to hundreds of metres across and tens of kilometres in length), and simple (non-branching) in planform.</td>
</tr>
<tr>
<td><strong>Gullies</strong></td>
<td><img src="Gullies.png" alt="Image" /></td>
<td>Composed of a head alcove, one or more channels and depositional aprons. Some are sinuous, though others are straight in morphology. They are commonly located on steeper slopes.</td>
</tr>
<tr>
<td><strong>Low-Centred Polygons</strong></td>
<td>Polygonal features that have depressed interiors and troughs with raised shoulders. Generally located to the north to south-east region of the outer ejecta.</td>
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<tr>
<td><strong>High-Centred Polygons</strong></td>
<td>Polygonal features that have depressed troughs with high interiors relative to their boundaries. No discrimination was made between kilometre-scale polygons and those less than 100 metres in diameter.</td>
<td></td>
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<tr>
<td><strong>Clastic Polygons</strong></td>
<td>Polygonal features that are low-centred with clastic margins. They are commonly located in the north to south-east region of the outer ejecta blanket.</td>
<td></td>
</tr>
<tr>
<td><strong>Boulder Fields</strong></td>
<td>Rough-textured, low albedo features containing boulders too small to resolve at CTX level. They cover large expanses at the margin of the outer ejecta blanket.</td>
<td></td>
</tr>
<tr>
<td><strong>Mantling Deposits</strong></td>
<td>Material that drapes the topography thereby smoothing it. Located across the study area.</td>
<td></td>
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<tr>
<td><strong>Relief Filling Deposits</strong></td>
<td>Material that infills topographic lows. Commonly found where the terrain undulates.</td>
<td></td>
</tr>
<tr>
<td><strong>Multiple Layers</strong></td>
<td>Material that displays layering, commonly found within the outer ejecta blanket.</td>
<td></td>
</tr>
<tr>
<td><strong>Texture</strong></td>
<td>Surface textures often associated with mantling deposits includes: “stippled” terrain, basketball terrain, linear terrain, brain terrain and wrinkled terrain.</td>
<td></td>
</tr>
<tr>
<td>Feature</td>
<td>Description</td>
<td></td>
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<tr>
<td>-------------------------</td>
<td>----------------------------------------------------------------------------</td>
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</tr>
<tr>
<td>Pits</td>
<td>Depressions in the ground with simple, non-scalloped edges. Common throughout the study area.</td>
<td></td>
</tr>
<tr>
<td>Scallops</td>
<td>Flat-floored depressions with scalloped edges. Common throughout the study area.</td>
<td></td>
</tr>
<tr>
<td>Isolated Mounds</td>
<td>High-relief dome features, they can be found with superposing mantling deposits. Common throughout the study area.</td>
<td></td>
</tr>
<tr>
<td>Hummocky Terrain</td>
<td>Terrain displaying multiple mounded features or “hummocks”. Generally found in undulating terrain within the outer ejecta.</td>
<td></td>
</tr>
<tr>
<td>Flat Pitted Ground</td>
<td>Heavily pitted, relatively dark surface type. Channels are commonly incised into it.</td>
<td></td>
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<td>--------------------------------------------------------</td>
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</tr>
<tr>
<td>Impact Structures</td>
<td>Any clearly identifiable impact crater found across the study site.</td>
<td></td>
</tr>
</tbody>
</table>

**Table 4.2**: Summary of the landforms and surface types included within the grid mapping.
4.2.2.1 Dunes

Dune fields (Figure 4.1) are generally large-scale features found within the crater interior, though smaller isolated dunes have been observed. They are easily recognised as distinctive dark areas and they are often associated with channel terminations (Balme et al., 2013a). Dunes can also cover features of interest so it is useful to know where they are located. The formation of dunes occurs as a result of the action of sand-moving winds blowing across the surface of Mars (Cutts and Smith, 1973; Parteli and Herrmann, 2007).

![Figure 4.1: CTX mosaic image showing the type example for dunes.](image)

4.2.2.2 Viscous Flow Features

Viscous flow features (VFFs, Figure 4.2) are recognised as features that indicate the slow, plastic deformation or movement of ice and debris down slope (Squyres, 1978; Squyres, 1979; Milliken et al., 2003; Arfstrom and Hartmann, 2005; Head et al., 2005). Characteristics of VFFs include evidence of flow around or over obstacles (Milliken et al., 2003), surface lineations (Squyres, 1978; Mangold, 2003; Milliken et al., 2003, Head et al., 2006a), compressional ridges or extensional troughs (Milliken et al., 2003; Arfstrom and Hartmann, 2005; Head et al., 2005), a lobate morphology (Squyres, 1978; Mangold, 2003; Milliken et al., 2003; Head et al., 2005) and moraine like ridges (Arfstrom and Hartmann, 2005). VFFs are split into a number of sub types including lobate debris aprons (LDAs; Squyres, 1978; Squyres, 1979; Mangold, 2003; Milliken et al., 2003; Berman et al., 2009; Souness and Hubbard, 2012), lineated valley fill (LVF; Squyres, 1979; Mangold, 2003;
Milliken et al., 2003; Berman et al., 2009; Souness and Hubbbard, 2012), concentric crater fill (Milliken et al., 2003; Berman et al., 2009; Levy et al., 2009a) and glacier like forms (GLFs; Arfstrom and Hartmann, 2005; Head et al., 2006a; Madeleine et al., 2009; Souness et al., 2012; Souness and Hubbard, 2012).

VFFs are of interest as a strong indicator of ice in the ground (Holt et al., 2008; Plaut et al., 2009; Hubbard et al., 2014). Previous work by Dickson et al., (2009) has also indicated that fluvial channels within Lyot crater could originate from these features. In Lyot crater, VFFs are located along the along the inner peak ring and the southern crater rim (Dickson et al., 2009). They are convex-outwards in plan-view and convex-upwards in long and cross-profile and often have an irregular pitted surface with occasional polygonal fractures, perhaps indicating the sublimation of ice (Head et al., 2005; Levy et al., 2006).

4.2.2.3 Channels and Inverted Channels (small)

Channels (Figure 4.3) are commonly found throughout the interior of the crater and within the inner ejecta blanket (Hobley et al., 2014). Most originate within the crater rim or central peak ring (Balme et al., 2013a). They are commonly small (up to hundreds of metres across and up to tens of kilometres in length), and simple (non-branching) in planform, although some contributory networks occur (Balme et al., 2013a). A small number of ridges associated with these channels are observed within Lyot Crater (Hobley et al., 2014), that could potentially be inverted channels.
Channels and inverted channels are of interest as they are interpreted as fluvial in origin due to their relatively low slopes and sinuous nature (Dickson et al., 2009). Thus, they record part of the history of water in Lyot crater.

**Figure 4.3**: CTX mosaic image showing the type example of a channel. The downslope direction is from the bottom to the top of the image.
4.2.2.4 Gullies

Gullies on Mars are commonly found on the walls of craters and can be found on any other type of hillslope (Malin and Edgett, 2000; Malin et al., 2006; Balme et al., 2006; Dickson et al., 2007; Berman et al., 2009). They are composed of a head alcove, one or more channels and depositional aprons (Malin and Edgett, 2000). Some channels are sinuous, but others are straight in morphology (Malin et al., 2006).

Gullies are of interest as they might be the result of the action of fluids, such as liquid water (e.g. Head et al., 2008, Conway et al., 2011, de Haas et al., 2015). In particular, it has been suggested that the source of such a fluid might be ice deposits related to glaciation (Milliken et al., 2003; Arfstrom and Hartmann, 2005; Head et al., 2008). Thus, it is useful to know where gullies are found across the study area. It should be noted that the process by which gullies form on Mars, and whether it is related to liquid water is uncertain. Other research indicates that gullies might form through dry slope processes (e.g. Treiman, 2003; Pelletier et al., 2008), or the action of CO$_2$ (e.g. Hoffman, 2002; Cedillo-Flores et al., 2011; Dundas et al. 2017).

![Figure 4.4: CTX mosaic image showing the type example of a gully. The downslope direction is from the bottom to the top of the image.](image)

4.2.2.5 Low-Centred Polygons

Low-centred polygons (Figure 4.5) have depressed interiors and troughs with raised shoulders (Levy et al., 2009b). Within Lyot crater, they are generally located to the north
and south east regions of the outer ejecta and are visible on areas of higher topography (Balme et al., 2013a).

Such polygons could be analogous to a variety of different polygon types including ice-wedge polygons (Levy et al., 2009b, 2010; Ulrich et al., 2011), sand-wedge polygons (Levy et al., 2009b; Levy et al., 2010; Ulrich et al., 2011) and sublimation polygons (Marchant and Head, 2007; Levy et al., 2006, 2010). They are often considered to be indicative of periglacial processes and thus are of interest as they likely indicate the freezing and thawing of ice (Soare et al. 2014; Mangold, 2005; Levy et al., 2009b).

**Figure 4.5:** CTX mosaic image showing the type example for low-centred polygons.

4.2.2.6 High-Centred Polygons

High-centred polygons (Figure 4.6) have depressed troughs with high interiors relative to their boundaries, they are commonly slightly convex upwards (Levy et al., 2009b). No discrimination in the grid mapping was made between kilometre-scale polygons, found commonly within the interior of Lyot crater, and polygons less than 100 metres in diameter. Kilometre-scale polygons were mapped as a unit for a future geological map (Chapter 5).

High-centred polygons might form in a variety of ways, for instance as sublimation polygons with preferential sublimation of ice at polygon troughs (Levy et al., 2008a, 2009), as unstable, degrading ice-wedge polygons (Levy et al., 2010), or via desiccation (El Maarry
et al., 2010, 2012, 2014). As in the previous case, high-centred polygons could indicate the freezing and thawing of ice.

4.2.2.7 Clastic Polygons

Clastic polygonal networks (Figure 4.7) are common within the Lyot study area (Balme et al., 2013a). They are limited to the north and south east regions of the outer ejecta and appear to occur on areas of higher topography (Balme et al., 2013a). They are enigmatic due to the presence of clasts which appear to demarcate the edge of the polygons (Balme et al., 2013a). The polygons are generally 4 or 5 sided and 100 – 200 m in diameter with clasts ranging up to several meters in size (Balme et al., 2013a).

Due to the size and clastic nature of these polygons the mechanisms involved in their formation are not well understood. Clasts might be the result of freeze-thaw cryoturbation processes (Balme et al., 2009; Gallagher et al., 2011), frost creep (Gallagher et al., 2011) or boulder clustering due to a “ratcheting” process which is the result of the locking of boulders within a CO$_2$ frost layer (Orloff et al., 2013). The clasts might also be the result of gravitational slumping whereby boulders fall into thermal contraction cracks (Levy et al., 2010), or a previously unreported mechanism (Brooker et al., 2018; Chapter 7). Previous work has inferred that these features are associated with outer ejecta material (Balme et al., 2013a), which indicates a genetic relationship with this material. It is due to this
relationship and the enigmatic nature of these polygons that they were chosen as a feature to be mapped.

Figure 4.7: CTX mosaic image showing the type example for clastic polygons.
4.2.2.8 Boulder Fields

Boulder fields (Figure 4.8) cover large expanses of ejecta within the Lyot crater study area. They generally appear to be located at the edges of the outer ejecta blanket (Balme et al., 2013a). They are identifiable as rough textured, low albedo features (Balme et al., 2013a). Many boulders are too small to resolve at CTX level, but the texture can be easily associated with boulders by comparing CTX and higher resolution HiRISE images (up to 25 cm/pixel; McEwen et al., 2007).

Clasts such as boulders, cobbles, pebbles and finer sediments are common in Martian permafrost environments (Levy et al., 2011). Some of the boulder fields also display possible self-organised patterns (Balme et al., 2013a). These areas are reminiscent of sorted stone circles observed in Elysium Planitia, and at Lomonosov crater, which might be formed as a result of freeze-thaw processes (Balme et al., 2009; Barrett et al., 2017). There is also the potential that these features are related to the clastic polygonal networks. Thus, it is useful to know where they are located as they could provide information about both the process by which clastic networks form, as well as the locations of ice.

![Figure 4.8: CTX mosaic image showing the type example for boulder fields.](image)

4.2.2.9 Mantling deposit

A mantling deposit (Figure 4.9) is recognised by its draping of the topography, thereby smoothing it, and it is present across large parts of the martian mid- to high-latitudes (Mustard et al., 2001; Head et al., 2003). The material comprising the mantling deposit is
likely an air-fall deposit composed of a mixture of dust and ice (Mustard et al., 2001; Kreslavsky and Head, 2002a; Schon et al., 2012). Such material is likely the youngest in the study area (Pan et al., 2018). Head et al. (2003) dated a similar mantling deposit using crater counting and found it was “very youthful” at less than 10 Ma. Schon et al. (2012) and Willmes et al. (2012) found even younger ages of ~0.4-5 Ma. In general, mantling deposits are considered to be the youngest geological units, and are thought to have formed within the Late Amazonian (Morgenstern et al., 2007; Schon et al., 2012). The deposition of mantling deposits is commonly interpreted as resulting from differences in obliquity leading to changes in latitudinal insolation (Kostama et al., 2006; Schon et al., 2012; Dickson et al., 2015). Due to the links to climatic differences, and the possibility of water ice, it is useful to locate such mantling deposits.

Figure 4.9: CTX mosaic image showing the type example for mantling deposits.
4.2.2.10 Relief Filling Deposits

Relief filling deposits (Figure 4.10) are recognised by the infilling of topographic lows. Such deposits are likely to be another, but discontinuous, form of mantling deposit, where underlying rougher material of higher topography has thinner mantle coverage, or is visible through the deposit (Mustard et al., 2001; Kreslavsky and Head, 2002a). In the Lyot crater study area, it is often found near to polygonal networks and boulder fields where the terrain undulates.

![Figure 4.10: CTX mosaic image showing the type example for relief filling deposits.](image)

4.2.2.11 Multiple Layers

Multiple layers (Figure 4.11) are interpreted as areas where there are multiple generations of mantling deposit present (Kreslavsky and Head, 2002a; Milliken et al., 2003; Head et al., 2003; Hobley et al., 2014). Multiple layers are commonly found within the outer ejecta blanket.

Mantling deposits can often be observed as layered deposits covering older terrain (Kreslavsky and Head, 2002a; Milliken et al., 2003; Head et al., 2003; Levy et al., 2011). Older mantling deposits can become degraded over time due to the insolation-driven sublimation of ice (Head et al., 2003; Arfstrom and Hartmann, 2005; Dickson et al., 2015), as is likely the case in areas of Lyot crater (Dickson et al., 2009; Hobley et al., 2014). The layered form of the deposit suggests the deposition and removal of material as a result of multiple cycles (Kreslavsky and Head, 2002a; Head et al., 2003; Dickson et al., 2015).
Therefore, it could represent Martian climatic variations, likely resulting from changes in martian orbital parameters (Murray and Ward, 1973; Ward, 1973, 1974; Laskar et al., 2004). It is for this reason that multiple layers were considered important to locate, as it likely provides an insight into climate-controlled deposition of ice and dust within the area.

Figure 4.11: CTX mosaic image showing the type example for multiple layers.
4.2.2.12 Texture

Surface textures are often observed associated with mantling deposits within the study area (Figure 4.12). Textures included in this category are: basketball terrain (Kreslavsky and Head, 2002a; Head et al., 2003; Kostama et al., 2006), linear terrain (Head et al., 2003; Kostama et al., 2006), brain terrain (Levy et al., 2009b) and wrinkled terrain (Head et al. 2003; Kostama et al., 2006). There are also surfaces that appear “stippled” which might be the result of small polygonal networks that cannot be resolved within CTX images. These terrains are not often distinguishable from one another in CTX images, and so have been grouped together as “texture”. Such surface textures are often attributed to the presence of ice, thus potentially providing further evidence that mantling deposits are ice rich (Head et al., 2003; Mangold, 2005; Kostama, 2006; Levy et al., 2009a; Levy et al., 2009b; Hobley et al., 2014).

Figure 4.12: CTX mosaic image showing the type example for texture.
4.2.2.13 Pits

Pits (Figure 4.13) are identified as areas where depressions in the ground occur with simple, non-scalloped edges. They can be isolated, or located near to other pits, and are found relatively commonly throughout the study area. Pits are likely the result of the sublimation of ice (Head et al., 2005; Kadish et al., 2008), though some could potentially be softened impacts superposed on mantle deposits (Kostama et al., 2006).

![Figure 4.13: CTX mosaic image showing the type example for pits.](image)
4.2.2.14 Scallops

Scallops (Figure 4.14) are flat-floored depressions with scalloped edges (Séjourné et al., 2011). They have circular to elliptical shapes with diameters ranging from tens of meters to several kilometres, and depths extending to tens of meters (Costard and Kargel, 1995; Zanetti et al., 2010; Séjourné et al., 2011). They can appear as a number of smaller pits that have combined to form a scallop and exhibit bright bands and stepped shapes (Soare et al., 2007; Séjourné, 2011).

Scallops form as the result of the degradation of an ice-rich mantle by sublimation or the thaw of ice (Morgenstern et al., 2007; Lefort et al., 2010; Ulrich et al., 2010; Zanetti et al., 2010; Séjourné, 2011). They might also form as the result of the interactions of pooled water with ice-rich permafrost through the collapse of a basin as pooled water evaporated or sublimed (Costard and Kargel, 1995; Soare et al., 2007).

Figure 4.14: CTX mosaic image showing the type example for scallops.

4.2.2.15 Isolated Mounds

Isolated mounds are high relief dome features (Figure 4.15). They can be found with superposing mantling deposits or they can stand out from mantling deposits. Such mounds might be frost mounds like pingos which indicate the presence of permafrost (Dundas et al., 2008; Burr et al., 2009; de Pablo and Komatsu, 2009). Pingos are thought to form by the growth of an ice-core and the subsequent up-doming of the ground (Dundas et al., 2008; Burr et al., 2009; de Pablo and Komatsu, 2009). Mounds might also be caused by the
preferential preservation of degrading mantling deposits (Dundas and McEwen, 2010; Hauber et al., 2011), or result from the cratering process and represent lumps of ejecta or central peak/peak-ring material.

Figure 4.15: CTX mosaic image showing the type example for isolated mounds.
4.2.2.16 Hummocky Terrain

Hummocky ground (Figure 4.16) is defined as terrain displaying multiple mounded features or “hummocks”. This terrain is generally found in undulating terrain within the outer ejecta, and often near to polygons and boulder fields. Hummocks can result from freeze-thaw processes (Mangold et al., 2004; French, 2007), via the upward displacement of material as a result of ice lenses (French, 2007). The mounded features which compose hummocky ground are far smaller than the isolated mounds presented in section 4.2.2.15.

Figure 4.16: CTX mosaic image showing the type example for hummocky terrain.
4.2.2.17 Flat Pitted Ground

Flat pitted ground (Figure 4.17) is a surface type that is heavily pitted and relatively dark. The pitted surface type could be indicative of a desiccated mantling deposit, and is the unit into which channels are commonly incised (Dickson et al., 2009; Hobley et al., 2014). Pitting might result from the interaction of warm ejecta with snow-or ice covered ground, or, melting or sublimation of ice after ejecta emplacement (Jones et al., 2011). This surface type will be classed as a morphological unit in future mapping (Chapter 5).

Figure 4.17: CTX mosaic image showing the type example for flat pitted ground.
4.2.2.18 Impact Structures

Impact structures (Figure 4.18) were mapped, as different crater morphologies could provide evidence of processes that have occurred in and round Lyot crater. For instance, pedestal craters could indicate the deflation of debris and/or the sublimation of ice surrounding the crater (Kadish et al., 2008, 2010). Impact craters might also be later used to date units via crater counting, so it useful to know where they are located.

Figure 4.18: CTX mosaic image showing the type example for impact structures.

4.2.3 Grid Mapping

To conduct the grid mapping, the CTX mosaic was divided into a grid of 1680 squares, each 20 x 20 kilometres in size. In ArcMap 10.1, this was done through the creation of a polygon feature-class shapefile, where each grid square was a polygon object. The shapefile attribute table was then given a new column for each landform/surface type. I determined that, as this method is being used for reconnaissance purposes, alternating squares across the study area would be viewed, as this is sufficient to indicate the distribution of landforms/surface types. These squares were viewed at a scale of 1:20,000, and each selected landform was described as either “dominant” (2), “present” (1), “possible” (P), “absent” (0) or “no data” (N) within the grid squares’ attribute table. In this way a coarse-resolution map is created which shows the distribution of the different landforms across the area.
In the creation of this map there are certain attributes present for the purposes of quality control use. These include:

- **CTX Quality**: This is a measure as to whether or not the CTX is of a good enough standard to mark down whether or not features are present e.g. are there clouds or blurred areas. Ranked as 1 or 0.

- **CTX Coverage**: This is a measure as to how much of the grid square is covered by CTX data and if there are any areas that have no data. Ranked as a percentage.

- **CTX Mappers Sign Off**: This is an attribute that, once filled by the mappers’ initials, will leave the grid square opaque; thereby indicating that it has been looked at.

- **Comment**: This is an area for any general comments about features within the grid square that are not covered by the attributes, and/or the noting down of particular facts about features of interest e.g. the type of crater morphology, the distribution of a feature etc.
4.3 Results

Within this section reconnaissance grid mapping results for the landforms and surface types are presented. Only the landforms and surface types for which patterns are found in their distributions are displayed, other results that are not visible in this section have been placed within Appendix A.

Figure 4.19 shows the squares that were mapped within the study area as black squares. All figures have the central peak (yellow), inner peak ring (blue), crater rim (red), inner ejecta extent (pink) and outer ejecta extent (green) marked as lines, so that the geographic context of the various features of interest can be easily identified. The figures in the following sections have dark green squares if the feature is “dominant” (2), light green squares if the feature is “present” (1), yellow squares if the feature is “possibly” present (P), and black squares if there is “no data” (N).

**Figure 4.19:** Grid mapping squares that have been mapped across the Lyot crater study area shown as an opaque black colour. Features such as the different ejecta extents, crater rim, inner peak-ring and central peak are displayed in different colours to allow easy distinction of which regions of the crater features are located in.
4.3.1 Viscous Flow Features

Figure 4.20 shows the location of viscous flow features (VFFs) across the mapped area. They are predominantly located on the southern crater rim and north-facing slopes in the interior of Lyot crater. VFFs likely develop preferentially on pole-facing slopes where they are largely sheltered from melting due to solar insolation (Milliken et al., 2003; Arfstrom and Hartmann, 2005). VFFs in the outer ejecta blanket are concentric crater fill deposits found within relatively large impact craters in these areas. Some VFFs are also seen to the far south of the mapping area and are associated with Deuteronilus Mensae deposits, and not with Lyot crater. VFFs have been mapped as a morphological unit in Chapter 5.

Figure 4.20: Grid mapping result displaying where viscous flow features were observed.
4.3.2 Small Channels and Flat Pitted Ground

Figure 4.21 shows that small channels are located predominantly within the crater interior and in the inner ejecta blanket. This might indicate that liquid water could only be found in these areas, or that channels in the outer ejecta have become overlain by younger deposits such as mantle units.

![Grid mapping result displaying where channels were observed.](image)

Figure 4.21: Grid mapping result displaying where channels were observed.

Channels are commonly observed incising the flat pitted ground (Dickson et al., 2009; Hobley et al., 2014), thus it is expected that flat pitted ground and channels would broadly be located in the same areas. It can be seen in Figure 4.22, that flat pitted ground is observed predominantly within the inner ejecta and crater interior. This unit is more widespread than the channels, though the channels are often found associated with it. This indicates that although the channels might incise this unit, the presence of this unit does not determine if the channels are present or not. The flat pitted ground was mapped as a morphological unit named the pitted floor unit within Chapter 5, and the potential relationship with the channels was investigated further within Chapters 5 and 6.
Figure 4.22: Grid mapping result displaying areas where flat pitted ground was observed.
4.3.3 Polygonal Networks (high-centred, low-centred and clastic)

Figure 4.23 shows that clastic polygonal networks identified with certainty are found only within the outer ejecta blanket. They extend further than expected to the south-east which indicates that the outer ejecta might extend further in this area than anticipated. Clastic polygons are also observed closer to the crater rim in the south-west than expected, which could be indicative of closer ejecta deposition as a result of an angled impact (Kenkmann et al., 2014). The clastic polygons are located in a band ranging from the north to south-east within the outer ejecta blanket. To the north of the crater there is a region located close to the outer ejecta margin where clastic polygons are absent; this corresponds to an absence of boulder fields (see section 4.3.4), which is likely the result of features in this location being covered by mantling deposits (see section 4.3.5).

Figure 4.23: Grid mapping result displaying areas where clastic polygons were observed.
Figure 4.24 shows that low-centred polygonal networks are found in similar areas to the clastic networks. Unlike for clastic polygonal networks and boulder fields (see section 4.3.4), low-centred polygonal networks are not completely absent in the north, close to the outer ejecta margin. This could suggest that clasts are draped by mantle and thus clastic polygon networks have a surface expression as low-centred networks with no clasts. The similar distribution of low-centred and clastic polygonal networks indicates that they might be related. Low-centred polygons can be seen further to the west of Lyot crater than clastic polygons.

Figure 4.25 shows the grid mapping result for high-centred polygons. These polygons are predominantly found in a latitudinal band crossing the centre of Lyot crater. They do not appear to be related to clastic and low-centred polygon networks. Large kilometre scale polygons, often rectilinear, are found predominantly in the crater interior. These large networks are distinctive and appear unrelated to smaller high-centred polygons observed, due to this it has been mapped as a separate morphological unit in Chapter 5.
Figure 4.25: Grid mapping result displaying areas where high-centred polygons were observed, with preliminary mapping of the kilometre-scale polygon terrain shown in pink.
4.3.4 Boulder Fields

Figure 4.26 shows that boulder fields are concentrated predominantly within the outer ejecta and tend to demarcate the edges of the outer and inner ejecta blanket. The boulder fields extend beyond the ejecta margin to the south-east which might indicate that the ejecta blanket originally extended further in this direction. To the south-west, boulder fields are found relatively close to the crater rim. This is likely due to the impact occurring at an angle and ejecta therefore depositing closer to the crater rim in this area. There is also a location to the north of the crater where no boulder fields are observed; this is likely the result of overlying mantle units obscuring boulder fields (see Section 4.3.5).

Figure 4.26: Grid mapping result displaying areas where boulder fields were observed.
4.3.5 Mantling deposits, Relief Filling Deposits and Multiple Layers

Mantling deposits and relief filling deposits occur throughout the Lyot crater study area (see the results for mantling and relief fill within Appendix A). Figure 4.27 displays the distribution of multiple layers, potentially indicative of multiple generations of mantling deposits. Multiple layers were observed particularly in the outer ejecta, though less layering is observed to the southwest of the crater, coinciding with where the later Deuteronilus Mensae deposits are found. This could offer information about the extent of the climatically-controlled deposition of dust and ice-rich airfall material.

Another interesting feature of this result is that there is an area to the north of Lyot crater where multiple layers are a dominating feature (i.e. they cover nearly almost all the surface). Furthermore, there are two smaller regions to the east and west where multiple layers of mantling deposits are observed. In these locations boulder fields, low-centred polygons and clastic polygons are commonly absent, which might indicate that these features are being obscured by the mantling deposits.

![Figure 4.27: Grid mapping result displaying areas where multiple layers of mantling deposits were observed.](image)
4.4 Further Investigation of the Polygons

To further investigate the distribution of the clastic polygons, a pre-existing shapefile containing point data displaying the approximate locations of clastic polygons – supplied by Matthew Balme – was used. These point data contain a point for each cluster of polygons that covered a region of hundreds of metres to kilometres in area. My work built upon these data by adding extra points to the shapefile, thus, the resulting distribution map displays the locations were polygons are most prevalent. Figure 4.28 demonstrates that polygons are most common in an arc ranging from the north to the south-east region, within the outer ejecta blanket. They are located in a relatively confined band of approximately 200-350 kilometres from the crater rim. This provides further evidence that the polygons have a genetic link to the impact event.

Figure 4.28: Lyot crater study area displayed using colourised Mars Orbiter Laser Altimeter (MOLA) topographic data overlain on a MOLA hillshade. The crater peak, inner peak ring, crater rim, inner ejecta scarp and outer ejecta extent are indicated by black lines. Regions of clastic polygonal features are marked as white circles. This Figure is also presented as Figure 2 by Brooker et al. (2018).
4.5 Conclusions

Within this chapter landforms potentially indicative of water/ice were located across the Lyot crater study area using grid mapping. These results were used to select terrains and landforms to include in an in-depth geomorphological map (presented in Chapter 5), as well as to select landforms for further study. Features highlighted for further study are the clastic polygonal landforms and fluvial channels.

Fluvial channels are generally constrained to the crater interior and inner ejecta blanket. This association could indicate a relationship with this material, though previous work has suggested a link to glaciers found within Lyot crater (Dickson et al., 2009). Although channels are observed within the inner ejecta blanket (Figure 4.21), there is a higher concentration of channels within squares mapped in the crater interior. This could indicate that water, and thus potential melting, was more prevalent within the crater interior. This might provide evidence of a microenvironment within Lyot crater as invoked by Dickson et al. (2009). Thus, they are a feature that could represent the melt of atmospherically-derived deposits such as the mantling units and viscous flow features. The study of these channel features is presented in Chapter 6.

Clastic polygonal features are found in a distinctive arc within the ejecta from the north to the south-east, constrained within the outer ejecta blanket. This distribution indicates a genetic link to the impact event, possibly as a result of deposited ice-rich material that has been exhumed from a specific depth during the Lyot-forming impact event. Thus, this material is potentially representative of ancient subsurface ice deposits. The study of these features is presented in Chapter 7.

Furthermore, the work conducted in this chapter formed the basis for the definition of morphological units within a more in-depth mapping project that provides context for fluvial activity within Lyot crater. These units include the mantling deposits, VFFs, flat pitted ground and kilometre-scale polygon terrain which were included in the grid mapping. The mapping of these units is presented in Chapter 5.
Chapter 5: Geomorphological Mapping of Lyot Craters’ Interior

5.1 Introduction

To understand the history of water at Lyot crater, the emplacement of materials and the modification history of this crater needs to be better constrained. Through the separation of the terrain into various geomorphological units based upon a variety of criteria, followed by a study of their stratigraphic relations, information such as whether water-related landforms of interest are associated with particular units and at what stage in the craters’ history these features formed, can be discerned. Toward this aim, the following chapter presents the approach and results of geomorphological mapping of the interior of Lyot crater. The science outcomes of the map will be discussed in context with the results of the following chapters.

5.1.1 Background

The style of mapping used within this chapter involves the dividing of an area into a variety of distinct units based upon surface properties such as albedo, degradation state and the inferred vertical associations between landforms and units. Hence it is “morpho-stratigraphic” mapping. Mapping is conducted by first identifying boundaries between units and marking them as: certain, when the boundary is clear; approximate, when there is a boundary but the exact position is not clear; and gradational, when there is not a boundary, instead one unit gradually changes into the next. A variety of geomorphological features of interest such as channels, viscous flow features and gullies were also distinguished and mapped across the area. Features are identified based upon their spatial and textural properties (Chapter 4).

Within the previous chapter, the work focussed on identifying and locating landforms and surface types of interest across the entire Lyot study area. Since the study area for Lyot crater is large (>600 km²) I decided that more detailed mapping should focus on a smaller section of particular interest. As viscous flow features and potential fluvial channels are constrained predominantly within the crater rim (Chapter 4), the interior of Lyot crater was selected as the focus of mapping. This is because these features are probably related to recent water sourced from the atmosphere (Dickson et al., 2009; Hobley et al., 2014) and so are relevant to my research. Mapping can help determine the stratigraphic relationships of the landforms and therefore establish whether they are primary features, or likely to be sourced from the atmosphere after the impact event. It is also important to constrain
whether such features are related to any particular morphological unit and, in particular, if the channel features are associated with potentially ice-rich material, such as the icy mantling units or viscous flow features, and thus could be the result of the melting of these deposits – this topic is explored further within Chapter 6.

Within the following chapter, the surface features that make up the various units mapped within Lyot crater are described and interpreted to better understand the modification and post-deposition history of the crater. The location and relation with fluvial and glacial features is also discussed to support interpretations of the history of water in the interior of Lyot crater.

5.1.2 Research Questions

The overriding aim of this work is to better understand the stratigraphy of Lyot crater and the materials that infill it, and, in doing so, to gain deeper insight into the water-related activity which formed the various fluvial and glacial features located within the crater interior. To further this aim, several research questions are posed which will be studied further and discussed throughout this chapter.

5.1.2.1 What is the stratigraphic history of Lyot crater?

The initial part of this chapter is focussed upon the description of a variety of units which have been mapped across the interior of Lyot crater. A unit is recognised by surface characteristics that make it unique and identifiable when compared with other surface types, and that has a discernible boundary (although this can be gradational or inter-fingering) with other units. Observational data is then used to distinguish which units are superposing others, and thus have a younger relative age. In this way a stratigraphic history of Lyot crater can be constructed.

5.1.2.2 Are there relationships between stratigraphic units and/or geomorphic features of interest in the crater?

An important line of inquiry is to distinguish whether particular units are associated with geomorphic features of interest, such as channels. Within Chapter 4 it has been noted that channels are seemingly associated with a surface type I have termed “flat pitted ground”. This surface type has been shown to be pervasive across the mapping area (Figure 4.22) and, although channels are occasionally present in areas where “flat pitted ground” was not observed, this does not mean that there is not an association between this surface type and the channels. The detailed mapping provides further data to reveal if there is a relationship between the channels and a particular unit and, through study into this units’
relation to other units, an estimate as to the onset of fluvial activity relative to other features can be distinguished. Another important relationship to distinguish is if the area from which the channels originate is linked to a particular surface type or morphological feature (e.g. viscous flow features, mantle units). This will provide further information as to how the channels originated and an upper bound on when the potential episode of fluvial activity that formed these channels occurred.

Some units across the region have more gradational boundaries. In these cases it is useful to distinguish the relationship between these units to identify if they are indeed separate units, or the same unit with a different degradation state.

5.2 Method

5.2.1 Map Sheet

Mapping was conducted using the context camera (CTX) mosaic within ArcGIS 10.1. These data were chosen because the resolution (6 m/pixel; Malin et al., 2007) is sufficient for the recognition of textural features, but not so high that mapped units are too small to identify on a map of the entire interior region. CTX data are also the highest resolution data available that covers the entirety of the interior region of Lyot. Contacts were mapped through the recognition of surface differences at the resolution of the available images. The map is presented on Map Sheet 1 and a simplified version is presented in Figure 5.1. The focus of the mapping work was on the deposits emplaced after the formation of Lyot, so ‘bedrock geology’ is not mapped separately.
Figure 5.1: Map of the interior of Lyot crater, overlain on a THEMIS daytime base map with a transparency of 50%.
5.2.2 Age Determination

Typically, mapping of a region involves the derivation of estimated ages for the various units, which provides additional information on top of determination of their stratigraphic relationships. Within planetary mapping, estimated ages are generally the result of impact crater counting. This is a technique whereby the size-frequency distribution of impact craters on a particular surface unit is used to derive an age (e.g. Tanaka, 1986; Hartmann, 1999; Michael and Neukum, 2010). Crater counting is a vital technique in planetary science, though its reliability is decreased due to several factors such as: the use of small diameter (< 1km) impact craters (Xiao and Strom, 2012), the size of the area for which crater counts are being measured (particularly if looking at areas less than 1000 km²; Warner et al., 2015), the presence of secondary craters (McEwen and Bierhaus, 2006; Robbins and Hynek, 2011) and uncertainty in the identification of craters from other features such as sublimation pits (Michael and Neukum, 2010). In addition, it must be recognised that an impact crater-derived age is a “crater-retention model age” – it is not a unit formation age in the strictest sense.

Lyot crater has been dated to early Amazonian in age based upon crater counting (Tanaka et al., 2005) and stratigraphic relationships (Dickson et al., 2009). It has a large number of secondary craters, located up to a minimum of 520 km from the impact crater (Robbins and Hynek, 2011). Previous work (see Chapter 4) has also demonstrated that there are a number of surface features indicative of sublimation that could be mistaken for small impact craters. The relatively young age of the crater, coupled with the presence of a high volume of secondary impact craters, relatively small surface area of individual units, and the presence of features that could be mistaken as impact craters could lead to unreliable dating of units. Therefore, it was decided that crater counting should not be attempted for the units in this region and that relative stratigraphy would be used instead.

5.3 Crater Units

The crater units (Figure 5.2) potentially represent the original bedrock material of the crater, though heavily modified by the impact event and later fluvial and/or glacial processes operating within the area. The units within this group are the lowest stratigraphically and are overlain by the mantling and pitted units. It is not certain if these features represent “basement” or not, though this will be examined and explained further within the following section and within the discussion.
5.3.1 Central Crater Unit (CcU)

The central crater unit has a high relative albedo compared to the other mapped crater units (Figure 5.3). It is rough-textured and often fractured, giving it a ‘polygonal’ texture in places (Figure 5.4). The unit is the topographically highest material within the interior of Lyot crater (Figure 5.5). Its location and topography indicate that this unit represents the central peak formed during the impact event.

Why this unit has a distinctive high albedo is not well understood. Initially, when I observed this (especially around the large dune field in Figure 5.3) I assumed that the high reflectance indicated the presence of CO$_2$ frost on the surface of the dunes rather than the underlying unit showing through. The lack of any such frost on the dunes implies that the high albedo is unlikely to be the result of preferential frost accumulation. It is also unlikely to be the result of a layer of dust that is only being deposited on top of the central peak material. Therefore, the albedo is either the result of the composition of this unit or it is a result of the unit being well-cemented and relatively bright in colour when compared with the overlying deposits, and therefore reflecting more light. Previous work by Carter et al. (2010), Pan et al. (2017) and Pan and Ehlmann (2018) has detected hydrated minerals at Lyot crater, however it is noted that the sites where most of the hydrated minerals were
detected is not the central peak material but instead the peak ring, crater rim and ejecta material (Carter et al., 2010; Pan et al., 2017; Pan and Ehlmann, 2018). CRISM data for the central peak material show that the unit has a strong signature for olivine or iron phyllosilicates, indicating a potential hydrated signature (Figure 5.6). Similarly, work by Pan and Ehlmann (2018) has found an unidentified hydrated signature. This indicates the high albedo could be related to a high hydrated mineral content.

**Figure 5.3:** CTX mosaic showing two larger outcrops of the central crater unit and bright ‘specks’ of this unit within a large dune field. The material has a high albedo and could be mistaken as ice on the dunes. Black lines represent mapped boundaries. The red box within the inset map indicates the extent of the main image.
Figure 5.4: CTX mosaic showing the central crater unit. The unit has a rough texture and appears fractured. This outcrop is surrounded by the smooth mantle unit. Black lines represent mapped boundaries. The red box within the inset map indicates the extent of the main image.

Figure 5.5: Map of the interior of Lyot crater displaying the location of the central crater unit presented on a THEMIS daytime base map.
Figure 5.6: CRISM-derived product for mafic mineralogy with mapped boundaries (fine grey lines) and major linear features overlain. Red (OLINDEX) indicates olivine or iron phyllosilicates. Green (LCPINDEX) indicates low Ca-pyroxene. Blue (HCPINDEX) indicates high-Ca pyroxene. In general, the central crater unit is associated with a crimson red indicating a high olivine or iron phyllosilicate content. The red box within the inset map indicates the extent of the main image.

The central crater unit is stratigraphically lower than the dune field, smooth mantle units and the various pitted units (Figure 5.7). The stratigraphic relationship with the other crater units cannot be constrained as they are not located in the same region.
Figure 5.7: CTX mosaic with morphological units overlain. The central crater unit has not been coloured and is the lowest unit stratigraphically in the region, although highest topographically – it is an inlier. The flat pitted units overlie the central crater material and the smooth mantle unit is observed to be draping slopes and infilling depressions. The red box within the inset map indicates the extent of the main image.

5.3.2 Inner Crater Unit (Icu)

The inner crater unit is similar in morphology to the central crater unit however it generally has a smoother and less fractured appearance (Figure 5.8). The unit is visible as topographically high mounds and ridges, often being revealed from beneath textured and smooth mantle units, and from beneath the various pitted units (Figure 5.8). The less fractured appearance of this unit - when compared with the central crater unit - could be the result of the abrasion of the unit by aeolian material, or obscuration of the unit by a layer of dust or mantle, thus giving it a smoother appearance.
Figure 5.8: CTX mosaic with morphological units overlain. The inner crater unit has not been coloured and is observed as topographically high mounds and ridges across the region. Although this unit still has a relatively high albedo it is not as high as the central crater unit. The red box within the inset map indicates the extent of the main image.

Previous work by Carter et al. (2010, 2015), Pan et al. (2017) and Pan and Ehlmann (2018) has reported the presence of hydrated minerals within the inner crater rim unit of Lyot crater (Figure 5.9). Such material has been interpreted as possible hydrated silicates (Carter et al., 2010; Pan and Ehlmann, 2018), chlorite (Pan and Ehlmann, 2018) and Al-clays (Carter et al., 2015). The topography and location of the inner crater material indicates that it represents Lyot craters’ peak/ring material; it is topographically the highest material within the crater interior. The presence of hydrated minerals in this region could represent the excavation of hydrated material during the impact event to form the ring material, or the subsequent modification of this unit by hydrothermal activity shortly after the impact event (Pan and Ehlmann, 2018).
Figure 5.9: Map of the interior of Lyot crater displaying the location of the inner crater unit presented on a THEMIS daytime base map. The sites of hydrous mineral detections, made using CRISM data from Carter et al. (2015) that intersect with the inner crater unit, are represented as green circles.

5.3.3 Rugged Floor Unit (Rfu)

The rugged floor unit is located in the south-west interior of Lyot crater between the inner and outer crater units (Figure 5.10). The unit has a similar morphology to the inner crater unit as it has a higher albedo than surrounding materials, and has a rough textured, yet not fractured, surface (Figure 5.11). It is, however, less textured than the inner crater unit and is overlain by the inner crater unit (Figure 5.11). For this reason, I conclude it should be classified as a crater unit: it does not appear to be a surficial unit and is stratigraphically the lowest material in the region it is located.
Figure 5.10: Map of the interior of Lyot crater displaying the location of the rugged floor unit presented on a THEMIS daytime base map.

Figure 5.11: CTX mosaic with morphological units overlain. The rugged floor unit has not been coloured. The rugged floor unit is observed as topographically higher material when compared with surrounding pitted and drift related units but is overlain by the inner crater unit. The inner crater
unit is generally more rough textured when compared with the rugged floor unit and is topographically higher. The red box within the inset map indicates the extent of the main image.

Another characteristic feature of this unit is that it is topographically lower than the other crater units (Figure 5.12). It is found in the topographically lowest part of Lyot crater and could represent the deepest material excavated during the impact event (Grieve et al., 1981). There is no CRISM data available for this unit so it is unknown whether it exhibits a hydrated mineral signature.

**Figure 5.12:** CTX mosaic with MOLA topographic data overlain. The inner crater unit is topographically higher than the rough floor unit. The red box within the inset map indicates the extent of the main image.

### 5.3.4 Outer Crater Unit (Ocu)

The outer crater unit is, unsurprisingly, the outermost of the crater units mapped and marks the crater rim and the boundary outside of which the inner ejecta, as opposed to crater interior, would be mapped. I decided that no further mapping would be conducted beyond the crater rim, so this unit is only seen in small amounts on the map (Figure 5.13). It is visible within the walls of the crater exposed where textured mantle is not present (Figure 5.14).
Figure 5.13: Map of the interior of Lyot crater displaying the location of the outer crater unit presented on a THEMIS daytime base map.

Figure 5.14: CTX mosaic with morphological units overlain. The outer crater unit has not been coloured. This unit represents the crater rim and wall. It is visible where textured mantle and viscous flow features are not present, perhaps as a result of sublimation. The unit, though rough
textured on close examination, appears to have been smoothed. The red box within the inset map indicates the extent of the main image.

The outer crater unit is morphologically similar to the other crater units and is stratigraphically lower than the pitted and mantle units. A key difference when compared to the other crater units is that it is characterised by a smoother texture. This is perhaps a result of the surface being smoothed by winds associated with the crater rim, or the viscous flow of icy mantling material (Figure 5.14). Its stratigraphic relationship to the other crater units is unknown because it is commonly covered by textured mantle and thus there is no clear boundary between it and the other crater units.

Work by Carter et al. (2015), Pan et al. (2017) and Pan and Ehlmann (2018) have indicated the presence of chlorite-prehnite on the outer rim of Lyot crater. Prehnite has been located in a number of crater walls in a variety of regions on Mars and is formed at temperatures of between 200°C and 400°C (Murchie et al., 2009; Carter et al., 2010; Ehlmann et al., 2011). The presence of prehnite indicates that this crater rim unit could preserve evidence of hydrothermal alteration (Pan and Ehlmann, 2018). This could have occurred within crustal material before the formation of Lyot crater and then been exposed by the impact event, or it has been formed through hydrothermal alteration as a result of the impact event (Murchie et al., 2009; Carter et al., 2010; Ehlmann et al., 2011; Pan and Ehlmann, 2018).

5.4 Pitted Units

Pitted units are the most common surface type in the mapping region (Figure 5.15). They can be potentially viewed as an intermediate unit between the crater units and mantle deposits. They overlie the crater units and underlie the mantling deposits. These units are located in parts of the crater interior that are flat-lying, in-between topographically high crater units visible as mounds and ridges. They are termed “pitted” as the common characteristic of these units is the presence of pits, often infilled with dark and smooth material. It can also be seen in Figure 5.15 that channel landforms are associated with the pitted units across the mapping region. This association will be discussed further in Sections 5.7 and 5.8.
Figure 5.15: Map of the interior of Lyot crater displaying the location of the pitted units presented on a THEMIS daytime base map. Units are located across the entire mapped region and constitute the majority of this region.

5.4.1 Pitted Floor Unit (Pfu)

The pitted floor unit is the most common unit found across the mapping region (Figure 5.16). It can be seen in figure 5.16 that this unit is located in the region between the inner and outer crater units. It is not as prevalent in the south-west of the mapping region as a result of the overlying of textured mantle and the presence of the rugged crater floor unit. Interestingly this also correlates with the lowest part of the Lyot crater floor. The flat pitted ground is best preserved in the interior region of Lyot crater. Beyond the margins of the crater, the pits, for which the unit is named, become far less well-defined.

The surface morphology of this unit is distinctive when compared to the other units mapped across this region. It is characterised by a rough texture, with a generally low-relief surface containing pits of various sizes (Figure 5.17). These pits vary in form from near perfect circles, which could represent impact craters, to small irregularly-shaped and degraded pits. On close inspection, it can be seen that the pits appear to contain, or reveal, a dark smooth unit (Figure 5.18). This could be representative of the infill of depressions by a later smooth mantle deposit. The highly pitted nature of this unit appears to indicate loss of a volatile component - normally on Mars, this would suggest the presence of once ice-rich ground that has lost ice through sublimation, leading to the formation of numerous

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pits. In this case the unit could represent a degraded mantling deposit, or an ice-rich sedimentary cover analogous to thermokarst terrain on Earth (Soare et al., 2008).

**Figure 5.16:** Map of the interior of Lyot crater displaying the location of the pitted floor unit presented on a THEMIS daytime base map.
Figure 5.17: CTX mosaic with morphological units overlain. The pitted floor unit has not been coloured. Pitted floor material is located in flat and low-lying regions of the crater interior. It is characterised by a rough texture at decameter-scale with a heavily pitted surface. The red box within the inset map indicates the extent of the main image.

Figure 5.18: HiRISE image (ESP_026308_2315) showing a close-up view of the pitted floor unit. It can be seen that it contains numerous pit features at a variety of scales. Some of the larger circular
pits containing a smooth, dark infill are likely impact craters, although they have a degraded rimless appearance. This could be a result of impacts into ice-rich material. The red box within the inset map indicates the extent of the main image.

### 5.4.2 Dark pitted unit with depressed polygons (Ddu)

Although this unit has been mapped as a separate unit from the pitted floor unit, it is most likely a subtype of the pitted floor unit. The unit is characterised by the same rough textured pitted surface as the previous unit, with one striking difference: this unit has large kilometre-scale trough-like polygonal features (Figure 5.19). The morphology and possible origins of the polygonal landforms will be discussed in further detail within Section 5.7. The unit surrounds the central crater unit within the peak ring, perhaps suggesting that the polygonal features could be related to the impact event which formed Lyot crater (Figure 5.20).

![Figure 5.19: CTX mosaic showing the south-eastern region of the dark pitted unit with depressed polygons with morphological units overlain. The unit of interest is not coloured. Broadly, this unit has a similar morphology to the last unit discussed with the exception of the presence of large kilometre-scale polygonal landforms. The red box within the inset map indicates the extent of the main image.](image-url)
Figure 5.20: Map of the interior of Lyot crater displaying the location of the dark pitted unit with depressed polygons presented on a THEMIS daytime base map.

This particular unit has a range of surface characteristics depending on the location within the crater. The south-east region of the unit features relatively “pristine” polygonal forms with occasional large pits which appear to have been infilled by a younger smooth mantling unit (Figure 5.19). The western and northern regions of the unit (Figure 5.21) features more degraded polygonal forms with large numbers of dark pits (often clustered together). Polygons edges (troughs) are discontinuous and many appear almost channel-like in form. This has led me to conclude that some of the polygons have been later modified by fluvial activity which has made use of these pre-existing depressions in the landscape. Another point to note is that in many cases the troughs are connected by pits. Recent work by Glines and Gulick (2018) has indicated that this assemblage of potential channels interconnected by pits might represent a thermokarst paleolake assemblage. This supports the view that the pitted units are likely ice-rich in nature.
Figure 5.21: CTX mosaic showing the western region of the dark pitted unit with depressed polygons with morphological units overlain. The unit of interest is not coloured. The polygons in this region are more degraded than those present in the south-eastern region, as is the surface in general, and far more dark pits can be seen, often clustered together. Some polygon margins appear similar in morphology and pattern to fluvial channels. The red box within the inset map indicates the extent of the main image.

5.4.3 Dark pitted unit with inverted polygons (Diu)

As with the previous unit, this unit represents a change in character of the pitted floor unit as opposed to a complete change of unit type. This particular unit is the smallest mapped and is located within the peak ring region of Lyot crater (Figure 5.22). This unit is characterised by a similar surface texture to the flat pitted unit, however with distinct differences. The surface is rough in texture, though smoother than the other pitted units. Surprisingly, this unit has far fewer pits present and any pits that are present are shallower, smaller and more degraded (Figure 5.23). Similarly to the previous unit, this unit features large polygonal landforms, though in this case they have a pseudo-polygonal form, more similar to the less continuous polygonal forms located in the western region of the central crater (Figure 5.21). The defining characteristic of this unit, however, is that the polygonal landforms in this region are inverted in form; the troughs are not present, and the polygon boundaries stand out instead as topographically high features in the landscape (Figure 5.23).
Figure 5.22: Map of the interior of Lyot crater displaying the location of the dark pitted unit with inverted polygons presented on a THEMIS daytime base map.

Figure 5.23: CTX mosaic showing the dark pitted unit with inverted polygons with morphological units overlain. The unit of interest is not coloured. Broadly, this unit has a similar morphology to the last unit however the surface in general appears smoother and less pitted. The polygons in this
case stand out as topographically high ridges of material. The red box within the inset map indicates the extent of the main image.

Although this unit is not located as close in association with the central crater unit as the previous unit, it could represent the southernmost extent of this pitted material within the inner crater ring which has been modified by erosion of the landscape. The erosional process is hard to determine, but aeolian abrasion is a candidate as this process occurs elsewhere on Mars. This is supported by the presence of a large aeolian deposit (i.e. a dune field) in this region which potentially overlies not only the southernmost extent of the central crater unit, but the dark pitted unit with inverted polygons too. In this case the inverted forms of the polygonal features, and the relative smoothness of the landscape when compared to the previous unit, could be explained by abrasion of the surface and the revealing of more resistant, potentially coarse-grained material that was deposited in pre-existing fractures at the polygon margins. This will be discussed further in Sections 5.7 and 5.8.

The relationship of this unit to the other units is interesting. Similarly to the other pitted units it is stratigraphically higher than the crater units and lower than the mantling and surficial materials units. Its boundary with the pitted floor unit, however, appears gradational (Figure 5.23 and 5.24) with the surface becoming less pitted and raised ridges becoming visible as one transitions into the other. It is inferred that, if the boundary between this unit and the dark pitted unit with depressed polygons were visible, it would also have a gradational boundary. It is also inferred that the dark pitted unit with inverted polygons is a modified version of the dark pitted unit with depressed polygons. The latter unit is probably a modified version of the pitted floor unit that has undergone fracturing to form the large-scale polygonal features.
Figure 5.24: CTX mosaic showing the dark pitted unit with inverted polygons with morphological units overlain. The unit of interest is not coloured. The boundary between the units though gradational is still distinct. It can be observed that the pitted floor unit contains distinctive pit features containing infilling dark material, whereas the dark pitted unit with inverted polygons is far smoother and less pitted in texture. The red box within the inset map indicates the extent of the main image.

5.5 Mantling Units

Mantling units are located across the entirety of Lyot crater but are particularly pervasive around the crater rim, inner peak ring and within the south-west region of the crater interior (Figure 5.25). Previous work mapping the extent of mantle deposits across the region (Chapter 4) indicates that the mantling deposits extend across almost the entirety of the Lyot crater study area. Aside from some of the smaller crater ejecta deposits and aeolian surficial material, the mantling units are stratigraphically the youngest in the entire region. They could represent latitude-dependent circum-polar mantles deposited as a result of changes in obliquity (e.g. Mustard et al., 2001; Kreslavsky and Head, 2002a; Head et al., 2003); this will be discussed further in Section 5.8. Another point of interest is that the possible origination points for the various channel features often appears associated with mantling units, this potential relationship will be discussed further in Chapter 6.
5.5.1 Smooth Mantle Unit (Smu)

Aside from aeolian surficial materials, the smooth mantle unit is stratigraphically the youngest material mapped in the region (Figure 5.26). It is the only unit to have been observed infilling and/or draping possible fluvial features such as the channels and fan deposits (discussed further in Chapter 6). The unit consists of mantling deposits that have undergone little to no modification, indicated by this unit's generally very smooth surface (Figure 5.27). It is found commonly within pre-existing depressions and resting against the slopes of topographically high features such as the exposed ridges and mounds of the crater units (Figure 5.27). The only definite modification of this unit observed is the presence of gullies where this unit has been deposited against/on steep slopes, such as on the crater wall and central peak (Figure 5.27).
Figure 5.26: Map of the interior of Lyot crater displaying the location of the smooth mantle unit presented on a THEMIS daytime base map.

Figure 5.27: CTX mosaic showing the smooth mantle unit with morphological units overlain. The unit of interest is not coloured. The smooth mantle unit is observed infilling depressed regions and draping the slopes of the central crater unit. The surface texture is smooth with little to no
modification of the surface. Gullies are observed to have formed on the steep slopes of the central crater unit (sloping down towards the NW in this example) and indicate that these features are stratigraphically younger in age than the smooth mantle unit. The red box within the inset map indicates the extent of the main image.

The smooth mantling unit is occasionally observed grading into the textured mantle unit rather than superposing it. This could be a result of the viscous flow of this material on steep slopes leading to textures on the mantle units surface that grade into underlying textured units, or the unit is only present as a thin surficial layer which is removed on slopes due to the effects of solar insolation (Figure 5.28).

![Figure 5.28: CTX mosaic showing the smooth mantle unit with morphological units overlain. The unit of interest is not coloured. In this example the smooth mantle unit appears to exhibit a small amount of texture though it is generally still far smoother than the textured mantle unit. The boundary between the two units is gradational in nature and might be explained as the same unit being modified by the presence of a change in slope or the revealing of the other unit from beneath the smooth mantle. The red box within the inset map indicates the extent of the main image.](image)

The thickness of the unit appears to vary. In certain regions of the crater the unit is thick enough to remove any indication of the texture of the material beneath the unit (e.g. Figure 5.27). In other areas, the smooth mantle does not completely mask the textural properties of the feature or unit that it is deposited on (Figure 5.29).
Figure 5.29: CTX mosaic showing the smooth mantle unit with morphological units overlain. The unit of interest and the fan deposit are not coloured. The smooth mantle unit is observed draping the viscous flow features, fan and channel end, yet the impressions of the texture on the fan deposit, as well as the textural features of the viscous flow features can still be observed. This indicates either material has been preferentially ablated away in these regions or the unit is less thick in this area. The red box within the inset map indicates the extent of the main image.

5.5.2 Textured Mantle Unit (Tmu)

The textured mantle unit is the most common mantle unit across the mapping region (Figure 5.30). It is most commonly associated with the crater rim and inner peak ring and is found on generally steep slopes (Figure 30). The unit is broadly defined as any mantling unit that displays a textured surface.

The most common surface texture of this unit is heavily pitted, where pits are often aligned in approximately parallel rows (Figure 5.31) to give a “ridge and furrow” appearance. The pitted texture is oriented perpendicular to the downslope direction indicating that this texture could be indicative of the flow of this material. This fits with the observed location of the unit, which is usually on steep slopes. The unit is stratigraphically younger than the crater and pitted units but is stratigraphically lower than the smooth mantle unit. The unit does, however contain a number of generations of mantling deposits within it, as indicated by the presence of layers in many locations in the south west of the crater interior (Figure 5.32).
**Figure 5.30:** Map of the interior of Lyot crater displaying the location of the textured mantle unit presented on a THEMIS daytime base map.

**Figure 5.31:** CTX mosaic showing the textured mantle unit with morphological units overlain. The unit of interest is not coloured. Textured mantle exhibits a highly pitted surface texture possibly indicative of the sublimation of ice. The red box within the inset map indicates the extent of the main image.
Figure 5.32: HiRISE image (ESP_033811_2295) showing a close up view of the stippled texture in the textured mantle unit that is commonly visible in areas where layering is observed.

Figure 5.33: CTX mosaic showing the textured mantle unit with morphological units overlain. The unit of interest is not coloured. In this example, the textured mantle unit has a stippled surface texture with multiple mantling layers visible, this probably indicates multiple depositional events. The red box within the inset map indicates the extent of the main image.
The other common surface texture exhibited by this mantling unit is a stippled texture often visible in regions where layering of the mantle is present. This could represent the sublimation of ice from the deposit, leading to pit formation that gives it a stippled texture as viewed in CTX and HiRISE images (Figure 5.32). In Figure 5.33 it can be seen that the layering is most visible near to the rugged floor unit in the region indicating that erosion of this unit (probably due to sublimation) might be enhanced in this area. The layering is probably indicative of multiple generations of mantle being deposited in this region, potentially as a result of climate cycles controlled by changes in orbital parameters (e.g. Mustard et al., 2001; Kreslavsky and Head, 2002a; Head et al., 2003).

5.6 Surficial Units

Within the region a number of additional features have been mapped (Figure 5.34). They are classified as surficial units but represent features of particular interest which cannot be represented fully as linear features (i.e. viscous flow features and fan deposits), and loose surficial deposits which obscure stratigraphically lower units (i.e. aeolian surficial material) and young impact crater ejecta. They are discussed in more detail in the following section.

Figure 5.34: Map of the interior of Lyot crater displaying the location of the surficial units presented on a THEMIS day base map. Image Credit: NASA/JPL/ASU.
5.6.1 Ejecta Deposits (Ed)

Ejecta material in this case does not indicate ejecta deposited as a result of the formation of Lyot crater, but rather ejecta deposited as result of the formation of smaller (km-scale) craters since the Lyot-forming impact. Such ejecta deposits are small and generally stratigraphically younger than the crater and pitted units. It is, however, rare to observe even very small impact craters and associated ejecta that are located on top of the smooth mantling unit. Young craters and their ejecta are most commonly covered by a surficial mantle through which the crater rims and extents of the ejecta blankets are still clearly visible (Figure 5.35). These small craters and ejecta are not associated with any specific geomorphological features and, being a natural consequence of exposure to the martian impact environment, are of little importance to the history of water at Lyot crater. For this reason, these units will not be discussed further.

![Figure 5.35: CTX mosaic showing an example of ejecta material with morphological units overlain.](image)

The unit of interest is not coloured. It can be seen that this particular crater formed after the inner crater unit, pitted floor unit and textured mantle, as the ejecta superpose these units. It is however draped by a surficial layer of the smooth mantling deposit. The red box within the inset map indicates the extent of the main image.

5.6.2 Aeolian Surficial Material (Asm)

There are a number of aeolian deposits across the region. These deposits are in the form of dune fields or isolated dunes composed of dark material. These deposits are
located in the depressed regions between the central peak and inner peak ring, and between the inner peak ring and crater rim in the southern region of the crater. The aeolian deposits appear to be the youngest features within the crater as they overlie all the other material and are not seen to be covered by any other units. This is demonstrated by the largest dune field in Lyot crater as shown in figure 5.36.

The second largest dune field, located in the south of the crater (Figure 5.37), overlies a distinctive fluvial assemblage consisting of two channels and three fan deposits (discussed further within Chapter 6). Although the dune fields formed later than these fluvial deposits there could be an association between the dunes forming in this region and the deposition and later remobilisation of sand-grade material deposited here by prior fluvial activity.

Figure 5.36: CTX mosaic showing the largest aeolian surficial material within the mapping area with morphological units overlain. The unit of interest is not coloured. The central crater unit can be seen underneath this dune field and is exposed in two larger outcrops in the centre of the deposit. The aeolian material overlies all the other units within the region. The red box within the inset map indicates the extent of the main image.
Figure 5.37: CTX mosaic showing the second largest dune field within the mapping area with aeolian surficial material and fan deposits coloured. It can be seen that this dune field appears to overlie a distinctive fluvial assemblage consisting of two channels and three fan deposits. This association could indicate a link to sand-grade material deposited as a result of the fluvial activity. The red box within the inset map indicates the extent of the main image.

Recent work by Widmer and Diniega (2018) has recorded the formation of alcoves on the brinks of a number of sand dunes within Lyot crater dune fields, indicating that there are active aeolian processes operating within the crater. Such alcove formation could be the result of seasonal CO₂ frost (e.g. Diniega et al., 2010) or dry granular flows due to purely aeolian processes (e.g. Horgan and Bell, 2012). As seasonal frost has not been observed within this region, the formation mechanism of these alcoves is most likely the result of dry granular flows due to aeolian deposition followed by over-steepening and failure of the dune brink (Horgan and Bell, 2012). The dune fields will not be discussed in further detail within this thesis other than to state that they are likely the youngest features within the crater interior, are still active today, and might indicate a source of sand-grade material from fluvial sinks.

5.6.3 Fan deposits (Fd)

Fan deposits are present at the terminations of a number of channels across the mapping region. They are positive relief features and can be divided into a number of
distinct morphological types. The potential origin of these deposits will not be discussed within this thesis chapter, as Chapter 6 focusses on the fan deposits and the related channels in more detail.

Stratigraphically, these features appear to have formed at the same time or shortly after the formation of the channels. This is indicated by the deposits overlying the channels but with some of the fan material sitting within the channel (Figure 5.38). These features are, however, overlain by mantling deposits (Figure 5.29). This indicates that their stratigraphic age is younger than the crater and pitted units, but older than the smooth mantle unit.

Figure 5.38: CTX mosaic showing an example of the fan deposits with morphological units overlain. The fan deposit is not coloured. In this example it can be seen that although the fan deposit can be viewed it is still covered by a layer of mantling material. The unit is linked to two channels and the inverted deposit overlies the end of these channels but is constrained to the channel margins. The red box within the inset map indicates the extent of the main image.

5.6.4 Viscous Flow Features (VFFs)

Viscous flow features (VFFs) are located on the crater rim and inner peak ring predominantly in the south of the crater (Figure 5.39). Three of these features have also been mapped on the north-west crater rim. A viscous flow feature has been mapped where surface texture and morphology indicates the flow of material. This is in the form of surface
lineations and ridges, as well as a generally lobate form (Figure 5.40, Milliken et al., 2003, Arfstrom and Hartmann, 2005, Head et al., 2010). The mapped VFFs are located on slopes. They appear to be formed from materials with similar surface morphology to textured mantle units, although in HiRISE images rocks and polygonal fractures can be observed distinguishing VFFs from this unit (Figure 5.41).

Figure 5.39: Map of the interior of Lyot crater displaying the location of the viscous flow features presented on a THEMIS daytime base map.
Figure 5.40: CTX mosaic showing an example of viscous flow features with morphological units overlain. The viscous flow features are not coloured. The viscous flow feature has a lobate plan-view shape and a surface containing ridges and pits. The red box within the inset map indicates the extent of the main image.
Figure 5.41: HiRISE image (ESP_043358_2295) showing a close-up view of a viscous flow feature. Polygonal features and rocks can be clearly observed on the surface indicating that this unit is not the same as the textured mantle unit.

In Figure 5.42 the ridges located on the surface of the VFFs can be clearly observed. Ridges at the downslope end of the features have been interpreted as potential flow ridges, formed due to accumulated rock and/or dust being pushed to the edges of the flow (Milliken et al., 2003, Arfstrom and Hartmann, 2005). Ridges located higher within the main body of the features have been interpreted as compressional ridges indicative of the ductile flow of material (Milliken et al., 2003, Arfstrom and Hartmann, 2005). Surfaces are often pitted, with large troughs of material that appear to have been sublimed away (Soare et al., 2017). The presence of polygonal fractures is also indicative of the freezing and thawing of ice-rich material and has been observed on debris-covered glaciers in Antarctica (Levy et al., 2006). This provides evidence that such features might be ice-rich. Another indication of this composition is that their locations are commonly pole-facing slopes (Figure 5.39). Such slopes are largely sheltered from ablation due to solar insolation, so ice-rich features might preferentially be located in such areas (Milliken et al., 2003, Arfstrom and Hartmann, 2005). Furthermore, Shallow Subsurface Radar (SHARAD) data indicates many VFFs are composed of massive H$_2$O ice with minimal lithic content (Holt et al., 2008; Plaut et al., 2009; Hubbard et al., 2014). For the reasons listed above these features are classified as ice masses similar to terrestrial debris covered glaciers, which fits with previous observations of similar features on Mars (e.g. Head et al., 2005, Milikovich et al., 2006, Head et al., 2010).
Figure 5.42: CTX mosaic showing an example of viscous flow features with morphological units overlain. The viscous flow features are not coloured. Slope is downwards from bottom of image to top. This image clearly shows the ridges that appear oriented perpendicular to a possible flow direction. Ridges at the flow edge are interpreted as flow ridges. Ridges higher in the body of the glacier are interpreted as compressional ridges. The surface is pitted possibly hinting at the sublimation of ice from the VFF surface. The red box within the inset map indicates the extent of the main image.

VFFs in Lyot crater are occasionally located near to, or in association with, channel features. In some cases, it appears that channels are located at the edges of the VFFs which could be indicative of the VFFs providing a source of water for the channel activity. This relationship is uncertain as in many cases it appears that channels are being revealed from beneath the VFFs rather than being sourced from them. This will be discussed further within Chapter 6.

5.7 Geomorphological Features of Interest

This section contains information about features that have been mapped as linear features or are particularly indicative of a unit's morphology. Such features are either of particular importance to the history of water within Lyot crater (e.g. channels) or provide information about the modification of material and therefore the overall history of the crater (e.g. km-scale polygonal ground).
5.7.1 Channels

Channel features are not discussed in great detail within this section, as Chapter 6 is devoted to detailing channels and fan deposits in more detail. Channels are classified as a depressed or inverted curvilinear feature that is relatively continuous and displays sinuosity characteristic with formation by fluvial processes. Such channels were mapped as a linear feature along their centrelines. Only two examples of the features mapped were inverted in relief, though some channels occasionally display an inverted section (often in association with the channel termination). Channels are kilometres in length and tens to hundreds of metres in width. They are commonly unbranching, but branched examples are also present (Figure 5.43). The centres of the channels are often infilled with a dark smooth material that could represent material transported and deposited as a result of fluvial activity, or could be later infill.

![Figure 5.43: CTX mosaic showing an example of channels with morphological units overlain. The pitted floor unit into which these channels are incised is not coloured. Channels in this example appear to be associated, or at least near to, features that are ice-rich. Two of the channels display branching as part of a contributory network. The largest of the channels terminates in a depression with an associated inverted fan deposit. The red box within the inset map indicates the extent of the main image.](image)

Figure 5.43: CTX mosaic showing an example of channels with morphological units overlain. The pitted floor unit into which these channels are incised is not coloured. Channels in this example appear to be associated, or at least near to, features that are ice-rich. Two of the channels display branching as part of a contributory network. The largest of the channels terminates in a depression with an associated inverted fan deposit. The red box within the inset map indicates the extent of the main image.

Channels are found incised into the pitted ground units and are often located near to mantling units and other such potentially ice-rich deposits. In terms of their stratigraphic
age, they formed after the pitted units, but before or at the same time as mantling units, though this will be discussed further in the following section and chapter.

### 5.7.2 Gullies

Gullies are located on steep slopes within Lyot crater, such as the central peak and crater rim. In the mapping process they were defined as an incised channel with a distinct alcove and occasionally an associated fan (Figure 5.44). They are stratigraphically younger than most of the units mapped in Lyot crater, and fan deposits associated with gullies have been observed overlying VFFs. This indicates that, along with the dune fields, they are the youngest features in the mapping area. The gullies incise into crater units that have been revealed from beneath mantle units, and deposit material on top of the latter unit (Figure 5.44).

![Figure 5.44: CTX mosaic showing an example of gullies on the Lyot crater’s western rim. Gullies have a distinctive alcove, channel and, in some cases, associated fan. The largest of the gullies in this image appears to be depositing material on top of a viscous flow feature. The red box within the inset map indicates the extent of the main image.](image)

The process by which gullies form is a contentious topic that is outside of the scope of this thesis. They might form as a result of the actions of water sourced from the melting of the frost deposited on the exposed crater units (e.g. Head et al., 2008, Conway et al., 2011, de Haas et al., 2015), from the avalanching of fine granular material (e.g. Treiman, 2003)
or the result of a CO₂ suspended flow (e.g. Musselwhite et al., 2001, Pilorget and Forget, 2016). However, the incision of crater units by gullies, and the fact that they are geologically recent, suggests an efficient erosion mechanism which is most consistent with the action of flowing water. If such features are indicative of the action of water then this implies that the most recent fluvial activity at Lyot crater is potentially the melting of deposits on mantled slopes of the crater units.

5.7.3 Km-Scale Polygonal Ground

Kilometre-scale polygonal ground has been found within the pitted units of Lyot crater (see Sections 5.4.2 and 5.4.3). There are three different morphologies into which the landform can be divided across the region: “pristine” polygons (Figure 5.19), “degraded” polygons (5.21) and “inverted” polygons (Figure 5.23).

The “pristine” polygons are located to the south-east of the central peak within a relatively flat and low-lying region. They have diameters of ~800 metres, though this varies across the network. Polygon margins (troughs) consist of depressions that are tens of metres in width. The network generally has 120° intersections. Relatively “degraded” polygons are more common and are located in an arc around the central peak of Lyot crater. The polygons in this region are still on the same scale and have a similar network morphology (when it can be clearly observed), however in very degraded regions such as the western region of polygonal ground (Figure 5.21), polygon troughs are smaller in width and polygons often appear to have multiple parallel troughs (or fractures) bounding them. Finally, the inverted polygons are difficult to fully describe as they are overlain by numerous deposits, so are not seen in full detail. Morphologically they appear similar to the degraded polygons, however polygon margins are characteristically higher than polygon centres.

The large size of these polygons is difficult to explain. Taking the “pristine” polygons as representative of idealised network morphology prior to modification, they appear similar to thermal contraction polygons. However it is difficult to explain the size of these polygons through the thermal contraction of ice cemented sediment (which is the usual process by which such networks form on Mars e.g. Marchant et al., 2002; Ulrich et al., 2011; Brooker et al., 2018). Another polygon type with similar network morphologies are desiccation polygons, formed as a result of the dehydration of volatile-rich material. Intermediate-scale polygons of up to ~300 metres observed on Mars have been attributed to this polygon morphology, but none on the scale observed here (e.g. El Maarry et al.,
It is more likely that polygons in this region are the result of tectonic processes, possibly linked to the impact that formed Lyot crater. McGill and Hills (1992) propose that large polygons can form as a result of the combination of desiccation and the differential compaction of material deposited onto an area of rough topography. The size of the polygons is controlled by the topography over which the material is deposited (McGill and Hills, 1992). This could form polygons on the scale observed however the morphology of this polygon type is very different to the “pristine” examples observed, though could potentially explain the more “degraded” polygons.

The polygon landforms in this case are likely all interconnected as they form in the same pitted material and in approximately the same region of Lyot crater (wrapped around the central crater unit). It is probable that they formed these different morphologies either as a result of differences in the material in which they are forming (i.e. material is thicker, more ice-rich etc.), or they have been modified after formation of the polygonal fractures.

In the case of the inverted polygons these are likely the result of the modification of a pre-dating fracture network. This modification might be the deposition of coarser material in fractures that has then been revealed through downwasting of the surrounding landscape, or is the result of the exploitation of the fractures by hydrothermal fluids depositing minerals and forming a vein network. It is not certain if the “degraded” polygons are the result of differences in the material in which the polygons have formed or a difference in the formation process. It can be seen however that in areas these polygons have been modified by later fluvial activity (Figure 5.21). It is inferred from the sinuous nature of many margins in the western region that fluvial activity has exploited the pre-existing fractures. Potentially, pits at the intersections of margins could have become infilled with water forming small ponded bodies of water (Glines and Gulick, 2018).

Superposition of mantling units on to the polygon landforms indicates that the polygonal features formed after or during deposition of the pitted units, but before the mantling units were deposited. The inverted polygons are also overlain by mantling deposits and aeolian material, so this modification must have occurred prior to their deposition.

5.8 Discussion

5.8.1 History of Lyot Crater Units

Referring back to the research questions for this chapter, the stratigraphic history of the area can now be assembled from the various mapping components and their
relationships to each other. Although the region is divided into nine separate ‘units’ (not including surficial units which formed at varying times in the history of the crater), these units can be combined to form three groups: the crater units, the pitted units and the mantle units. Broadly the formational history of the units across the region is that the crater units were formed first from material uplifted during the impact event that formed the crater, the pitted units were then deposited via infill of the crater terrain by an extensive and relatively thick ice-rich mantle, this unit is then overlain by later mantle units as a result of obliquity cycles. The intricacies of the deposition and modification of the various contained units is more complex (Figure 5.45).
**Figure 5.45:** Summary of the history of the units mapped across Lyot crater providing unit descriptions and interpretations. This figure is featured as part of Map Sheet 1.
5.8.1.1 Crater units

The crater units are stratigraphically the oldest material, indicated by them being superposed by all the units and by the rough-textured and degraded appearance attributed to impact gardening as a result of being exposed on the martian surface the longest. It is difficult to distinguish between these units. They appear to have only very slight differences in surface texture, and perhaps some compositional differences – as described by Carter et al. (2010 and 2015), Pan et al. (2017) and Pan and Ehlmann (2018). The primary differences between the units are their topographic positions and their positions within the crater interior. The central crater, inner crater and outer crater units represent the central peak, inner peak ring, and crater rim material respectively, and thus are approximately the same age. The rugged floor unit, however, is different to the other units: it is topographically lower-lying and is somewhat intermediate in morphology between the crater rim and inner peak ring material. Other than these characteristics, the unit is similar in character to the other crater units. The location of this unit does correspond with the location of the textured mantle unit so it might represent inner peak ring material modified as a result of emplacement or removal of the textured mantle. Whatever the explanation for the slight morphological difference, it appears to have the same relative age to the other crater units.

Surface differences in the various crater units could be a result of both a difference in the composition of the material as a result of the depth from which it originated during the impact event (the deepest material is most likely to be revealed in the crater centre (Grieve et al., 1981)), and the way in which these units have been modified during and after the impact. It is unsurprising, for instance, that the material in different regions exhibits differences in how fractured the units are, and that the units in closest association with aeolian surficial material have experienced surface abrasion.

The hydrated signatures detected amongst the crater units are attributed to either processes which occurred prior to the Lyot-forming impact, or hydrothermal activity which occurred shortly afterwards (Pan and Ehlmann, 2018). If the signatures have been revealed as a result of exhumation during the impact event, they do not represent the history of water within the crater, but the history of water in the area prior to the crater’s formation (Pan and Ehlmann, 2018). Alternatively, the hydrated minerals formed as a direct result of the impact event and represent one of the first stages of aqueous activity to have occurred after the crater formed (Pan and Ehlmann, 2018).
The history of these units is best revealed through the modelling of the impact event, as in this way the depth from which the material originated can be estimated and whether this corresponds to the potential location of past and present subsurface ice or groundwater in the region (see Chapter 8).

5.8.1.2 Pitted units

Pitted units are somewhat enigmatic in their origins. Previous work by Hobley et al. (2014) interpreted this material as dissected mantle (Mustard et al., 2001). Similarly, Dickson et al. (2009) interpreted this unit as stippled mantle. Looking at the evidence presented in the form of textural features indicative of the sublimation of ice, most notably the pits for which this unit is named, I also find that these units are most likely a mantle unit deposited and dissected much later than the original impact event. These units therefore represent the oldest surface expression of mantling deposition in the mapping area. The time period between the impact event and the deposition of this mantling deposit is hard to constrain, but given the lack of km-scale impact craters in all the mantled deposits, it is presumably a significant portion of the age of Lyot crater itself. Previous work by Dickson et al. (2009) involved the dating of the stippled mantle into which channels are incised. They calculated a Middle Amazonian age (~1.5 – 0.78 Gyr), supporting a significant time lapse between the formation of Lyot crater (~3.3 – 1.6 Gyr) and the deposition of the pitted units (Dickson et al., 2009).

This mantle unit is interpreted as being the result of ice-rich airfall material being deposited as a result of differences in obliquity leading to climate cycles and the net deposition of ice at this location (e.g. Head et al., 2003). In this case, the material has been deposited and then sufficient time has passed such that the climate has shifted from net deposition, to net loss of ice. Hence the mantle deposits began to desiccate as ice sublimed from the surface, leaving the pitted dissected mantle unit seen today. This also indicates that the mapped mantle units are representative of the most recent period of net deposition as they are not as dissected.

The three pitted units, although mapped as separate units, are the same deposit that has undergone different modification processes. Working hypotheses to explain these processes are proposed here.

The two pitted units featuring km-scale polygons are difficult to explain. They could form where mantle particularly rich in volatiles has been deposited over the pre-existing impact crater topography. Desiccation of this unit, coupled with the expression of the underlying topography, would lead to differential compaction, causing the formation of
the large polygonal landforms (McGill and Hills, 1992). The morphology of the more “pristine” polygons is difficult to explain by this hypothesis. The polygons could result from thermal contraction in a particularly thick and ice-rich layer around the central peak, though the polygon size is difficult to explain (Brooker et al., 2018). In the region to the south of the central peak (where the pseudo-polygons are located), temperatures and conditions were potentially sufficient to have allowed the circulation of hydrothermal fluids through the sub-surface (Pan and Ehlmann, 2018) and the formation of a mineralised vein network which exploited pre-existing fractures. This unit has then been downwasted, leaving behind topographically high expressions of the resistant mineralised fills. Alternatively, the ridges could result from the deposition of coarse-grained aeolian material which has become cemented and similarly downwasted revealing the more resistant fill material (Brooker et al., 2018). The formation mechanism of the various polygon morphologies is uncertain and further work is needed to explore their history.

Of particular interest to this chapter’s research are features representative of water and the relationship that they have with the various mapped units. The pitted units are of particular importance in this respect as they are closely related to the potential fluvial deposits (channels and fan deposits). Such features appear to form on the pitted units and so this provides an upper time boundary on when the fluvial activity that formed them occurred. The activity must have occurred after the deposition of the pitted units and polygon formation, indicated by fluvial modification of some of the polygon margins, therefore indicating a time-lag between the Lyot-forming impact event and the onset of channel formation. This implies that fluvial activity is unlikely to be the result of latent heat from the impact event. The formation of the channels will be discussed in further detail in the next chapter.

5.8.1.3 Mantle units

Mantling units are the youngest non-aeolian units mapped in Lyot crater. They exhibit a range of textures indicative of an ice-rich nature. They are interpreted as airfall deposits, rich in ice and dust, deposited as a result of changes in climate due to differences in obliquity (e.g. Head et al., 2003). This is the same mechanism by which the pitted units formed however the less desiccated appearance of the mantle units indicates that they were deposited in a more recent period of mantle deposition. The presence of textured and smooth deposits is interpreted as being broadly reflective of the relative age of the various mantle layers: smooth deposits being relatively younger, or potentially less icy or
having a less porous lithic upper layer, and textured deposits being older or initially more ice-rich with more porous insulating deposits.

The presence of a larger volume of mantling material in the southern region of Lyot crater is probably a result of the region being better protected from solar insolation as the crater rim slope is poleward facing, or the mantle was preferentially deposited here. In some regions of the mapping area, layering of the mantle has been observed, indicating multiple generations. This could be evidence for climate cycles driven by planetary-scale obliquity cycles. The smooth mantle unit is the most recently deposited of these units and is present over all the other deposits – aside from the aeolian drift material – as a thinner veneer of draping material. There is no layering observed within this unit indicating it is likely the most recently deposited mantle. This smooth mantle unit is also observed infilling topographic lows, so could represent the last remnants of a once more continuous blanketing unit that has been largely removed on flatter lying areas.

Viscous flow features are interpreted as debris-covered glaciers (Head et al., 2005, Milikovich et al., 2006, Head et al., 2010). They superpose the crater and pitted units, but are often superposed by the smooth mantle unit. The viscous flow features are commonly found associated with the textured mantle unit. I propose that they formed during the same periods of net ice deposition as the mantling units.

Gullies are interpreted to have formed both before and after the deposition of the youngest mantle units; there are examples of gully fans both superposing and being superposed by the mantling deposits. This is potentially the most recent evidence of fluvial activity in Lyot crater – assuming these features formed by aqueous processes. The aeolian deposits are the only unit to superpose the mantling units in all cases, indicating that aeolian processes are the most recent activity in Lyot crater.

5.8.2 Implication

The wider implications of this work are that the fluvial activity related to the formation of channels is unlikely to be associated with the impact event. The presence of the channel deposits on pitted units indicates that the fluvial activity that formed them occurred after the impact event and after sufficient time had passed for a mantling unit to be deposited on the crater interior. This means that the channels record more recent water activity, sourced from the atmosphere, rather than the melting of ice deposits released during the impact event. The channels and associated deposits will be discussed in more detail in the
next chapter, with the aim being to reveal more about the timing, style and magnitude of fluvial activity within Lyot Crater.

The other side to the story of water in Lyot crater is related to the history of ice across the region. Lyot records both periglacial or ground-ice processes (e.g. potential thermokarst landforms; Glines and Gulick, 2018) and glacial activity (e.g. viscous flow features). This history is evident due to the deposition of ice-rich mantles, as well as the flow of icy material down the slopes of the crater to form viscous flow features. Both are interpreted as being younger than the fluvial activity which formed the channels.

5.9 Conclusions

Detailed mapping of the interior of Lyot crater has been conducted, building upon the grid mapping presented in Chapter 4. This in-depth geomorphological mapping has been conducted to provide context for the fluvial and glacial activity recorded in the crater interior, by revealing the history of the deposits and surface features within this region.

The stratigraphically oldest deposits are the crater units which are composed of the central, inner and outer crater units, and the rugged floor unit. These represent the crater-floor material including the central peak, inner peak ring and crater rim. These units have approximately the same relative age and formed during the Lyot-forming impact.

The pitted units were probably deposited as an airfall deposit as a result of climate cycles influenced by changes in obliquity. Within the centre of the crater, km-scale polygonal features formed by a process that it currently unexplained. In a region to the south-west of the central peak, hydrothermal activity might have led to the formation of the inverted polygonal forms as a result of the deposition of minerals within the fractures to form a “vein” network, or aeolian material was deposited in the fractures and became cemented. The hosting material then gradually became desiccated and sublimed over time as ice was removed, and material was eroded by aeolian abrasion. The channel and associated fan deposits form after deposition of the mantle units and formation of the polygons. Their origins will be discussed, with context provided by the map, in Chapter 6.

Finally, the mantle units were deposited on top of the older, more degraded airfall deposits. These mantling units are the youngest across the region and represent ice and dust deposited during the latest period or periods of net ice deposition as a result of climate change driven by differences in obliquity. During these same periods, snow and ice accumulated and formed the debris-covered glaciers mapped as viscous flow features. The most recent activity involves the incision of gullies into crater units that have been revealed
from beneath mantle units, and the formation of aeolian dune fields, which may still be active today (Widmer and Diniega, 2018).

Mapping has revealed that the fluvial activity which formed the channels occurred after the deposition of the pitted units and, although they are associated with ice-rich deposits, the fluvial channels formed before the viscous flow features, as they are revealed from beneath these features. Further work in the next chapter focusses on these channels and their associated deposits to better understand how they formed.
Chapter 6: Fluvial Landforms in Lyot Crater’s Interior

6.1 Introduction

The work within this chapter builds upon the geomorphological mapping of Chapter 5 and presents the science outcomes of the analysis of the potential fluvial landforms mapped within the crater interior. The two landforms of particular interest are the small channels and their associated fan deposits presented in Chapter 5. The channel and fan assemblage appear morphologically similar to terrestrial river channels with a terminal depositional fan. To better understand if these features could be fluvial in nature a study of their morphology, setting, and hence potential origins has been conducted.

6.1.1 Background

Channels within and around Lyot crater have been the focus of previous studies by Dickson et al. (2009) and Hobley et al. (2014), as discussed in Chapter 2. The main mechanisms proposed by these studies are formation due to the melting of glacial deposits as a result of a local microenvironment resulting from the low elevation of Lyot crater (Dickson et al., 2009), and formation under a regional ice sheet as a result of global climate change or climatic excursions (Hobley et al., 2014). The fan deposits have previously been interpreted as alluvial fans in various stages of degradation (Dickson et al., 2009), or as being the result of the rapid freezing of water released from the margins of an ice sheet (i.e. “aufeis-like” deposits) alongside ice-marginal depositional lobes (Hobley et al., 2014).

Hobley et al. (2014) interpret the channels as forming under a regionally extensive ice sheet due to observations that channels transition to ridges which then transition to a terminal fan deposit. This assemblage is interpreted as being morphologically analogous to subglacial channels transitioning to eskers, followed by the deposition of ice-marginal deposits and/or “aufeis-like” deposits (Figure 6.1). Hobley et al. (2014) also noted convexity within various channel profiles which was interpreted as the incision of the channel against the local gradient – which commonly occurs when a flow is pressurised, such as under a constraining ice sheet. Conversely, Dickson et al. (2009) found that channels follow regional gradients, and suggest that they originate from the glacial features, as some channels appear to have a direct association with them, but formed subaerially.
These two working hypotheses can be tested by studying the channels and their associations in more detail, and by using terrestrial morphological features indicative of glacial processes as analogues. The fan deposits have not previously been studied in-depth other than to note the range of morphologies, average areas, and the modification that they have undergone. This chapter will therefore also present the first in-depth study of these fan deposits, and will present further study of the channels, using the map for context, to test previously presented hypotheses.

6.1.1.1 Glacial Channels

There are a number of different glacial channels which can be broadly separated into categories based upon their location in reference to a current, or former, glacier. These categories include supraglacial, englacial, subglacial and proglacial channels (Menzies, 2002; Benn and Evans, 2010).

Supraglacial and englacial channels result from the surface melting of glacial deposits leading to the incision of channels (Menzies, 2002; Benn and Evans, 2010). On Earth, these channels are common on temperate glaciers and are less common on cold-based glaciers – though they can still occur as a result of limited amounts of meltwater generated by absorption of solar radiation (Hambrey and Glasser, 2012). Sedimentary landforms formed by deposition in these channels can occur by both supraglacial and englacial processes and can become progressively lowered to the surface as a result of the ablation of the glacier (Benn and Evans, 2010). Evidence of englacial processes includes eskers which form as a
result of sediments deposited within ice-walled channels being gradually lowered to the surface as underlying ice melts (Benn and Evans, 2010). Such eskers are heavily disturbed and not as well-preserved as those which form through subglacial processes (Banks et al., 2009; Benn and Evans, 2010). Further evidence which could indicate supraglacial and englacial drainage is kame and kettle topography (Menzies, 2002; Benn and Evans, 2010). This topography is composed of a collection of mounds and ridges (kames) interspaced with pits (kettles) which form as a result of the reworking of debris by supraglacial and englacial drainage systems (Menzies, 2002; Benn and Evans, 2010).

A variety of different types of subglacial channels exist; they can appear morphologically similar to subaerial channels but there are characteristics which can distinguish them. One of the most diagnostic characteristics is that subglacial channels are able to ascend topography, as hydraulic potential is determined by water pressure and elevation, meaning channels don’t necessarily follow bed slope (Shreve, 1972; Benn and Evans, 2010). This leads to channels oblique to topographic gradients with undulating long profiles (Menzies, 2002; Jørgensen and Sanderson, 2006; Benn and Evans, 2010; Galofre et al., 2018). The deposition of material within subglacial channels leads to the formation of eskers which, when the glacier is removed, can be preserved as ridges with undulating long profiles (Shreve, 1972; Menzies, 2002; Benn and Evans, 2010).

Proglacial fluvial activity is often characterised by a large volume of sediment transport when compared to non-glacial activity (Church and Gilbert, 1975; Benn and Evans, 2010). On Earth, this results in the deposition of extensive outwash plains (called sandar) with typically braided streams (Kehew and Lord, 1986; Warburton, 1990; Slaymaker, 2011; Benn and Evans, 2010). When water is released from ice-dammed or proglacial lakes, channels (called spillways) can be carved over a number of (sometimes catastrophic, high intensity) episodes (Figure 6.2; Kehew and Lord, 1986; Benn and Evans, 2010). Such channels have approximately uniform channel widths, high depth-to-width ratios and discrete cutbanks (Kehew and Lord, 1986, 1987). Spillways evolve from anastomosing channels which over time become an increasingly focussed flow within a deeply incised inner channel (Benn and Evans, 2010).
Another type of channel which forms as a result of glaciation are marginal, and sub-marginal, lateral channels which form along, or just beneath, the ice margin as a result of meltwater (Figure 6.3; Greenwood et al., 2007; Syverson and Mickelson, 2009; Benn and Evans, 2010). These channels are particularly well-developed in cold-based glaciation where little meltwater penetrates their frozen beds (Greenwood et al., 2007; Syverson and Michelson, 2009; Benn and Evans, 2010), though care must be taken in using them as evidence of cold-based glaciation (Syverson and Michelson, 2009). These channels are particularly useful as they form parallel to the ice-margin and can therefore record previous extents of ice (e.g. Dyke, 1993; Cofaigh et al., 1999; Greenwood et al., 2007).

**Figure 6.3:** Taken from Figure 9, Syverson and Michelson (2009). Image shows large lateral meltwater channels (indicated by arrows) which have formed parallel to the margin of Burroughs Glaciers in south-east Alaska.
6.1.1.2 Fan Deposits

Previously proposed formation mechanisms for the fan deposits include formation as alluvial fans (Dickson et al., 2009), or “aufeis-like” formation and/or ice-marginal depositional lobes (Hobley et al., 2014). Within this study we include deltas as they are morphologically similar to alluvial fans, with the primary distinction being their formation within a standing body of water.

In terrestrial settings, alluvial fans develop where a sediment-loaded stream emerges from a region in which it was confined in a relatively straight, steep and narrow channel, and transitions into zones where sediment transport capacity decreases, typically as a result of a sudden change in gradient (Figure 6.4; Rachocki, 1981; Blair and McPherson, 1994a, 1994b; National Research Council, 1996). This leads to the deposition of sediment in a fan-shaped landform approximating a cone emanating from a single, well-defined apex (Blair and McPherson, 1994a; National Research Council, 1996). The radial profiles of alluvial fans are typically concave or virtually straight (Blair and McPherson, 1994a, 1994b; National Research Council, 1996). Fans with a long radius tend to have more gentle slopes (Rachocki, 1981; Blair and McPherson, 1994b). Alluvial fans commonly form in arid and semi-arid environments where sediment load is high (National Research Council, 1996). They can also be associated with subglacial channels (Colgan et al., 2003).

Figure 6.4: Taken from Figure 14.1a, Blair and McPherson (1994a). Image shows an example of an alluvial fan from south-western Death Valley, California.
Alluvial fans on Mars are difficult to distinguish from deltas as they are very similar in plan-view morphology. Deltas form where sediment is deposited at the edge of a standing body of water, such as a lake or ocean, due to a decrease in the sediment transport capacity (Seybold et al., 2007). Delta deposits often have channels which shift across the deposit (avulse) leading to multiple locations of deposition (e.g. Stouthamer et al., 2011). Alluvial fans can prograde into standing water forming deltas named ‘fan deltas’ (Nemec and Steel, 1988; Postma, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012). Fan deltas have deeper channels and thicker backfill deposits when compared to alluvial fans (Van Dijk et al., 2012). Experimental results also indicate that alluvial fans prograde further than fan deltas (Van Dijk et al., 2012). Distinguishing between alluvial fans and deltas is therefore implicitly related to identification of the environment in which they formed. A morphological indicator for the involvement of a standing body of water is a fan profile that has a steepened front as a result of the development of foreset slopes influenced by subaqueous deposition (Axelsson, 1967; Nemec, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012).

Hobley et al. (2014) invoke a formation procedure for the fan deposits in Lyot which is similar to the formation of aufeis. Aufeis (also known as “up-ice”, “icings” or “naleds”) are often observed on proglacial floodplains in the High Arctic (e.g. Hodgkins et al., 2004; Wainstein et al., 2008). Aufeis occur where water is released onto the surface and freezes in layers (Kane, 1981; Hodgkins et al., 2004; Yoshikawa et al., 2007; Wainstein et al., 2008). Another possible analogue for the fan deposits is that they formed as ice-marginal depositional fans (Hobley et al., 2014). These fans are typically steeply sloping and composed of outwash material (Colgan et al., 2003). They can be interpreted as equivalent to alluvial fans (Blair and McPherson 1994a, 1994b; Krzyszkowski, 2002).

6.1.2 Research Questions

The work presented in this chapter aims to study the fluvial deposits located in the interior, and on the inner ejecta blanket, of Lyot crater, and to determine if these channels and fan deposits represent a surface expression of recent water sourced from the atmosphere. If these channels do represent such fluvial activity then they could represent one of the most recent periods of channel-forming activity on the surface of Mars. To guide analysis of the fluvial features, a number of research questions were posed which are discussed in detail within this chapter.
6.1.2.1 Where are the channel and fan deposits located?

The first section of this chapter focuses on mapping the centre-lines of all of the channels, as well as demarcating the fan deposits associated with some of them, located within the crater interior and across the inner ejecta blanket. These regions have been selected via preliminary mapping: the results of chapter 4 indicate that the channels are only found in these areas. It is then important to distinguish where these channels and fan deposits are located in relation to the morphological map presented in chapter 5, and the topography of the study area. This provides information as to the timing and potential origins of the fluvial activity.

6.1.2.2 Did the channels form subaerially or under the cover of an ice sheet?

Previous work on the channels has proposed different hypotheses regarding whether the channels form subaerially (Dickson et al., 2009), or under the cover of an ice sheet (Hobley et al., 2014). The line of evidence used for both is slope-related, with Hobley et al. (2014) indicating that there is evidence of the flow of water upslope indicating formation resulting from the pressurised flow of water under an ice sheet, and Dickson et al. (2009) finding that channels follow regional gradients implying that they result from the gravity-driven flow of water downslope. To test these hypotheses I analyse the slope of these channels to see if I also find evidence of flow of water uphill. If such an uphill flow is indicated, the channel is put into the context of the map to check that the elevation values are not the result of superposing material, or taken from inverted sections of channel. Additionally, I will discuss, with the map as context, if there is any evidence of a regionally extensive ice sheet. This will provide additional information as to how the channels have formed.

6.1.2.3 Are channels associated with ice-rich material?

A crucial piece of information to distinguish is the source of the water which formed the channel features. Both Dickson et al. (2009) and Hobley et al. (2014) propose that the melting of ice-rich material is the source of water for the channels. However, Dickson et al. (2009) indicates more specifically that the sources of water are glacial features located within Lyot crater (i.e. viscous flow features). Although the relationship with ice-rich deposits has been proposed, there has not been an in-depth study of whether channels (and by extension fans) are associated with deposits interpreted as ice-rich. Using the geomorphological map created in chapter 5 for context, I analyse whether the channels are sourced from ice-rich deposits, and in particular if those deposits are the viscous flow
features. This provides further information as to the potential source of the water which formed the channels and, by extension, likely the fan deposits.

6.1.2.4 How do the fan deposits form?

Although the gross morphologies of the fans have been briefly described (Dickson et al., 2009; Hobley et al., 2014) and areas of the deposits measured (Hobley et al., 2014), the fan deposits have not previously been analysed in detail, nor have morphological measurements other than area values been taken. This chapter describes the first in-depth study of the fan deposits and assesses the various potential formation mechanisms. Through better understanding of the formation of the fan deposits the origin of the fluvial activity which formed the channels can be better understood.

6.2 Method

6.2.1 Mapping

6.2.1.1 Channels

To investigate the formation hypotheses, I focussed first on locating and mapping channel features and fan deposits across the crater interior and inner ejecta blanket. The channels have previously been mapped within the crater interior as part of the work in chapter 5, however to ensure that all the channels have been mapped for this work I expanded the search area to include the inner ejecta blanket and adopted the grid mapping approach used in chapter 4. The CTX mosaic, which covers the entire area within the inner ejecta blanket, was studied at full resolution. All squares in the grid were examined in this study (as opposed to examining every other square as was done for the general grid mapping discussed in chapter 4). Mapping does not extend beyond the inner ejecta scarp, as previous grid mapping identified that the smaller channels, of interest to this study, are not observed in the outer ejecta blanket.
Figure 6.5: CTX mosaic with grid used for the mapping of channel features and fan deposits overlain. The grid is composed of 20 x 20 kilometre squares which are coloured pink if the channel features and fan deposits have been mapped inside that square. The crater rim and inner ejecta extent are marked as black lines.

All channels and fan deposits in the study area were mapped using CTX data, as this is available for the entire area. Channels, and potential channels, were mapped as two separate classes: “certain”, if the channel morphology is distinctive and a fluvial origin is deemed likely; and “uncertain” if the channel is less distinctive but a fluvial origin is thought possible (Figure 6.6). This is to ensure that all potential channels are captured in the study so that when the general trend of the location of channels across the mapping area is discussed there is no biasing as a result of the preferential preservation of channels in particular areas. The separation of channels into these two classes also ensures that the quantitative measurements are only conducted on channels that are well preserved, allowing better interpretations of their origins.

All channels were digitised down their centre-lines using ArcMap 10.1. The GIS project used a Cassini projection centred on the 30° East meridian. Some channels have discontinuities along their path, in these cases each section was mapped separately to ensure accuracy. A total of 157 certain and 470 uncertain channel sections were mapped.
Figure 6.6: CTX mosaic showing examples of channel features mapped as certain and uncertain. A & C) Example of a channel mapped as ‘Certain’. The channel is a distinctive depressed feature within the landscape. It has a dark infill and is sinuous. The channel has been mapped down its centre-line. B & D) Example of a channel mapped as ‘Uncertain’. The channel is less distinctive and is visible due to a difference of albedo within the channel. It is a less distinctive depression and is less sinuous. The channel has been mapped down its centre-line.
6.2.1.2 Fan Deposits

When a fan-like deposit was observed, the deposit was digitised and a series of attributes describing that fan were added to the shapefile. Additional observations were recorded in an Excel spreadsheet (Appendix B). The attributes recorded in the shapefile include: (i) whether it has a bounding scarp, (ii) if there is an inverted section of channel just before the deposit, (iii) if the deposit is situated in a basin, and (iv) a section for other information (e.g. degradation state, if there are impact craters, if it is covered by mantle etc.). The observations, recorded within a Microsoft Excel spreadsheet, consist of: (i) a short description of the deposit and associated channel morphology, (ii) the type of surface it is associated with, (iii) if it is associated with mantle units, (iv) the CTX image within the mosaic that it is on, and (v) the deposit location. Using these observations the fan deposits were divided into five morphological types. These types are discussed in section 6.4. A total of 23 fan deposits were mapped within the crater interior and inner ejecta blanket.

6.2.2 Channel Measurements

Measurements of channel length, elevation, and average slope were conducted for all “Certain” channel sections within the dataset to look for trends or signatures: for instance do channels occur within a specific elevation range or have a constrained range of mean slopes? These measurements were made using the ArcMap 10.1 ‘Add Surface Information’ tool, with inputs consisting of polylines for the channels and MOLA topography for the elevation information. The key parameters extracted are the channel length, mean elevation and mean slope angle. The output data was then exported so that statistical values could be calculated (e.g. mean, minimum, maximum, first quartile, third quartile, median, standard deviations and skewness values). The complete channel dataset is presented in Appendix C. Alongside these measurements, observations of the various units that the channels are associated with were made.

Sinuosity indexes were calculated for 20 channels randomly selected from the “Certain” channel dataset. To calculate these values, “valley length” polylines were created manually within ArcGIS 10.1 for each channel, recording the general path of the channel without the shorter wavelength deviations (Figure 6.7). The sinuosity index is then calculated using the equation:

\[ \text{Sinuosity} = \frac{\text{Channel Length (m)}}{\text{Valley Length (m)}} \]  
(Eq. 6.1)
A sinuosity of 1 would indicate a channel that flows directly, with no divergence from the “valley path”. The greater the value is above 1, the more the channel deviated from a simple path, and for values of sinuosity above ~1.5 it can be defined as “meandering”.

Figure 6.7: Example of sinuosity measurements conducted on two channels with CTX mosaic as a base map. The estimate “valley length” is created manually and assumes a direct flow so that sinuosity can be estimated.

Additionally, a focussed analysis of channel profiles and relationships was conducted for three channel systems (Figure 6.8). These systems were chosen as they are of particular interest due to unit associations, locations, observed relationships with other channels, or relationships with fan deposits. This detailed analysis included the construction of channel profiles, taking into consideration any breaks in channel continuity. This was achieved through the creation of a point every 200 metres along the channel polyline for which an elevation value was taken using MOLA data. Where available, elevation values were taken using HRSC data, for which points were created every 100 metres. This data was then plotted to create channel profiles. MOLA data was available for all of the channels of interest and is suitable to show the broad trend of the channel slopes. The suitability of MOLA data for this study was tested through the comparison of MOLA data, taken every 100 metres, with HRSC data taken for part of channel 1, to see whether there was a significant difference in the values (Figure 6.9). The measured profiles were than
compared. All graphs, boxplots and histograms were created using R (R Core Team, 2013) within RStudio (RStudio Team, 2015).

Figure 6.8: THEMIS daytime base map displaying the channels selected for more in-depth analysis. These channels were selected to represent different morphological types of channel observed within and around Lyot crater. They also capture a number of fan deposit morphologies and the relationship of the channels with these deposits.

Figure 6.9: Profile 1 showing a comparison of the channel profile created using MOLA data and HRSC data sampled every 100 metres. The profile indicates that MOLA data is sufficient to capture the general trend of the channel slope. Note that the channel is superposed by the smooth mantle unit and a large dune field (marked by blue and orange lines respectively). The dune field in
particular has led to an increase in the measured elevation values for this part of the profile. The convex form of the plot for the channel portion inside the dune field is interpreted as being the result of the measuring of the aeolian surficial materials as opposed to the elevation of the channel.

6.2.3 Fan Deposit Measurements

Once all of the fan deposits were mapped, their perimeter lengths and areas were calculated and added to the attribute table. Additional data were then collected using the ‘Add Z Information’ tool in ArcGIS 10.1. This additional data included (i) the mean elevation and (ii) average slope angle of the MOLA points within the mapped fan deposit shapefile. The full table of fan deposit data is available in Appendix D. These data were used to analyse if the fan deposits are found at a specific elevation range or have particular slope angles. This provided further information to be used in an interpretation of their origins.

Longitudinal profiles for the fan deposits were constructed using the same method as that used to create the channel profiles – though in this case the profile was measured along the steepest straight-line path from the end of the channel (the start of the fan) to the distal edge of the fan. A sample spacing of 50 metres was used rather than a spacing of 200 metres. The profiles provide only a broad representation of the elevation in the region of the fan deposits, due to the coarse resolution of MOLA data compared to their size, so are not suitable for in-depth analysis of the deposits themselves. To create a better representation of the fan deposit profiles, higher resolution data was needed. (Figure 6.10). A HiRISE DTM (Digital Terrain Model) which covers part of one of the fan deposits is available. This data has been used to generate a profile of this fan deposit. HRSC DTM data also has a suitable resolution (though the 100 m/pixel resolution is coarse when compared to HiRISE). Figure 6.10 shows that the HRSC data does not capture the profile of the fan deposit, or a suitable broad representation of the elevation, and is therefore not used for the fan deposit profiles.
6.3 Channel Observations and Results

During the mapping process, the channels were scrutinised for associations and morphological features that could provide information about their formation processes. These qualitative observations are a crucial window into the channel formation process, and provide insight into the potential source of the water which led to their formation.

6.3.1 Channel Observations

The channels are predominantly located within the crater interior. Channels within the inner ejecta blanket are commonly associated with the crater rim deposits; a few channels are observed as far into the ejecta blanket as the inner ejecta scarp (Figure 6.11). Within the interior of Lyot crater the majority of channels are observed originating from higher topography regions, such as the crater rim and inner peak-ring, and terminating in areas of lower topography. Other than the relationship with steeper slopes, the channels are not preferentially located in any particular region of the crater interior – the lack of channels in the south-west region is interpreted as being the result of the superposition of the textured mantle unit in this area. The channels tend to form perpendicular to MOLA-derived contours (Figure 6.11), suggesting that flow followed the steepest path.
Figure 6.11: MOLA topography overlain on a THEMIS daytime base map with contours and mapped channels shown. The locations of the fan deposits are displayed as point data. The crater rim and inner ejecta scarp are seen as black lines.

Channels are commonly observed incising pitted units, though some crater units are also incised by channels (Figure 6.12). Mantle units superpose the channels and thus were emplaced after channel formation. Similarly, the aeolian surficial material and viscous flow features also superpose the channels; this relationship will be discussed in further detail within this chapter. Based upon the superposition of units, the channels therefore formed after deposition of the pitted units, but before the mantle units.
Figure 6.12: THEMIS daytime base map with geomorphological units, contours and channels overlain. Fewer channels are observed in the south-west region, this is likely a result of the superposing textured mantle units in this region. It can also be seen that channels are predominantly located within the pitted units.

Figure 6.13: CTX mosaic image showing a channel which incises the pitted floor unit. The channel is well-defined and contains an infilling unit that could represent material deposited by the flow of
water in the channel. A smooth mantling unit (surrounding the small hills) superposes part of the channel. Where this mantling unit is present, the trace of the channel is not seen.

The larger, well-preserved channels have an infill which is frequently darker than the unit it is incised into. It can be seen in Figure 6.13 that when a mantle unit does overlay a channel then a surface expression of the channel is typically not visible.

Channels have different morphologies that appear somewhat controlled by the region of the crater in which the channels form. In general, larger channels occur near to the crater rim. These channels are generally wider and greater in length than others away from the crater rim. The channel morphologies are also more complex, featuring branching channels with stream orders of 2 and 3 (Figure 6.14), and occasional braiding-like reaches, often at their termination (Figure 6.15). Highly sinuous reaches are observed with morphology similar to regular or transitional channel patterns (Schumm, 1963). Fan deposits are most commonly associated with this channel morphology. Channels in the central area, mainly associated with the inner peak-ring, are commonly smaller, with lower widths and lengths. The channel morphologies are simpler with no branching and common stream orders of only 1 (Figure 6.16). Some channels form in the inner ejecta blanket, though this is less common (Figure 6.17). These channels are often discontinuous: they are covered by later mantling units, and in some cases their morphology appears to have been removed by disruption or deflation of the pitted floor unit. They display no branching, and stream orders above 1 are not observed. Some channels have widths more similar to those found near to the crater rim, and few fan deposits are observed. Channel patterns tend to be straight to transitional in morphology (Schumm, 1963). The more complex channels, and well-defined channels, are associated with the crater rim.
Figure 6.14: CTX mosaic image showing a channel which incises the pitted floor unit. The channel is well-defined and contains an infilling unit that could represent material deposited by the flow of water in the channel. The channel is sinuous to meandering in planform, and contains features interpreted as potential mid-channel bars or small anabranches. There is obvious branching and a stream order of 3. The downslope direction is from the bottom to the top of the image.
Figure 6.15: CTX mosaic image showing a channel which incises the pitted floor unit. The channel is well-defined and contains an infilling unit that could represent material deposited by the flow of water in the channel. The channel displays branching and reforming (anabranching) and perhaps even braiding near its termination. The downslope direction is from the bottom to the top of the image.

Figure 6.16: CTX mosaic image showing a number of small channels which incise the pitted floor unit. The channels are not well-defined, and appear to have a darker infilling unit that could represent material deposited in the channel during, or after, flow. The channel is relatively straight with no branching and a stream order of 1. The downslope direction is from the top to the bottom of the image.
Commonly the crater rim and wall are covered by textured mantle and viscous flow features (Figure 6.12). In several locations, the channels appear to originate from beneath the margins of mantle or viscous flow features. Figure 6.18 shows an example of a channel which is visible at the end of a viscous flow feature. Specifically, this channel appears to originate from a depression at the foot of a viscous flow feature – possibly a mantle-filled impact crater. The channel is small, narrow and branched at the proximal end. The branching might represent two channels as opposed to a single branched channel as there appears to be a small ridge left between the two “branches”. In one location, two channels have been observed which appear to be parallel to one another alongside the textured mantle unit. This could be evidence of a glacial formation if these represent glacier-marginal meltwater channels. Both channels follow the local gradient (Figure 6.19), however, suggesting they were not controlled by the presence of now-removed topographic features such as glaciers, unless they were relatively thin.
Figure 6.18: CTX mosaic image showing a channel which incises the pitted floor unit. The channel is well-defined, though short and narrow. The channel displays possible branching, although there is a small gap between the two “branches”, so this could indicate two channels rather than one branched channel. The channel is relatively straight in form. The downslope direction is from the bottom to the top of the image.
Figure 6.19: CTX mosaic image showing two channels which incise the pitted floor unit. The channels are relatively well-defined and contain an infilling unit that could represent material deposited by the flow of water in the channels. The channels are revealed from beneath the textured mantle unit. At the edge of the mantle unit, an expression of one of the channels can be seen through the unit indicating that it might continue beneath the deposit. The downslope direction is from the top to the bottom of the image.

Observations have been noted of locations where the mantle units have sublimed away, and are near to ice-rich deposits such as the viscous flow features and textured mantle unit. In these locations, the channels can be seen to originate high up, near to the crater rim itself (Figure 6.20). Where textured mantle has not receded to the crater rim, the channel can still be observed in regions where the mantle has been removed (Figure 6.21). Therefore, channels have either been partially formed under the textured mantle and viscous flow features, or the units have been deposited on top of the features later. If there is evidence of the channels cross-cutting high topography rather than diverting around it, or not following what are now the steepest downslope paths, this could indicate that channels formed under the pressure of ice and therefore channels could have formed as subglacial channels, with the viscous flow features as a source of water for the channels.

Figure 6.20: CTX mosaic image showing a channel revealed from underneath a textured mantle unit which drapes the crater wall. The channel is well-defined where the textured mantle does not superpose it. Where the textured mantle has not been sublimed away the channel is less visible. This channel is relatively straight and terminates in a fan deposit. The downslope direction is from the bottom to the top of the image. The crater rim is at the bottom of the image.
Figure 6.21: CTX mosaic image showing a channel revealed from underneath a textured mantle unit which drapes the crater wall. The channel is well-defined where the textured mantle does not superpose it. Where the textured mantle has not been removed, the channel is less visible. This channel is well-defined, meanders and has a dark infill. The downslope direction is from the bottom to the top of the image.

Some of channels within the region between the inner peak-ring and crater rim have inverted sections. Two channels in particular have complex inverted morphologies (Figure 6.22A and Figure 6.22B). These channels are also linked within a complex system including fan deposits (Figure 6.22B) and are associated with aeolian surficial material. The aeolian material might represent the reworking of sand-grade material deposited as a result of fluvial activity (Figure 6.22C). This particular system is one which will be discussed in further detail as a study area in section 6.5.
Figure 6.22: A series of CTX mosaic images of a fluvial system consisting of channels with inverted sections and several fan deposits. A) Inverted channel which terminates in a fan deposit. The inverted section has a complex morphology with indications of multiple channel bodies and therefore multiple depositional episodes. The termination of the channel is associated with aeolian surficial material. The downslope direction is from the bottom to the top of the image. B) A second short inverted channel section which terminates in a fan deposit. The channel has a complex
morphology with evidence of multiple depositional episodes and possibly avulsion where it transitions into a fan deposit. The channel and deposit are associated with aeolian surficial material. **C)** A depressed section of channel overlain by aeolian surficial material. The channel contains infilling material with a “ridge and furrow” texture. The downslope direction is from right to left in the image.

In the central crater region, within the dark pitted unit with depressed polygons, fluvial activity appears to have taken advantage of a pre-existing fracture network (i.e. depressed polygons), thereby modifying the fractures and giving them a more sinuous appearance (Figure 6.23). This region has been interpreted by Glines and Gulick (2018) as thermokarst terrain. In particular the channels which appear to connect pits were interpreted as a beaded lake system.

![Figure 6.23](image.jpg)

**Figure 6.23:** CTX mosaic image showing channels that incise the dark pitted unit region that contains depressed polygons. The channels are relatively well-defined and contain a dark infilling material. The channels are revealed from beneath the textured mantle unit. These channels are at the edge of the depressed polygon region. To the right of the image, fluvial channels appear to have taken advantage of the polygon network. Channels connect depressed pits and could represent a beaded lake system as interpreted by Glines and Gulick (2018). The downslope direction is from the top left to the bottom right of the image.
6.3.2 Channel Measurements

Table 6.1 summarises statistical values of the key parameters of 157 channel sections mapped as “Certain”. For a table of all the data collected for these channels see Appendix C.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mean Elevation across section length (m) [2.s.f]</th>
<th>Length of section (m) [2.s.f.]</th>
<th>Average Slope of section (°) [2.s.f.]</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mean</strong></td>
<td>4300</td>
<td>8200</td>
<td>4.3</td>
</tr>
<tr>
<td><strong>Maximum</strong></td>
<td>3100</td>
<td>63000</td>
<td>15.0</td>
</tr>
<tr>
<td><strong>Minimum</strong></td>
<td>-6600</td>
<td>150</td>
<td>0.46</td>
</tr>
<tr>
<td><strong>Count</strong></td>
<td>157</td>
<td>157</td>
<td>157</td>
</tr>
<tr>
<td><strong>Median</strong></td>
<td>-3900</td>
<td>5000</td>
<td>3.7</td>
</tr>
<tr>
<td><strong>1st Quartile</strong></td>
<td>-5000</td>
<td>2200</td>
<td>1.9</td>
</tr>
<tr>
<td><strong>3rd Quartile</strong></td>
<td>-3500</td>
<td>9200</td>
<td>6.0</td>
</tr>
<tr>
<td><strong>Standard Deviation (SD)</strong></td>
<td>930</td>
<td>10000</td>
<td>3.0</td>
</tr>
<tr>
<td><strong>Skewness (SK)</strong></td>
<td>-0.77</td>
<td>3.20</td>
<td>0.97</td>
</tr>
</tbody>
</table>

Table 6.1: Key parameters summarised for all “Certain” channel sections analysed within the study.

Note that some of the channels included in the study are tributary channels, or discontinuous channel segments and are therefore disproportionately short in length with a low average slope angle. This causes a skew of the length data to the left (i.e. greater positive skew).

The distribution of length, elevation and average slope for the channels is broad, indicted by the interquartile ranges in Figure 6.24. The boxplots show that channels do not preferentially occur at a specific range of elevations within the crater (an interquartile range of more than 1.5 km). Average channel slope values indicate that they most commonly form on slopes of between 2 and 6°. Continuous channel section lengths are generally constrained to values of between 2 and 10 kilometres – but this itself shows these are significant systems – certainly they are larger than almost all Mars gullies (e.g. Malin and Edgett, 2000; Heldmann and Mellon, 2004; Balme et al., 2006) but much shorter than valley systems (e.g. Hynek and Phillips, 2003; Hynek et al., 2010).
Figure 6.24: Boxplots displaying channel length, mean elevation and average slope values. The boxes represent the interquartile range. The bold horizontal line is the median. The whiskers represent the upper and lower quartile. Outliers have been removed.
Figure 6.25 displays values for channel elevation with hypsometric curves for the entire study site and crater interior overlain. These curves were generated from the MOLA topography data by calculating the number of pixels that lie within each elevation range. Values are skewed towards the right indicating that channels might form preferentially in higher elevation areas, above -4200 metres, though this skew is not as significant as that for channel lengths (Table 6.1). The hypsometric curve for the entire study site indicates that a large percentage of the site lies at this elevation (-4200 metres and above generally corresponds to the inner ejecta blanket which makes up a large percentage of the region). A histogram normalised to the percentage area of the study region in each elevation range (Figure 6.26) amplifies the skew of the data and indicates that channels do commonly form at elevations above -4200 metres; this is thought to correspond with channels forming on the crater walls, as opposed to the ejecta blanket.

Figure 6.25: Histogram displaying channel elevation with hypsometric curves for the entire study site and crater interior only overlain. The mean value is marked as a red line.
**Figure 6.26**: Histogram displaying channel elevation normalised to the percentage area of the study site within each elevation range. The mean value is marked as a red line.

Figure 6.27 displays values for the average slope of channels, with overlain curves showing the percentage area of the study site with each slope value for the entire study site and crater interior. These values are skewed to the left which could be a result of data for discontinuous and short sections of channel, that do not have a large variation in slope due to their short lengths, affecting the data. Slopes are generally no greater than 8°. The curve for the entire study site indicates that a large percentage of the study site has low slopes of near to 0°. The curve for the crater interior alone shows that that the peak of 0° is a result of the inner ejecta blanket skewing the results. A histogram normalised to the percentage area of the study region in each slope value (Figure 6.28) amplifies the skew of the data and indicates that there is a distinct cut-off at 8°, indicating that channels do not commonly form on slopes above this value.
**Figure 6.27:** Histogram displaying average channel slope with curves showing the percentage area of the site and crater interior at each slope value overlain. The mean value is marked as a red line.

**Figure 6.28:** Histogram displaying average channel slope normalised to the percentage area of the site each slope value. The mean value is marked as a red line.
Measurements of estimated valley length were used to calculate sinuosity values for 20 channels within the crater interior, as outlined in section 6.2.2. Table 6.2 presents statistical values for the 20 channels used for sinuosity calculations.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Valley Length (m) [2.s.f.]</th>
<th>Channel Length (m) [2.s.f.]</th>
<th>Sinuosity [3.s.f.]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>18000</td>
<td>21000</td>
<td>1.13</td>
</tr>
<tr>
<td>Maximum</td>
<td>54000</td>
<td>63000</td>
<td>1.37</td>
</tr>
<tr>
<td>Minimum</td>
<td>3700</td>
<td>4200</td>
<td>1.03</td>
</tr>
<tr>
<td>Count</td>
<td>20</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Median</td>
<td>13000</td>
<td>14000</td>
<td>1.11</td>
</tr>
<tr>
<td>1st Quartile</td>
<td>8600</td>
<td>9800</td>
<td>1.07</td>
</tr>
<tr>
<td>3rd Quartile</td>
<td>23000</td>
<td>26000</td>
<td>1.14</td>
</tr>
<tr>
<td>Standard Deviation (SD)</td>
<td>13000</td>
<td>15000</td>
<td>0.08</td>
</tr>
<tr>
<td>Skewness (SK)</td>
<td>1.60</td>
<td>1.60</td>
<td>1.84</td>
</tr>
</tbody>
</table>

Table 6.2: Key parameters summarised for 20 “Certain” channels for which estimate valley polylines were manually generated. Sinuosity was calculated as outlined in section 6.2.2.

Sinuosity values for the channels are mainly between 1.0 and 1.2 as shown by the interquartile range in Figure 6.29. These values indicate that channels are commonly ‘straight’ in morphology (sinuosity = ~1.0) – containing minor bends with no regularity – with some that are ‘transitional’ (sinuosity = ~1.2) – containing flat curves which repeat (Schumm, 1963). For a channel to be considered ‘regular’ – containing smoothly-curved ideal meanders – the sinuosity would be ~1.5 (Schumm, 1963). Values of ~1.7 and ~2.1 would be classified as ‘irregular’ – containing low amplitude meanders superimposed on a larger pattern – or ‘tortuous’ – containing meanders that have become deformed with a very irregular pattern – respectively (Schumm, 1963).
6.4 Fan Deposit Observations and Results

As with the channels, fan deposits were scrutinised for associations and morphological features indicative of their origins. A better understanding of how these deposits form could provide insights into the duration of the fluvial activity which formed them, as well as the local environment. In this section, key observations of the fan deposits are outlined, followed by the presentation of the quantitative fan deposit measurements.

6.4.1 Fan Deposit Observations

Fan deposits are located predominantly within the interior of Lyot crater, specifically in the region between the crater rim and inner peak-ring (Figure 6.30). No fan deposits were found within the inner peak ring. These fan deposits generally form within topographic lows (Figure 6.30) and are commonly associated with the larger channels that are hundreds of metres wide, kilometres in length, and which occasionally display branching (Figure 6.31). Seven fan deposits are located outside of the crater rim, though five of these are found within crater rim materials, or are sourced from channels which originate from the outer crater unit. Two fan deposits are located outside of the crater rim materials; both occur at the boundary of the inner ejecta scarp and mark the furthest example of the fan deposits.
Figure 6.30: MOLA topography overlain on a THEMIS daytime base map with contours and the locations of mapped fan deposits displayed as point data. The crater rim and inner ejecta scarp are seen as black lines. Fan deposits occur in the regions of low topography within the crater interior, and relatively low topography within the inner ejecta blanket.

Figure 6.31: CTX mosaic image showing a channel revealed from underneath a textured mantle unit/viscous flow feature that drapes the crater wall. The channel terminates in a small basin as a
fan deposit. The fan deposit appears degraded. The downslope direction is from the bottom-right to the top-left of the image.

All of the fan deposits are associated with channels. Many of these channels transition to an inverted section just prior to the deposit (Figure 6.32). These inverted sections of channel are often contiguous with the fan deposit material. In one section of the crater, pictured in Figure 6.22, there are a number of fan deposits associated with inverted channels/inverted channel sections. These particular inverted sections preserve a complex history of deposition and potential channel migration which will be discussed further in section 6.5 as a study site of interest.

![Figure 6.32](image)

**Figure 6.32:** CTX mosaic image showing a channel revealed from underneath a smooth mantle unit which drapes the inner peak-ring. The channel terminates in a basin with a fan deposit, prior to the fan deposit the channel appears to transition to an inverted section which continues onto the fan deposit. The fan deposit appears fractured and degraded. The downslope direction is from the top-right to the bottom-left of the image.

Many of the fan deposits are overlain by mantling units, which can make surface textures hard to identify (Figure 6.33). Some of the fan deposits have a sudden break in topography at their edge (i.e. a scarp; Figure 6.34). These scarps are one of the defining characteristics of particular fan deposit morphologies that have been identified across Lyot crater. The various morphologies have been characterised as: ‘broad’, ‘complex’,
‘elongate’, ‘fractured’, and ‘other’ (for fan deposits that don’t fully fit into any of the other categories). These morphological types are discussed below.

**Figure 6.33:** CTX mosaic image showing a terminal fan deposit which is superposed by the smooth mantle unit. The channel and fan deposit are visible through this unit, though surface textures are obscured.

**Figure 6.34:** CTX mosaic image showing two channels with the same terminal fan deposit, with a lower lying “exit” channel. This terminal deposit has a scarp which marks its boundary and is overlain by a mantling deposit.
During mapping, observations were recorded within the attribute table attached to the fan deposit shapefile. These include the morphological class assigned and information as to whether or not the fan is (i) associated with a basin, (ii) has an inverted channel section, and (iii) has an obvious steep bounding scarp. The data collected is presented within Table 6.3, all the observations made of the fan deposits are presented in Appendix B.
<table>
<thead>
<tr>
<th>FID</th>
<th>Scarp?</th>
<th>Inverted Channel End?</th>
<th>Depression?</th>
<th>Fan Type</th>
<th>Other Information</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Uncertain</td>
<td>Yes</td>
<td>Yes</td>
<td>Elongate Fan</td>
<td></td>
</tr>
</tbody>
</table>
| 2   | Yes    | Yes                  | Yes         | Fractured Fan         | Impact crater on the surface  
|     |        |                      |             |                       | Formed at a sharp change in topography                                           |
| 3   | Yes    | Yes                  | No          | Broad Fan             | Impact crater on the surface  
|     |        |                      |             |                       | Evidence of avulsion                                                            |
|     |        |                      |             |                       | Linked to other fans and aeolian surficial material                              |
| 4   | No     | Yes                  | Uncertain   | Complex Fan           | Impact crater on the surface  
|     |        |                      |             |                       | Evidence of avulsion                                                            |
|     |        |                      |             |                       | Linked to other fans and aeolian surficial material                              |
| 5   | No     | No                   | No          | Complex Fan           | Linked to other fans and aeolian surficial material                              |
| 6   | No     | Uncertain            | No          | Elongate Fan          | Formed where fan "hits" a channel                                               |
| 7   | Uncertain | Yes                 | Uncertain   | Complex Fan           | Linked to other fans and aeolian surficial material                              |
| 8   | Uncertain | Yes                 | Yes         | Elongate Fan          |                                                                                  |
| 9   | Uncertain | Uncertain            | Uncertain   | Broad Fan             | Covered by mantling unit  
|     |        |                      |             |                       | Next to viscous flow features                                                   |
| 10  | Uncertain | Yes                 | Uncertain   | Broad Fan             |                                                                                  |
| 11  | Uncertain | Yes                 | Yes         | Broad Fan             | Two channels linked to this deposit                                              |
| 12  | Yes    | No                   | Uncertain   | Broad Fan             | Formed at a change in topography                                                |
| 13  | No     | Yes                  | Yes         | Fractured Fan         | Formed at a change in topography                                                |
| 14  | Yes    | No                   | Uncertain   | Elongate Fan          | Large pits on the surface                                                       |
| 15  | Uncertain | Yes                 | No          | Other                 | Possibly a section of inverted channel rather than a fan deposit  
|     |        |                      |             |                       | Superposed by a second fan from a different channel                             |
| 16  | Uncertain | Yes                 | No          | Other                 | Superposed on a second fan from a different channel                             |
| 17  | Uncertain | No                  | Uncertain   | Broad Fan             | Appears degraded/buried                                                         |
| 18  | Uncertain | No                  | Yes         | Elongate Fan          | Appears degraded/buried                                                         |
| 19  | No     | No                   | Yes         | Fractured Fan         | Appears degraded                                                              |
Table 6.3: Observations noted within the attribute table attached to the polygon shapefiles for the fan deposits.

<table>
<thead>
<tr>
<th>Number</th>
<th>Uncertain</th>
<th>Uncertain</th>
<th>Yes</th>
<th>Other</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>20</td>
<td>Uncertain</td>
<td>Yes</td>
<td></td>
<td>Other</td>
<td>Formed at a change in topography</td>
</tr>
<tr>
<td>21</td>
<td>Uncertain</td>
<td>No</td>
<td>No</td>
<td>Elongate Fan</td>
<td>Covered by a mantle unit</td>
</tr>
<tr>
<td>22</td>
<td>Uncertain</td>
<td>Yes</td>
<td>Yes</td>
<td>Other</td>
<td>Formed at a change in topography</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Large pits on the surface</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Appears degraded</td>
</tr>
<tr>
<td>23</td>
<td>No</td>
<td>No</td>
<td>Uncertain</td>
<td>Fractured Fan</td>
<td>Might be completely filling a depression.</td>
</tr>
</tbody>
</table>
6.4.1.1 Broad Fan

Broad fan deposits (Figure 6.35) have a characteristically greater width than length. The deposits appear to have a wide flat surface with one or more, sharp topographic breaks which demarcate its edge. On closer inspection in HiRISE images the surface of the fan is often ‘stippled’ in morphology. Many of these deposits are overlain by mantling units.

![CTX mosaic image showing a channel visible through a smooth mantle with a terminal broad fan deposit. This channel and fan are found within the outer crater unit. The fan deposit is characteristic of the broad fan morphology, with two topographic breaks at the edge and an extent that is wide and relatively flat.](image)

Figure 6.35: CTX mosaic image showing a channel visible through a smooth mantle with a terminal broad fan deposit. This channel and fan are found within the outer crater unit. The fan deposit is characteristic of the broad fan morphology, with two topographic breaks at the edge and an extent that is wide and relatively flat.

6.4.1.2 Complex Fan

Complex fan deposits (Figure 6.36) are found in one location only and are superposed by extensive aeolian surficial material. The fan seems to be formed from the amalgamation of several inverted channels. The deposits are relatively lobate in shape and seem to represent inverted channels that show evidence of channel migration. The morphology of the deposit potentially indicates avulsion, but could instead show both vertical and lateral migration. These channels are discussed in more detail as a specific study area in section 6.5.
Figure 6.36: CTX mosaic image showing a characteristic example of the complex fan morphology. These deposits appear to be an amalgamation of inverted channels and are lobate in shape. They are associated with superposing aeolian surficial material.

6.4.1.3 Elongate Fan

Elongate fan deposits (Figure 6.37) have many characteristics in common with broad fan deposits, with one distinguishing difference, which is that they have a high length/width ratio. Elongate fans have a long, relatively flat surface with stippled morphology (when examined in higher resolution images). The deposits have one, or more, sharp topographic breaks at the edge and are commonly overlain by mantling units.
Figure 6.37: CTX mosaic image showing a terminal elongate fan deposit. This deposit is characteristic of the elongate fan morphology, with its relatively flat surface, topographic break at the margin and high length/width ratio.

6.4.1.4 Fractured Fan

Fractured fans (Figure 6.38) are generally the smallest fan deposits observed within the mapped region. These deposits are circular to ellipsoidal in form with a highly fractured and/or dissected surface.
Figure 6.38: CTX mosaic image showing a terminal fractured fan deposit. This deposit is characteristic of the fractured fan morphology. It has a highly fractured/degraded surface and is ellipsoidal in plan view.

6.4.1.5 Other Fan
**Figure 6.39:** CTX mosaic image showing two terminal fan deposits. These deposits have been characterised as “Other” as they do not clearly fit into any of the previous morphologies. One of the fan deposits superposes the other. The underlying deposit has a greater length/width ratio. They are sourced from two separate channels.

Fan deposits were classified as “Other” if they did not clearly fit within any of the previously described morphologies (Figure 6.39). All of the deposits within this category are located within the inner ejecta blanket. These deposits are small in size which could explain why it is difficult to classify them.

### 6.4.2 Fan Deposit Data

A summary of statistical values for the key parameters gathered from 23 fan deposits is presented in Table 6.4. For a table of all the data collected for these deposits see Appendix D.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Area (km²) [2.s.f.]</th>
<th>Elevation (m) [2.s.f.]</th>
<th>Average Slope (°) [2.s.f.]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>2.8</td>
<td>-4800</td>
<td>2.5</td>
</tr>
<tr>
<td>Maximum</td>
<td>11.0</td>
<td>-3400</td>
<td>7.7</td>
</tr>
<tr>
<td>Minimum</td>
<td>0.12</td>
<td>-6300</td>
<td>0.7</td>
</tr>
<tr>
<td>Count</td>
<td>23</td>
<td>23</td>
<td>23</td>
</tr>
<tr>
<td>Median</td>
<td>1.7</td>
<td>-4500</td>
<td>2.2</td>
</tr>
<tr>
<td>1st Quartile</td>
<td>0.74</td>
<td>-6000</td>
<td>1.4</td>
</tr>
<tr>
<td>3rd Quartile</td>
<td>3.2</td>
<td>-3600</td>
<td>2.8</td>
</tr>
<tr>
<td>Standard Deviation (SD)</td>
<td>3.2</td>
<td>1100</td>
<td>1.6</td>
</tr>
<tr>
<td>Skewness (SK)</td>
<td>1.7</td>
<td>-0.17</td>
<td>2.02</td>
</tr>
</tbody>
</table>

**Table 6.4:** Key parameters summarised for the 23 mapped fan deposits.

The distribution of area and elevation for the fan deposits is broad, indicated by the interquartile ranges in Figure 6.40. The average slope however is more constrained. The boxplots show that these deposits commonly form on average slopes of between 1 and 3° which is less steep than the channel slopes. These results indicate that fan deposits do not preferentially occur at a specific range of elevations and the fan deposit sizes are varied.
Figure 6.41: Figure 6.41 shows the fan deposit area distribution in further detail. It can be seen that the deposit areas are most commonly lower than 1 km² and are therefore generally small, although the mean value is around 3 km². Figure 6.42 shows that the elevations on which the fan deposits are located are centred between the lowest topographies and the highest, with fewer deposits located between -5000 and -4500 metres. This could be indicative of the two interior crater regions in which fan deposits have been observed: the central crater region, and between the crater rim and inner peak-ring.

Figure 6.40: Boxplots displaying fan deposit area, mean elevation and average slope values. The boxes represent the interquartile range. The bold horizontal line is the median. The whiskers represent the upper and lower quartile. Outliers have been removed.
Figure 6.41: Histogram displaying fan deposit area. The mean value is marked as a red line.

Figure 6.42: Histogram displaying fan deposit elevation. The mean value is marked as a red line.
Figure 6.43 shows the average slopes on which the fan deposits form. It can be seen that they tend to form on slopes of between 1 and 3° as indicated by the boxplots above. The mean average slope on which the deposits form is 2.5°. These results indicate that the deposits tend to form on shallow slopes.

![Histogram of Average Fan Deposit Slope](image)

**Figure 6.43**: Histogram displaying average fan deposit slope. The mean value is marked as a red line.

To investigate whether particular fan deposit morphologies have specific characteristics such as smaller average areas, steeper slopes or occurrences at specific elevations, a number of boxplots have been generated. Figure 6.44 shows that fractured fans, and fans marked as other, have the smallest average areas, whereas broad fans have the largest. Broad fans also have the largest quartile range.
Figure 6.44: Boxplots displaying fan deposit area segregated by fan morphology. The boxes represent the interquartile range. The bold horizontal line is the median. The whiskers represent the upper and lower quartile. Outliers have been removed.

Figure 6.45 indicates that two fan morphologies are located in specific elevation bands. Complex fans are found only in the lowest elevation region of the Lyot crater interior. This could be a result of a bias in the data due to complex fans occurring in close proximity to each other. Fans categorised as ‘Other’ are only located in areas of high elevation (above -4000 m), this is because these fans are only found within the inner ejecta blanket, not within the crater interior. Finally, Figure 6.46 shows that fractured fans have high average slopes of above 3°, as well as the highest quartile range. This could be a result of the method by which average slope values were sampled, as the fractured fans have the smallest areas. The other fan morphologies have low average slopes of between 1 and 3°.
Figure 6.45: Boxplots displaying fan deposit elevation segregated by fan morphology. The boxes represent the interquartile range. The bold horizontal line is the median. The whiskers represent the upper and lower quartile. Outliers have been removed.
Figure 6.46: Boxplots displaying average fan deposit slope segregated by fan morphology. The boxes represent the interquartile range. The bold horizontal line is the median. The whiskers represent the upper and lower quartile. Outliers have been removed.

6.5 Analysis

6.5.1 In-depth analysis of fluvial assemblages

To better understand the process by which the channels and fan deposits form three study sites have been selected which represent well developed fluvial assemblages and characteristic channel systems (Figure 6.47). Further observations which build upon section 6.3 were noted, and in-depth analysis was conducted for these sites. There are two main hypotheses currently proposed for the formation of the channels at Lyot crater:

1. Channels were incised as a result of the subaerial flow of water sourced from the melting of ice-rich deposits (Dickson et al., 2009). This is inferred from the morphology of the channels and channel gradients appearing analogous to valleys carved through fluvial processes, combined with their proximity to ice-rich deposits, such as the viscous flow features (Dickson et al., 2009). In this scenario, the channels are indicative of proglacial drainage.
2. Channels were incised underneath a regionally extensive cap of ice and snow which has since been removed from the landscape (Hobley et al., 2014). This is inferred from channels crossing DTM-derived basin divides, and generated channel profiles indicating sections where channels ascend topography (Hobley et al., 2014). Inverted channel sections and fan deposits have been interpreted as eskers, and ice-marginal depositional lobes and/or “aufeis-like” deposits (Hobley et al., 2014).

![MOLA topography overlain on a THEMIS daytime base map with contours and mapped channels shown. The fan deposits are displayed as point data. The crater rim and inner ejecta scarp are seen as black lines. The black boxes display the locations of the three study sites that have been studied in-depth.](image)

**Figure 6.47:** MOLA topography overlain on a THEMIS daytime base map with contours and mapped channels shown. The fan deposits are displayed as point data. The crater rim and inner ejecta scarp are seen as black lines. The black boxes display the locations of the three study sites that have been studied in-depth.

If channels result from the pressurised flow of water underneath an ice sheet, instances of flow against the local gradient should also be found in this study. Therefore, for each study site, a channel profile for each main channel has been generated using the method described in section 6.2. These profiles were then scrutinised for evidence of sections where channels are potentially ascending topography. Where channels might ascend topography, the geomorphological map was scrutinised to check that the profile is generated from the channel, and not from a superposed later deposit, such as aeolian surficial materials or mantling units. Furthermore, the ‘Create steepest path’ tool in ArcGIS
was then used to find the steepest path from the uphill starting position of each channel to see whether these paths correspond to the mapped path of the channels. If there is a close agreement between the two paths, then this indicates that channels follow the steepest local gradient, and argue against the interpretation that the channels flow against the present-day topographic slope.

Further study of the fan deposits involved the analysis of fan deposit profiles to observe if there is a pronounced steepening of slope at the downslope end, which might indicate (partial) formation in a subaqueous environment (Axelsson, 1967; Nemec, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012). There is only one suitable profile available for analysis, which is a broad fan deposit within study site A. Another way in which the fan deposits were studied was through analysis of the contours in each site’s location to see if the deposits form within closed basins. To analyse whether the deposits occur within basins, a 25 metre contour map was generated from MOLA topography data. The contour map was set to a colourmap with reds and oranges representing high elevations and greens representing low elevations. This contour map has been used to look for evidence of enclosed basins in Lyot crater that could have contained pooled water.

### 6.5.1.1 Study Site A

This site encompasses the most extensive fluvial system within Lyot crater (Figure 6.48). It consists of four main channel segments, with a number of adjoining channels originating from the inner peak-ring and crater rim. There are also five fan deposits within the region: three complex fans and two broad fans. Taken together, this assemblage is interpreted as a drainage network with tributary channels joining onto a main channel which flows downslope. The fan deposits could therefore represent alluvial fans or deltas, depending on whether they formed sub-aerially, or in standing bodies of water. There is one fan that appears to be cut by another channel (Fan Deposit 6), which indicates that this channel and fan formed earlier, and is thus connected in space but is not contemporaneous with the other channels.
Figure 6.48: CTX mosaic image displaying the entirety of study site A which represents the most extensive fluvial system within Lyot crater. It consists of four main channel segments, with a number of adjoining channels originating from the inner peak-ring and crater rim, and five fan deposits. The red box within the inset map indicates the extent of the main image.

The main channel segments have been combined within a single profile to see if this channel follows local gradients (Figures 6.49A and B). The profile demonstrates that the channel does generally follow local gradients and has a consistently negative slope, though there is convexity and portion of positive slope visible in the lower section (Figure 6.49C). However, the convex section is not necessarily indicative of uphill flow; instead it matches the area where the channel is superposed by mantling units and aeolian surficial material. This indicates that there has been modification of this landscape after formation of the channel. Furthermore, profiles across the mantling units and aeolian surficial material have given thickness estimates of up to 100 metres, at least partially accounting for this uphill climb. This indicates that the measured elevation values in this region are not representative of the channel floor and the elevations recorded are too high due to the measurement of the elevation of the aeolian deposits.
Figure 6.49: A) CTX image of the highest section of the main channel. This section of channel is sinuous and follows the steepest slope, as indicated by the steepest slope analysis conducted. The channel terminates in a broad fan deposit. There is a second fan linked to a tributary approximately midway up the channel section that is cross-cut by the channel. The red box within the inset map indicates the extent of the main image. B) CTX image of the middle and lower section of the main channel. This section is overlain by a large aeolian deposit and terminates in a complex fan deposit, followed by another shorter inverted channel which terminates in a second complex fan deposit. The red box within the inset map indicates the extent of the main image. C) Channel profile for the main channel within the study site which is segmented into four channel sections. The channel broadly follows the local gradient. A region of convexity occurs within the lower channel section which correlates with superposing mantle units and an aeolian deposit.

One of these fan deposits (Figure 6.50) is covered by a 1 m/pixel DTM available on the HiRISE website (DTEEC_024831_2300_019372_2300_A01) so a profile was generated using this data (Figure 6.51). This profile shows the fan is mainly linear, with a slope of ~0.3° and a sudden steepening in gradient (to ~5°) at the downslope end, indicating the presence of a terminal scarp. The steepened margin of this deposit could represent foreset slopes deposited within a subaqueous environment (Axelsson, 1967; Nemec, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012), and thus that this fan was at least partially deposited within a standing body of water. The approximate depth of this standing body of water can be determined by measuring the height of the terminal scarp (Nemec, 1990; Van Dijk et al., 2012), giving an estimate of 50 metres.
Figure 6.50: CTX image showing fan deposit 1 within the study area. This deposit has been classified as a broad fan. It has a steep scarp at the deposit margin and is lobate in form. The surface of the unit is stippled and pitted indicating degradation. The section nearest to the channel is potentially fractured. The deposit is overlain by a mantling unit.

Profile of Fan Deposit 1 (HiRISE)

Figure 6.51: Profile for fan deposit 1 in the study area created using a HiRISE DTM (DTEEC_024831_2300_019372_2300_A01). The profile has a relatively straight profile with a sudden change in gradient at the end that is representative of a steepened scarp.
Of further interest is the continuation of the channel beneath this fan deposit (Figure 6.52). Such an assemblage could indicate the lowering of water level leading to the formation of a later channel which continued beyond the fan deposit; evidence to support this would include the presence of later incising channels on the fan deposit surface which are not observed. Two other mechanisms are proposed: (i) the lower stretch of channel formed prior to the fan deposit as a result of a change in the flow of water or sediment-load, or, (ii) the channel is an outlet channel formed as a result of the overflow of a standing body of water formed within a basin linked to the fan deposit (Fassett and Head, 2005; Mangold and Ansan, 2006). Figure 6.52 shows that the fan deposit formed at the edge of a depression, with the lower section of channel forming at the lower margin of this depression and connecting the system to a second depressed region which is currently host to aeolian drift material. This evidence supports the second hypothesis, as water from the channel would likely pool in a depressed region and the change in slope would lead to the deposition of material. As the water pools, a subaqueous portion of the fan forms leading to a steepened margin. It is feasible that the pooled water would eventually overtop or breach the lower basin margin leading to the formation of the lower section of channel. This hypothesis also explains the lack of channel between the fan deposit and lower channel section, as this region would be underwater. This hypothesis for channels forming below fan deposits will be discussed further in section 6.6.
**Figure 6.52:** CTX mosaic image showing the broad fan deposit linked to the highest channel section of profile 1a, with the 25 m MOLA contour map overlain. The contours indicate that the fan deposit is at the edge of a depression. The lower channel links the edge of this depression to another depressed region which is host to aeolian material.

**Figure 6.53:** CTX mosaic images showing four fan deposits found in close proximity to each other within the study site. **A)** A complex fan deposit superposed by later aeolian surficial material. The deposit has a rough texture with parallel surface lineations (perpendicular to inferred flow direction). **B)** A complex fan deposit superposed by later aeolian surficial material. The deposit is formed of complex inverted channel systems with possible evidence of channel migration and/or avulsion. The deposit abuts a second channel, so might not really be a fan deposit at all, and could be a complex contributory junction. **C)** A complex fan morphology superposed by later aeolian surficial material. The deposit is linked to a complex inverted channel with possible evidence of avulsion and/or channel migration. This is the lowest, and terminal fan in the system **D)** A broad fan deposit overlain by mantling material. The deposit has a discontinuous margin on the upslope side and terminates with a sudden change in gradient at the downslope end. It is linked to a channel which originates from the crater rim and transitions to an inverted section prior to the deposit.

The deposits visible in Figure 6.53A, B and C, are all classified as complex fans and occur in close proximity to each other. This system of fan deposits is associated with two channels: the main channel, which trends from east to west; and an adjoining channel, sourced from the crater rim, which trends from south to north. All of these complex fans
are superposed by later aeolian surficial material. One of the fan deposits (Figure 6.53A) is difficult to distinguish due to aeolian reworking, and the channel associated with the start of this deposit is not visible. However, the prior visible section is depressed with an infill that appears to have a ripple-like texture. This channel is sinuous but mainly buried by aeolian surficial material. The other two complex fans are linked to channels that are inverted and complex in morphology. There is evidence of multiple depositional episodes and potential channel migration (Figure 6.54). The fan morphology might indicate the amalgamation of a number of channels, or multiple channel forming episodes, leading to the formation of the fan-shape. Alternatively, the deposits might represent a braided system leading to the formation of an outwash fan. A further complexity is the close associations of these deposits with one other. If each fan represented the terminus of the flow, then they show changes in the location and elevations at which deposition occurred. The reason for these different depositional points is currently unknown. The relationship between the deposits is difficult to distinguish due to the overlying aeolian material. A tentative hypothesis, is that this could have occurred as a result of deposition in a standing body of water with a gradually lowering water level, however the topographically higher fans are not visibly incised by later channels (e.g. Fawdon et al., 2018). Another possible hypothesis is that the deposits formed in quick succession such that they essentially form at the same time. In this scenario each deposit formed in a slight depression in the landscape within a larger enclosing basin. The two higher depressions would be significant enough to cause a slight change in velocity leading to the deposition of sediment from the water flow, but not significant enough that a standing body of water forms. Each depression would become infilled and then spill over into the next successively forming fan deposits. This is similar to the “fill and spill” of basins in submarine environments by sediment, which forms fans (Prather et al., 1998; Sinclair, 2000; Beaubouef and Friedmann, 2000; Prather, 2000; Prather, 2003). A similar process in which basins are infilled and then spill over has been proposed on Mars for channels connecting shallow basins in Elysium Planitia (Balme et al., 2011), and basins connected by outflow channels in eastern Valles Marineris (Warner et al., 2013); however no fans have been observed in relation with these landforms.
Figure 6.54: HiRISE image (ESP_042422_2295) a section of the largest ridge present within Lyot crater. The ridge is revealed from beneath a mantling unit and is interpreted as an inverted channel. The channel displays evidence of possible lateral migration and/or layering, as well as a section of possible anabranching.

Figure 6.55 shows that the complex fans form within a large basin, with the terminal fan (Fan Deposit 7), forming at the edge of the lowest contours in this area. The fans form at decreasing elevations from east to west which could provide evidence for a receding body of water. The higher-standing fan deposits might not have visible incising channels as a result of the erosion and abrasion of fan surfaces as the channels were inverted and the aeolian deposits formed. Alternatively, fan deposits 4 and 5 formed separately in different depressions; one or both of these depressions were eventually overtopped leading to the formation of outlet channels which formed the terminal fan deposit in the lowest lying depression. Potential smaller depressed regions are indicated by the contours around fan deposits 4 and 5 which could support the latter hypothesis, though the depressions do not appear significant enough as to form pooled water. This might indicate that the coevolution of these deposits is a possibility; though to my knowledge such a process of sequentially forming fans by basin “fill and spill” has not been observed subaerially elsewhere on Earth or Mars. This assemblage of fans will be discussed further in sections 6.5.2 and 6.6.
Figure 6.5: CTX mosaic image showing the assemblage of complex fan deposits with the 25 m MOLA contour map overlain. The entire location has low elevation values compared to the entire study site. The contours indicate that the fan deposits are forming in a basin, with each deposit forming at decreasing elevation values trending from east to west. The contours near to fan deposits 4 and 5 indicate that they could be associated with slight depressions within the larger basin.
Figure 6.56: A) CTX image of the lowest section of the main channel and one of the main possible tributaries. The tributary channel (Profile 1b) features two sections of channel separated by a broad fan deposit that follows the steepest slope, as indicated by the steepest slope analysis. The red box within the inset map indicates the extent of the main image. B) Channel profile for one of the main tributary channels within the study site, constructed using HRSC data, which is segmented into two channel sections. The channel broadly follows the local gradient. A region of convexity occurs within the lower channel section which correlates with the point at which the channel becomes inverted. This channel terminates in fan deposit 4, discussed above.
Profile 1b (Figure 6.56A) shows the channel profile for a tributary channel that trends from south to north and connects with the complex assemblage of fans discussed previously. This channel is initially depressed but transitions into an inverted section in its lower reaches that is complex in morphology (Figure 6.54). The point at which inversion occurs is visible in the profile as a small convexity, and thus this convexity is not attributed to the upslope flow of water. This channel is separated into two sections: one which occurs above the fan, and another lower reach of channel beyond the fan margin. The channel generally follows local gradients aside from the small convexity. This is further supported by steepest slope analysis which indicates the channel generally follows the steepest slope in the local area. Interestingly, the channel gradient of the lower channel section is less steep than the upper section, potentially implying that the water in this stretch of channel would have a lower velocity and thus lower energy. Low gradient sections on Earth are commonly straight to weakly meandering with an aggradational regime (Grotzinger and Jordan, 2010). Such low gradient sections are commonly associated with floodplains (Jones et al., 1991; Grotzinger and Jordan, 2010), which are often occupied by shallow lakes (Howard, 1996). This could account for the complex morphology and inverted section observed in the lower reach of this channel section (Figure 6.54). Potentially this region represents a floodplain that has been differentially eroded, leaving behind the coarse-grained channel infill (Pain et al., 2007).

The fan in this assemblage (Fan Deposit 3; Figure 6.53D) is another example of a broad fan deposit which appears to have a steeptened margin that has formed half way through a channel, with a gap between the deposit and the next channel section. The contours in Figure 6.57 indicate that the fan deposit has formed at the edge of/within a basin. The lower channel section forms at the lower margin of the basin and continues following local slopes before terminating in the system shown in Figure 6.55. This provides further evidence of potential basin overflow forming discontinuous channel systems marked by broad fans that occur at the edge of/within perched basins.
6.5.1.2 Study Site B

This site focusses on a distinctive channel system involving two branched channels which are linked to a single fan deposit (Figure 6.58A). Below the fan is another section of channel with a possibly braided section of channel at its terminus. The two channels associated with the fan deposit originate from the textured mantle unit superposing inner crater material and are incised into the pitted floor unit. They have inverted sections which transition into the fan deposit. A profile of the channels involved within this system indicates that they generally follow local gradients (Figure 6.58B). This is further confirmed by steepest slope analysis which indicates that the channels incise where the slope is steepest. These channels are not located near to any currently visible viscous flow features.
Figure 6.58: A) CTX mosaic image of study site B. The channel system involves two branched channels terminating in a single broad fan, followed by a further channel terminating with a potentially braided section. The red box within the inset map indicates the extent of the main image. B) Channel profile for the channels within the study site which have been grouped into a single system. This system is composed of two “branches” and a single lower section. The channels broadly follow the local gradient.

The fan deposit (Figure 6.59) is classified as a broad fan. It is unique when compared to other fans at Lyot crater as it has two feeder channels. Both channels transition to short inverted sections which appear contiguous with the deposit. These short inverted sections
indicate differential erosion of the landscape in this region. The fan is overlain by mantling material making it difficult to distinguish surface textures, though the deposit does appear stippled. Qualitatively, the fan appears to terminate with a sharp change in gradient at the downslope end, similar to fan deposit 1. This implies that, at least the margin of the deposit might have been subaqueous, leading to the formation of foreset slopes (Axelsson, 1967; Nemec, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012). Furthermore, there is a terminal section of channel emerging from the mantling deposit which indicates that fluvial activity persisted beyond the margin of the deposit.

Figure 6.59: CTX mosaic image showing the broad fan deposit within study site B. This deposit has two adjoining inverted channel sections which appear contiguous with the deposited material. The deposit is overlain by mantling material.

Figure 6.60 shows that the fan deposit is forming at the edge of a clear depression in the landscape providing evidence for the pooling of water and a partial subaqueous component in this fans formation. The lower channel section forms at the lower margin of this depression and continues to follow local slope into another basin, at which point potential braiding of the terminus is observed. Similarly to the previous section, the channel profile indicates that the gradient of the lowest channel section is less than the higher sections. Possible braiding observed at the terminus could be the result of lower velocity flows leading to the deposition of material forming bars. This assemblage provides
some of the best evidence towards the fans with lower channel reaches (separated by a
gap), representing standing bodies of water that have overflowed.

![Diagram of CTX mosaic image showing a broad fan deposit linked to profile 2, with the 25 m MOLA contour map overlain. The contours indicate that the fan deposit is at the edge of a depression. The lower channel forms at the lower margin of the depression and continues into another basin where braiding of the terminus is observed.]

**Figure 6.60:** CTX mosaic image showing a broad fan deposit linked to profile 2, with the 25 m MOLA contour map overlain. The contours indicate that the fan deposit is at the edge of a depression. The lower channel forms at the lower margin of the depression and continues into another basin where braiding of the terminus is observed.

### 6.5.1.3 Study Site C

This study site focusses on a channel system located at the boundary between the crater rim and inner ejecta blanket (Figure 6.61). The system is composed of a channel with four minor tributaries that terminates in a broad fan deposit. There is a section of channel beneath the fan deposit which could be continuous with the channel and fan deposit. This is not certain, as there is a considerable gap of ~7 kilometres between the fan deposit and this channel section. The channel profile shows that the channel trends downhill, however there are two distinct sections of convexity within the top section of the profile. These sections of convexity are associated with two meander bends in the channel plan view, and the channels appear partially infilled or potentially inverted in this region which could affect sampled MOLA values (especially if the channel is inverted). It is thus uncertain as to whether this convexity is a true representation of the elevation in this region. Similarly to
the other sites, the lower section of channel has a generally lower gradient when compared to the higher channel section.

![Image of study site C](image)

**Figure 6.61: A)** CTX mosaic image of study site C. The channel system involves a single channel with a number of small adjoining channels terminating in a single broad fan. There is a lower section of channel that is morphologically similar but is separated from the fan by a gap. The red box within the inset map indicates the extent of the main image. **B)** Channel profile for the channel and channel section within the study site which have been grouped into a single system. The channel
profile broadly follows the local gradient, however there are two obvious convex sections at the upper end. It is uncertain whether these sections represent true convexity in the profile or are an artefact of the way in which elevations values are extracted.

The fan deposit within this study site is classified as a broad fan (Figure 6.62). It is overlain by a mantling deposit which makes identification of any surface textures difficult. The deposit spreads out directly after the channel leaves the confined space between two hummocks of outer crater unit. This deposition of material in a broad deposit once flow is unconfined is typical of an alluvial fan setting (Rachocki, 1981; Blair and McPherson, 1994a, 1994b; National Research Council, 1996). However, the fan has a distinctive qualitative steepening of gradient at the downslope end. This steepened edge hints at a subaqueous component of this fans formation and thus the fan might be more similar to a fan delta.

![CTX mosaic image showing the broad fan deposit within study site C. This deposit has one adjoining channel and is overlain by mantling material. There is a distinctive scarp at the downslope margin.](image)

**Figure 6.62:** CTX mosaic image showing the broad fan deposit within study site C. This deposit has one adjoining channel and is overlain by mantling material. There is a distinctive scarp at the downslope margin.

The presence of another lower stretch of channel is interesting and is demonstrative of a pattern. Similarly to the other occurrences of such an assemblage, contours at this location indicate that the fan deposit is formed at the edge of basin (Figure 6.63). The lower channel section was formed at the lower margin of this depression and could mark the
overflow of this basin leading to the formation of an outlet channel. This hypothesis will be discussed further in section 6.6.

Figure 6.6: CTX mosaic image showing a broad fan deposit linked to profile 3, with the 25 m MOLA contour map overlain. The contours indicate that the fan deposit is at the edge of a depression. The lower channel forms at the lower margin of the depression.

6.5.2 “Fill” Analysis

For fan deposits seen on Mars to be classed as deltas, the existence of standing bodies of water is required. These standing water bodies do not need to be extensive, just deep enough to cause a change in the transportation capacity of the channels leading to the deposition of material. To analyse where water might naturally pond within the crater a “fill” analysis has been conducted to identify local sinks across the study site. This analysis involved using the fill tool in ArcGIS to calculate a range of different fill levels for closed basins across Lyot crater. The ‘Diff’ tool is then used on the results to create rasters with values that represent theoretical “lakes” if all closed basins were filled. Finally, the raster calculator tool is used to extract values which represent the small basins, allowing them to be displayed as polygons representing different fill values. If these regions overlap with fan deposits and channel systems, this could provide support for the idea that there were deltas terminating the drainage systems within Lyot crater.
The fill analysis is presented in Figure 6.64. This analysis indicates that the majority of the channels across Lyot crater terminate within closed basins. Importantly, all the fan deposits are linked to the edges of closed basins, and many only require shallow fill levels of <100 metres (Figure 6.65A). Where there is a lower channel system beneath the fan deposit, these channels lie between two closed basins and appear to connect these basins (Figure 6.65B). Finally, a large system of channels and fan deposits (including the channels and deposits discussed in study site A) in the southern region of the crater interior is correlated with the deepest potential lake of ~800 metres depth (Figure 6.65C). This system appears to connect three potential standing bodies of water and could represent a significant drainage system. Each of the basins in this system are associated with a dune field, which could indicate enhanced erosion in these locations.

**Figure 6.64:** MOLA topography overlain on a THEMIS daytime basemap with the fill analysis displaying a number of theoretical fill levels overlain. It can be seen that the majority of channels terminate within small basins within the crater interior of varying fill levels.
Figure 6.65: MOLA topography overlain on a THEMIS daytime basemap with the fill analysis displaying a number of theoretical fill levels overlain. A) CTX mosaic image displaying a number of fluvial channels terminating in closed basins, many requiring shallow fill levels of <100 metres. B) CTX mosaic image displaying study sites A and B demonstrating that where there is a lower channel system beneath a fan deposit, the channels lie between two closed basins and appear to connect these basins. C) CTX mosaic image displaying a large system of channels and fan deposits (containing study site A) including the deepest potential lake of ~800 metres depth. The system appears to connect three potential standing bodies of water each of which is associated with a dune field.

6.6 Discussion

6.6.1 Channel Locations and Context

Channels and fan deposits are located predominantly within the crater interior. The channels are incised into the pitted units and in a few locations, possibly also the crater units. There is no evidence of the channels incising any of the mantling units or viscous flow features. Where mantling units are present in the same location as the channels, it is clearly observed that the mantling deposits superpose and obscure these features (e.g. Figures 6.13 and 6.21). Similarly, the fan deposits superpose the pitted units and are in turn superposed by later mantling units (e.g. Figure 6.33). This indicates that the channels and fan deposits formed after deposition of the pitted units, and before deposition of the
mantling units. As the pitted units are interpreted as an older, degraded mantling deposit, this means that enough time must have passed for these older mantle units to degrade before they were incised by the channels, as many of the channels are generally fresh-looking. This indicates that, although the water that formed the channels is likely sourced from the melting of ice-rich deposits within and around Lyot crater, these deposits were likely not melted due to remnant heat from the impact event. This agrees with, and is supported by, dating conducted by Dickson et al. (2009) who quantify a significant time lag between the impact crater formation and formation of the channels using crater counting (~0.8-1.9 Gyr).

Channels typically originate from the topographically higher crater units and trend downslope, terminating in topographically lower regions, many of which form enclosed basins or depressions. There is one particularly large fluvial system, shown in study site A, which extends from the east to the south of the crater within the area between the inner peak-ring and crater rim. This system is located adjacent to a number of viscous flow features. On closer observation of the relationship between the viscous flow features and the channels, it has been observed that, although a few of the channels appear to have a direct relationship with these features (e.g. Figure 6.18), they have not been seen directly originating from the viscous flow features themselves, but rather from mantling deposits that superpose or are located adjacent to them. Another point of note is that, in regions where the mantling deposits are degraded or sublimed away (e.g. Figures 6.20 and 21) the channels are revealed beneath the mantling units and originate from close to the crater rim (e.g. Figure 6.20). This indicates that the channels formed prior to the mantling deposits and the currently visible viscous flow features. This does not mean that viscous flow features are not a potential source of liquid water, as the currently visible deposits probably represent only the most recent period of glaciation, and channels might be associated with previous glacial forms, or a regionally extensive ice sheet that is no longer visible (as indicated by Hobley et al., 2014).

A further point to note is that some of the largest aeolian surficial material deposits seen in Lyot are spatially associated with, and superpose, the most striking channels and fan deposits: those within study site A (e.g. Figure 6.22). This indicates that firstly that enough time since channel and fan formation must have passed for the dune fields to have formed. It might also suggest that the source of the dunes is the fan and channel material – which is likely, given that such deposits commonly contain sand-grade materials.
Overall, I conclude that the channels formed relatively recently – after the formation and degradation of the pitted units – but significantly before deposition of the “young” units in the crater such as icy-mantles, viscous flow features and aeolian deposits.

6.6.2 Channel Relationships with Ice-rich Material

Channels across the region commonly originate from topographically high regions that are often superposed by mantling material and, occasionally, viscous flow features. Not all channels show a direct association with these deposits, and in cases where a direct relationship is implied, this relationship might simply be due to the preferential accumulation of ice-rich material against slopes within Lyot crater, thus superposing the channels. This means that the currently visible deposits at Lyot crater need not be the source of water for the channels. There are no observations that suggest the channels to be coming from fluvial systems inherent to the viscous flow features: there are no observations of channels emerging from portals within viscous flow features, nor of significant channel systems integrated with extant glacier like forms. The conclusion is that the present day viscous flow features have been deposited on top of existing channel systems.

The presence of multiple generations of mantling material, indicated by observations of layering, and the distinction of at least two or three different mantling morphologies indicative of different stages of degradation (i.e. smooth and textured mantle types, and the pitted units, thought to be heavily degraded, older mantling material), implies that there have been multiple episodes of deposition. As the channels are located only within the most degraded mantle deposits, and younger units must have been deposited since their formation, the deposits which provided the source of the water might be significantly degraded or even completely removed and no longer visible on the surface. Regardless, the direct association of these channels with the favoured locations of mantle deposition does imply a relationship with ice-rich material, though whether some of the sources of melt are past viscous flow features (i.e. glacial style deposits) is a more difficult question to answer.

Some of the channels originate from high up on the Lyot crater rim. Channels are also present within the central-most crater interior and inner ejecta blanket, which are not locations that host extant viscous flow features. These locations also show no remnant landforms that could be evidence for having hosted viscous flow features in the past. These observations taken together argue against the channels being associated with glacial
activity in general, although the observations do not rule out some exceptions to this generalisation.

A further distinction to make is whether the source of the water could be from the melting and release of subsurface ice deposited, or present close to the surface, as a result of the impact event that formed Lyot. As these deposits were emplaced during or immediately after the impact event, it follows that they would have been at least partially buried by the later mantling unit in which the channels have formed. In this scenario, the deposits would be present immediately beneath the surface at the time of channel formation. Therefore evidence of a groundwater-related origin of the channels has been looked for. If groundwater seeping from regions of high topography led to the formation of channels, then the presence of significant tributaries would be less likely (Malin and Carr, 1999), and channels would potentially appear more similar to sapping channels with a headward alcove formed due to the undermining and transportation of material from this region downslope (Higgins, 1982; Malin and Edgett, 2000). Channel morphologies within the crater do not appear morphologically similar to gullies. Also, it seems unlikely that significant aquifers would impinge on the surface at the top of the impact crater rim, rather than further down into the crater.

Furthermore, if these deposits are directly linked to the melting of subsurface material there would need to be a heat source. Previous work on channels with a potentially groundwater-related origin on Mars indicates that impact-generated hydrothermal systems related to impact cratering or volcanic activity could provide a source of heat (e.g. Brakenridge et al., 1985; Gulick, 1998), but observations argue against such an origin of the fluvial activity in Lyot. Firstly, the time lag between the formation of the crater and the formation of the channels (~0.8-1.9 Gyr; Dickson et al., 2009) is significant enough that any impact-related hydrothermal activity would have ceased (Abramov and Kring, 2005; Dickson et al., 2009; Osinski et al., 2012); an approximation of the upper limit of hydrothermal activity for a Hellas-sized basin is ~10 Ma after the impact (Abramov and Kring, 2005), which is two orders of magnitude lower than the time lag measured for channel formation at Lyot crater in a basin that is 10 times larger. Secondly, morphological evidence does not indicate the presence of any potential springs or igneous features indicative of potential hydrothermal activity or nearby volcanic activity. Thirdly, the channels are sourced mainly from the impact crater rim area, where extant icy deposits
exist, suggesting that this is a region where atmospherically-derived water is abundant, but this is also likely to be the area where remnant impact heating is least persistent.

Taken together, these points indicate an origin of water from more ancient subsurface ice linked to the impact event is unlikely, and supports a more recent origin linked to the deposition of ice-rich material sourced from the atmosphere.

6.6.3 Did the channels form subaerially or under the cover of an ice sheet?

I infer from the evidence presented here that the channels and fan deposits are linked to ice-rich material, deposited around the impact crater rim area – and probably derived from the atmosphere. However, the nature of this relationship must be further discussed. For example, Hobley et al. (2014) concluded that the channels formed under the cover of a thick ice layer. This conclusion was primarily based upon analysis of slopes across the study site, both within channel profiles and via a drainage analysis. They found evidence of flow upslope and channels crossing basin divides, both compelling lines of evidence for the pressurised flow of water (Menzies, 2002; Jørgensen and Sandersen, 2006; Benn and Evans, 2010; Hobley et al., 2014; Galofre et al., 2018).

Analysis of channel profiles within this study have indicated that the channels generally follow local slopes, but there are regions of convexity within some channel profiles (e.g. Figure 6.49C, 6.56B and 6.59B). It is uncertain if this convexity is a true indication of the upslope flow of water, as elevation values appear to have been affected by the superposition of later units, such as aeolian surficial material, or values are taken from inverted sections and therefore might not be representative of slope at the time of channel formation. Nevertheless, I disagree with the conclusions of Hobley et al. (2014). Qualitatively, I have not observed channels running parallel or oblique to topographic contours, and instead they all run nearly perpendicular to them. I have also noted that the channels flow around topographic obstacles, such as topographically higher crater unit outcrops, and generally follow the steepest slopes. Where there are high channel densities, channels converge on the lowest-lying channel, forming large networks that are broadly representative of a drainage network (e.g. Study Site A). This is supported in part by the fill analysis which indicates that the largest system within the crater terminates into one of the lowest-lying basins – which is also currently host to a large dune field. A further important point to note is that the current topography of the region is not necessarily representative of the topography at the time of channel formation. Since the cessation of fluvial activity, compaction, degradation and differential erosion of the landscape might
have influenced local topography, as indicated by the presence of sublimation pits, aeolian deposits and discontinuities in the channels. Although this effect is likely not significant enough to radically alter the topography (especially as channels are still visible on the surface), it is likely that it will have some influence on the slopes that we now measure – especially for those parts of the system that now appear to be inverted channels (e.g. Figure 6.6). Therefore, although convexity and uphill flow in the channel profiles cannot be completely discredited, there is no evidence that strongly argues against a subaerial formation. Other morphological indicators for channel formation under an ice sheet must be provided.

![Channel Profile 5](image)

**Figure 6.6**: Channel profile for a channel located in the interior of Lyot crater that is affected by degradation which has formed a large pit marked on the profile which has become infilled by mantling material.

To test the ice sheet hypothesis further, morphological indicators for the presence of an extensive ice sheet have been searched for across the mapped area. These include drumlins, mega-flutings, mega-scale glacial lineation, boulder fields, glacial drift and moraines (Kleman, 1992; Benn and Evans, 2010). No evidence for a regionally extensive ice sheet was observed. Hobley et al. (2014) suggest that the ice sheet left no evidence on the units that it would have superposed, and that the only remnants of the previously extensive ice sheet are the superposing mantle units. While it is likely that the later mantling units were once more extensive, this does not necessarily indicate that the fluvial activity occurred beneath these mantling units. I would still expect there to be evidence of erosion and glacial processes associated with an extensive, wet-based ice sheet underneath which channels have formed (Benn and Evans, 2010).
On the other hand, a sub-glacial environment in which few erosional features are observed could be indicative of cold-based glaciation (Sugden, 1978; Dyke, 1993; Atkins et al., 2002; Davis et al., 2006; Marchant and Head, 2007; Benn and Evans, 2010). Cold-based glaciation is also more likely to dominate on Mars due to the cold, hyperarid environment (Marchant and Head, 2007; Head et al., 2009). It should be noted that although such glaciation could preserve previously extant features such as the channels, the formation of sub-glacial channels is less likely, due to the glacier being frozen to its bedrock (Kleman, 1994; Benn and Evans, 2010; Hambrey and Glasser, 2012). Melt associated with such glaciation is typically supraglacial, often forming lateral channels which record the previous extents of the ice (Greenwood et al., 2007; Syverson and Michelson, 2009; Benn and Evans, 2010). Two channels mapped within the Lyot study area have been observed trending roughly parallel to the margin of the textured mantle unit (Figure 6.19). These channels follow local slopes, however, and are therefore not interpreted as lateral meltwater channels. Other features which could provide evidence of a cold-based system are ice-contact ridges and moraines (Head and Marchant, 2003; Hambrey and Glasser, 2012). No moraines or features that could be interpreted as moraines have been observed near to the channels, nor have distinctive parallel ridges been seen which could represent ice-contact ridges.

Although cold-based glaciation is likely to dominate glacial processes on Mars, there is evidence of wet-based glacial processes having occurred in the recent past (e.g. Gallagher and Balme, 2015; Butcher et al., 2017; Conway et al., 2018). These glacial processes do not indicate widespread wet-based glaciation on Mars, rather that wet-based glaciation has occurred on a few occasions in the Amazonian, and perhaps even under a polythermal glacial regime. Within this type of thermal regime, regions of the glacier are of sufficiently high temperature for melting, but most of the glacier is below the pressure-melting point (Ben and Evans, 2010; Hambrey and Glasser, 2012). Such a regime is possibly representative of a location in which the retreat of ice has occurred, with the wet-based area shrinking as ice thins (Hambrey and Glasser, 2012). In areas where a wet-based regime extends to the glacier (or ice sheet) margin, then subglacial channels discharge easily, otherwise subglacial waters will be released when sufficient pressure is built up leading to outburst events or a release of stored water, perhaps forming “aufeis” deposits (Benn and Evans, 2010). In this case, it would be expected that water was stored and then released, forming large erosive meltwater streams or spillways which become constrained to a single
main channel as repeated episodes of meltwater release occur (Benn and Evans, 2010). There are not large amounts of erosion associated with the channels in Lyot crater, as might be expected if they had such a catastrophic origin. Also, this formation mechanism does not explain the associated fan deposits or the contributory network patterns seen in the more well-developed channels. In the case of aufeis formation, the ice deposit would represent a terminus of the flow and could be linked to a sub-glacial channel which ultimately supplies the meltwater for its formation (Wainstein et al., 2008). Potentially fan deposits in Lyot crater could represent aufeis-like deposits, as suggested by Hobley et al. (2014). However, this seems unlikely given the direct association of the deposits with the channels, an assemblage that to my knowledge is not often observed, and the fact that aufeis deposits have not been commonly observed on Mars.

Polythermal glaciers have a high debris load, therefore moraines and features associated with debris deposition caused by ice-retreat might be expected (Hambrey and Glasser, 2012). Distinctive landforms also include kame and kettle topography and lateral meltwater channels (Hambrey and Glasser, 2012), none of which is observed within Lyot crater. A final discussion in relation to the formation hypothesis proposed by Hobley et al. (2014) is whether the inverted channel sections (detailed in section 6.3.1) could be interpreted as eskers.

Eskers appear morphologically similar to inverted channels making distinguishing between the two landforms difficult (Burr et al., 2009; Butcher et al., 2018). Inverted channels form through the differential erosion of fluvial features when channel infill develops a greater resistance then the surrounding terrain either due to the greater resistance of infilling material due to composition or grain-size, due to a process of cementation or induration, or by later infill by erosion material deposits such as lava flows (Pain et al., 2007; Burr et al., 2010). Many inverted channels have been recognised and interpreted on Mars (e.g. Malin and Edgett, 2003; Burr et al., 2010; Davis et al., 2016), as have a very few eskers (e.g. Gallagher and Balme, 2015; Butcher et al., 2016; Butcher et al., 2017). The basis for the distinction between the features is largely related to the morphology of the features and their associations with landforms indicative of the environment of formation.

Inverted “ridges” associated with the channels in Lyot crater are commonly short sections situated immediately prior to the fan deposits (e.g. Figure 6.32). There is only one location in which a large section of inverted ridge is observed (within study site A), where
a depressed channel transitions into a complex inverted section before terminating in a fan deposit. The morphology of the ridge is complex, with evidence of possible anabranching, aggradation or channel migration, as well as layering, all interpreted as being indicative of multiple episodes of deposition. The ridge is contiguous with a distinctive fan classified as complex. The assemblage and morphology of the ridge, alongside the lack of evidence for a subglacial environment, leads me to interpret it as an inverted channel. The correlation of the inverted section with a break in slope, and the presence of aeolian material could indicate that deposition of sand grade materials occurred in this location, followed by the removal of less-resistant material which exposed the more-resistant channel fill, which is now degrading to form sand dunes. It is noted that the gradient of this channel section is low, when compared to the upper reaches (Figure 6.56B). This indicates that the water in this stretch of channel would likely have a lower velocity and thus lower energy. Such low gradient sections are commonly associated with floodplains (Grotzinger and Jordan, 2010). Potentially this channel formed in a floodplain that has been removed due to the differential erosion of fine-grade material, leaving behind the coarse-grained channel infill (Pain et al., 2007), as discussed in section 6.5.1. The short, more common, ridge features are likely inverted channel sections rather than eskers. They form at the margins of the fan deposits which have been shown to occur at the margins of basins. Such topographic features influence the wind flow (Ruel et al., 1998; Greeley et al., 2002; Cardinale et al., 2012) and lead to the development of features such as dune fields in topographic lows such as impact craters (Greeley et al., 2002; Cardinale et al., 2012). Potentially the basins within which fan deposits form have caused enhanced wind erosion, which leads to the development of these small ridge sections which directly adjoin the fans in the basin. Considering the morphological evidence, there is simply not enough evidence to favour a subglacial esker hypothesis.

From the previous discussion, I postulate that a subglacial origin of the channels seems unlikely, based upon morphological evidence, and a subaerial hypothesis is favoured. This activity is more similar to cold-climate channel formation with the source of water being nearby ice deposits, but not necessarily viscous flow features, as no distinct relationship with these features is identified. This does not mean that the ice sheet hypothesis proposed by Hobley et al. (2014) is rejected, but my study does find little evidence to support it compared with the simpler hypothesis of sub-aerial flow.
6.6.4 How do the fan deposits form?

The fan deposits direct association with channels indicate that they involve fluvial processes in their formation, and their common location at the termination of channels, and positive relief, further indicates that they are a feature associated with the deposition of material. Therefore, it is inferred that the fan deposits form from fluvially-transported sediment. Based upon these points three hypotheses are proposed:

1. Fan deposits are related to proglacial activity and form from channels with high-sediment loads as a result of the deposition of this material in a glacial outwash plain (sandar). This material could later be left upstanding by differential erosion.

2. Fan deposits form at changes in topography where the sediment carrying capacity of channels decreases, leading to the deposition of alluvial fans.

3. Fan deposits form at the margins of a standing body of water present within the Lyot crater study region, and therefore represent deltas.

In hypothesis 1, channels would represent proglacial channels with a high sediment-load. Streams associated with outwash plains are typically braided due to the high sediment content, fluctuating discharges and steep gradients (Benn and Evans, 2010). Furthermore, outwash plains are located in close proximity with a glacier, so evidence of a former glacier terminus position in direct association is expected (Benn and Evans, 2010). The typical morphology of a sandar is a fan-shaped deposit with gentle slopes in association with a braided channel system (Kehew and Lord, 1986; Warburton, 1990; Slaymaker, 2011; Benn and Evans, 2010). As has been previously stated, the channels from which the fan deposits have formed are not necessarily associated with the viscous flow features interpreted as glaciers. Furthermore, only one type of fan is potentially associated with braiding (complex fans). The fans themselves are typically morphologically different from outwash plains, as many appear to have a steep scarp at the terminus and have relatively smooth surfaces, or extensively fractured or lineated surfaces, with no evidence of braiding patterns preserved within them. This could be a result of post-depositional modification and the overlaying of later mantling units obscuring surface texture; however, it is difficult to explain how many of these deposits have developed a steep scarp after deposition (i.e. broad and elongate fans), even if invoking inversion of relief (the terminal edge should be the thinnest, and most easily removed by erosion and hence should form the lowest relief
Complex fans are the only morphological type to display potentially braided channels directly associated with the deposit. These fans are found in close proximity to each other in a system that appears representative of a number of different depositional episodes. This could represent the gradual retreat of an ice sheet or glacier, forming different generations of outwash fan. However, there is no other evidence of an ice sheet or glacier having extended to this location. Also the different orientations of the fans and association with two channels which are oriented approximately perpendicular to each other is challenging to explain via this hypothesis. Taking all these observations together it is indicated that a proglacial outwash plain hypothesis is unlikely; there is not enough evidence for a direct association with glacial activity, and the morphology of the fan deposits generally does not conform to that expected in this scenario.

Hypotheses 2 and 3 are difficult to distinguish between as the resulting landforms generally appear morphologically similar. The distinguishing characteristic that separates a delta from an alluvial fan is the environment in which they form. Alluvial fans generally form at breaks in topography where the sediment-carrying potential of a stream or river decreases, often when a confined flow reaches an unconfined plain (Rachocki, 1981; Blair and McPherson, 1994a, 1994b; National Research Council, 1996). Deltas form as a result of a change in flow velocity when water from a stream or river enters a standing body such as a lake or ocean and deposits the material it is transporting (Seybold et al., 2007; Grotzinger and Jordan, 2010). In actuality the identification of whether a deposit is an alluvial fan or delta has further complications, as alluvial fans can be the source of material for a delta or prograde into a standing body of water forming deposits termed ‘fan deltas’ (Nemec and Steel, 1988; Postma, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012). In many ways alluvial fans and deltas have a continuum of form, depending on the environment and how that environment evolves over time. This makes distinguishing alluvial fans from deltas when the deposit is “fossilised” very difficult. A more robust identification should therefore be based upon looking at an assemblage of landforms, as well as the morphology of the deposit, to provide evidence of the environment in which the deposit formed. For instance, if a channel-associated deposit is interpreted as a delta, sufficient evidence of a standing body of water should be provided. Based upon this, fan deposits have been analysed to look at both the deposit morphologies, but also where the deposits occur. For instance, do deposits form in closed basins which might lead to the pooling of water?
Fan deposits in Lyot generally have relatively low slopes, with a mean of 2.5° (Figure 6.43). The fans do not all occur at the same elevations and their surface area has a large range. Taking each fan morphology separately (not including those characterised as ‘other’) it can be seen that the broad fans have the highest areas and fractured fans generally have the lowest, this could be related to formation times or indicative of different volumes of sediment being transported. Fans are not constrained to specific bands of elevation. This is important to note, as if these deposits were deltas and they all formed in a single body of water, they would be expected to exist at approximately the same elevation. Thus, if these deposits are deltas, then they formed in different standing bodies of water – possibly small enclosed depressions in which water ponded due to the gravity-driven process of flow down the steepest slopes (see Figure 6.64).

Broad and elongate fan morphologies have steeper margins – inferred from qualitative observation and a HiRISE fan deposit profile (Figure 6.51). These steeper margins are more characteristic of delta deposits (Axelsson, 1967; Nemec, 1990) and more specifically they are reminiscent of Gilbert-type deltas (Gilbert, 1885; Nemec, 1990; Colella and Prior, 2009; Gobo, 2014). Gilbert-type deltas are characterised by steep subaqueous delta slopes and form where streams/rivers debouch into standing water bodies that are significantly deeper than the feeder channel (Nemec, 1990; Gobo, 2014). Though the deposit margins are steep when compared with the rest of the deposit, the measured slope of the HiRISE fan deposit profile margin is ~5° which is still significantly lower than Gilbert-type deltas, which commonly have steep margins with slopes of up to 20°. Regardless, the steepened front is a morphological indicator for the involvement of pooled water which resulted in the development of foreset slopes (Axelsson, 1967; Nemec, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012). Rather than Gilbert-type delta, the deposits could represent fan deltas formed as a result of alluvial fans prograding into standing bodies of water. The thickness and height of the sloping end is determined by the water depth (Nemec, 1990; Van Dijk et al., 2009; Van Dijk et al., 2012). The measured scarp for fan deposit 9 was ~50 metres which provides a minimum water depth estimate.

Other observations which support pooled water in the formation of these deposits is the association of the fans with depressions in the landscape, and the occurrence of further sections of channel downslope of some of them (e.g. Study Sites A, B and C). These basins represent a topographic low which would be infilled by water from the channels forming small pooled “lakes”. In this scenario, the lower sections of channel beneath the fans could
represent outlet channels formed when the standing body of water, in which the fan deposit partially formed, overflowed (Fassett and Head, 2005; Mangold and Ansan, 2006). In this case, the channels could represent brief flood events (Mangold and Ansan, 2006). This hypothesis of pooled water in basins which are overtopped also explains the lack of channel between the fan deposit and lower channel section, as this region would be subaqueous and therefore it is unlikely that channels would be incised here. It also means that the channels formed before the fan stopped forming, which agrees with the lack of later incised channels on top of the fans. The fill analysis presented in Figure 6.6 further supports this by demonstrating that closed basins are associated with the fan deposits such that the lower channel sections correspond to the topographically low edges of these basins. This is a compelling piece of evidence for the lake and delta hypothesis and again highlights that the history of these fan deposits is complex.

Another type of fan deposit that is located within a large basin within the crater interior is the complex fan morphology; potentially the most striking in terms of its morphological similarity to terrestrial deltas. Complex fans have a distinctive association with inverted channels and, in particular, fan deposit 4 appears to preserve evidence of channel avulsion and vertical migration. Avulsion is a principle process in the formation of fluvial-deltaic successions (Stouthamer et al., 2000). Fan deposits of this morphology are found at the terminus of the largest channel system within an area picked out in my fill analysis as a potential location for a standing body of water. Furthermore, each fan could be associated with a different water level for this theoretical lake, which could account for their close association to one another. In this scenario, fans represent different generations of delta formed as a body of water dries up. Unfortunately, the fans are associated with an extensive dune field which has hidden any further evidence to distinguish if a standing body of water was present in this depression. An issue with this hypothesis is the lack of incising channels present on top of the topographically higher deposits. If these deposits formed first then it would be expected that later water flow would incise them forming late stage channels which connect with the lower delta formed at the new water level (e.g. Fawdon et al., 2018). Another potential mechanism by which this assemblage could form is a process involving the “fill and spill” of smaller depressions present within the larger basin discussed in section 6.5.1. In this scenario, smaller basins result in the deposition of some of the sediment-load but are not deep enough to lead to pooled “lakes”. This could lead to the coevolution of fan deposits. Although this is a possibility it seems more likely that the
simpler mechanism by which deposits form at the margins of a receding body of water is more likely. Evidence of later incising channels could have been removed as a result of differential erosion, the reworking of deposits to form aeolian drift material and/or the superposition of the aeolian material.

Fractured fans are potentially the only fan morphology that is more characteristic of an alluvial fan than a delta (Figure 6.67). These deposits do not typically have a scarp, are often found at a sharp change in topography, and form in regions where flow becomes unconfined. They also have the steepest average slope angle when compared to the other fan deposits. The lineated surface is likely the result of post-emplacement degradation and fracturing. These deposits are interpreted as alluvial fans.

Figure 6.67: CTX mosaic image of a fractured fan deposit that appears morphologically similar to an alluvial fan.

In conclusion, the fan deposits within Lyot crater are enigmatic and record a complex history of water. Broad and elongate fans could be analogous to fan deltas, indicated by their steepened margins and locations at the edges of depressions. Lower lying channels beneath a number of these deposits represent potential outlet channels formed as a result of the overflow or breaching of the margins of these depressions. Complex fans provide perhaps the best example of traditional deltas within Lyot crater. The locations of these fans, and a flood analysis, indicate that they could have formed at the margin of a receding
water body, with fans representing the different water levels over time. Fractured fans likely represent alluvial fan deposits and do not involve a standing body of water in their formation. This therefore indicates that the majority of deposits within Lyot crater are alluvial fans, or fan deltas (alluvial fans that have prograded into standing water). Such deposits are commonly associated with steep mountain streams (Blissenbach, 1954; Rachocki, 1981; Blair and McPherson, 1994a, 1994b; National Research Council, 1996). The average channel slope in the Lyot crater study site is steep, with slopes of between 2 and 6° (mean = 4.3°), indicating that they are most analogous to mountain streams. The fans have lower gradients of between 1 and 3° (mean = 2.5°). This provides further evidence for an alluvial fan environment.

An issue with the hypotheses presented are that the broad, elongate and complex fan morphologies require standing bodies of water. It is contentious to argue for water bodies of this depth within the Amazonian, as Mars environment is interpreted as cold and hyperarid (Marchant and Head, 2007). Some previous studies have indicated the presence of lakes within the Amazonian (e.g. Scott and Chapmann, 1991; Grin and Cabrol, 1997; Soare et al., 2008; Orosei et al., 2018), but these lakes are interpreted, or observed, as ice-covered (e.g. Scott and Chapmann, 1991; Grin and Cabrol, 1997; Orosei et al., 2018) or result from freeze-thaw processes (e.g. Soare et al., 2008). Furthermore, other than these deposits occurring within basins, there is no other evidence of lake-related landforms such as shoreline terraces, though these could be covered by the extensive mantling often found in these depressed regions. Although Amazonian lakes are not common the morphological evidence present in the Lyot study site indicate that the subaerial flow of water has occurred, and it is expected that such gravity-driven flows of water would pool within local depressions. The presence of fan deposits associated with depressions in the landscape, and the observation of lower channel sections that appear to represent outlet channels provide evidence that such pooled bodies of water were present. Therefore, this section concludes that Lyot crater has contained small standing bodies of water within the mid/late Amazonian, as indicated by the crater counting ages for the channels (Dickson et al., 2009; Hobley et al., 2014).

6.7 Conclusions

Fluvial landforms within and around Lyot crater were studied using qualitative morphological observations and quantitative analysis. Using the data collected, the current hypotheses proposed for the formation of the channels and associated fan deposits were
tested. The channels present at Lyot crater are amongst the youngest such channels visible on Mars (Fassett et al., 2010). They are kilometres long and have mean slopes of ~4.3° indicating that they are analogous to mountain streams. Channels are commonly unbranched, though branched channels are observed, largely in the region between the inner peak-ring and crater rim. Twenty-three of the channels terminate as fan deposits. These deposits are split into four main categories based upon their morphological characteristics.

Using the geomorphological map for context, the channels at Lyot crater are seen to originate from the locations of ice-rich units. These units are not the viscous flow features, but rather the mantling units. However, the existing mantling units are probably not the source of meltwater for the formation of these channels. This is because the mantling units superpose the channels, and the channels are revealed from beneath this unit rather than being sourced from it. The degradation state of these superposing mantling units imply that they are likely younger and therefore are not the source of the meltwater. The origin of channels from the locations of these features does however indicate that there could be a genetic link to regions of the crater which favour the deposition of ice-rich material, and therefore a previously deposited unit could be the source of the meltwater.

Quantitative measurements of the slope of channels indicate that there could be regions of convexity within the profiles of some of the channels, as demonstrated by Hobley et al. (2014), which could support a subglacial origin. However, this convexity cannot be relied upon as the basis for the rejection of a subaerial formation hypothesis. This is because the convexity could be the result of the superposition of later units (such as is likely the case within channel profile 1), or the values are taken from inverted channels which do not necessarily represent the slope at the time of channel formation. I searched for evidence of an ice sheet across the region to support the hypothesis of a subglacial origin, but no such evidence was found. As the channels within the region qualitatively follow local slopes and divert around topographic obstacles, a simpler subaerial hypothesis is favoured.

The origin of the fan deposits is more difficult to explain. Morphologically they resemble deltas or alluvial fans. In particular, fractured fans are proposed to represent alluvial fans formed where a channel reaches a sudden break in topography and becomes unconstrained. However, complex, broad and elongate fan morphologies are more challenging to classify. Evidence of fan formation within basins, and their associations with
lower channel sections which could represent outlet channels, indicate that pooled water could be involved in their formation. Complex fans morphologically resemble deltas, and the broad and elongate fans have steepened margins indicative of subaqueous formation. I suggest that these fans are morphologically more similar to deltas and fan deltas respectively. This indicates that standing bodies of water were present within Lyot crater during the mid/late Amazonian. However, explaining the presence of standing bodies of water during the Amazonian needs further work focussing on the stability of water within Lyot crater.

Taken together, this work indicates that the fluvial channels and fan deposits in Lyot crater are the result of the subaerial gravity-driven flow of water. The fan deposits indicate further complexity as they could represent the presence of ponded water. In particular, a complex system of channels and fan deposits within study site A could provide evidence for a receding standing body of water. The source of the meltwater which formed these landforms is suggested to be ice-rich airfall material. Both the channels and this airfall deposit were later covered by younger mantle units leading to preservation of the landforms and landscapes. Later erosion and sublimation led to the (partial) exhumation of these channels and fan deposits. Therefore, channels and fan deposits represent recent water sourced from the atmosphere.

The cause of the melting of ice deposits has not been explored within this chapter. Similarly to Dickson et al. (2009) I would propose that a combination of climate change linked to obliquity cycles and the unique low elevation of Lyot crater has led to the increased stability of liquid water, allowing it to flow without evaporating or boiling as rapidly as elsewhere on Mars. There is a big issue with this hypothesis, as it does not explain the higher elevation channels present within the inner ejecta blanket. Therefore an important avenue of further work would be the modelling of atmospheric conditions within and around Lyot crater to look at the stability of liquid water here under different inferred past climate regimes.
Chapter 7: Clastic Polygonal Landforms at Lyot Crater

7.1 Introduction

This chapter presents work on the clastic polygonal landforms, the distribution of which was mapped using grid mapping and point data mapping, as presented in Chapter 4. The distribution of these landforms demonstrates a distinct relationship with the outer ejecta blanket, with the highest concentration of clastic polygons being located within a specific band at 200-350 km from the crater rim within the outer ejecta blanket. This distribution strongly suggests that these features are linked to the impact event that formed the crater. As polygonal features on Earth are generally formed as a result of the freezing and thawing of ground ice, this could indicate that there was a layer of ice in this region of the Lyot study area. Such a region could represent the deposition of an ice-rich layer penetrated and then deposited within the ejecta blanket during the impact event.

The following chapter, beginning at Section 7.2, presents work that has been published in 2018 within the journal ‘Icarus’ as a paper entitled: ‘Clastic polygonal networks around Lyot Crater, Mars: possible formation mechanisms from morphometric analysis’. This paper is presented in its entirety and details a study of the clastic polygonal landforms, comparing them to a series of morphological analogues. This investigation uses grid mapping results to identify locations where 0.25m/pixel HiRISE data overlap with the known locations of clastic polygonal landforms.

I conceived the study, which included coding a new ArcGIS tool, and conducted all of the measurements and analysis as presented in this chapter. I would like to thank my co-authors for providing general editorial and intellectual support in the creation of this paper. In addition, I want to recognise the elements of the paper that were contributed by my co-authors: Matthew Balme contributed fieldwork photographs of sorted stone circles from Svalbard, Norway, Earth; Alex Barrett provided data on potential sorted patterned ground located around Lomonosov Crater, Mars, and sorted patterned ground from northern Iceland, Earth; and, Susan Conway provided fieldwork photographs of sorted patterned ground from western Iceland, Earth. Reviews of the paper were provided by Jay Dickson and Ernst Hauber, their excellent reviews improved the paper manuscript without changing the scientific content.
7.2 Clastic Polygonal Networks Around Lyot Crater, Mars: Possible Formation Mechanisms from Morphometric Analysis

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7.2.1 Abstract

Polygonal networks of patterned ground are a common feature in cold-climate environments. They can form through the thermal contraction of ice-cemented sediment (i.e. formed from fractures), or the freezing and thawing of ground ice (i.e. formed by patterns of clasts, or ground deformation). The characteristics of these landforms provide information about environmental conditions. Analogous polygonal forms have been observed on Mars leading to inferences about environmental conditions. We have identified clastic polygonal features located around Lyot crater, Mars (50°N, 30°E). These polygons are unusually large (> 100 m diameter) compared to terrestrial clastic polygons, and contain very large clasts, some of which are up to 15 metres in diameter. The polygons are distributed in a wide arc around the eastern side of Lyot crater, at a consistent distance from the crater rim. Using high-resolution imaging data, we digitised these features to extract morphological information. These data are compared to existing terrestrial and Martian polygon data to look for similarities and differences and to inform hypotheses concerning possible formation mechanisms. Our results show the clastic polygons do not have any morphometric features that indicate they are similar to terrestrial sorted, clastic polygons formed by freeze-thaw processes. They are too large, do not show the expected variation in form with slope, and have clasts that do not scale in size with polygon diameter. However, the clastic networks are similar in network morphology to thermal contraction cracks, and there is a potential direct Martian analogue in a sub-type of thermal contraction polygons located in Utopia Planitia. Based upon our observations, we reject the hypothesis that polygons located around Lyot formed as freeze-thaw polygons and instead an alternative mechanism is put forward: they result from the infilling of earlier thermal contraction cracks by wind-blown material, which then became compressed and/or cemented resulting in a resistant fill. Erosion then leads to preservation of these polygons...
in positive relief, while later weathering results in the fracturing of the fill material to form angular clasts. These results suggest that there was an extensive area of ice-rich terrain, the extent of which is linked to ejecta from Lyot crater.

7.2.2 Introduction

Terrestrial polygonal networks of centimetre- to decametre-scale patterned ground are common in cold-climate regions. They form by the thermal contraction of ice-cemented soils, in the case of fracture patterns, and/or the freezing and thawing of ground ice, in the case of patterned ground (e.g. Lachenbruch, 1962; Kessler and Werner, 2003). Patterned ground includes sorted patterned ground – frequently observed in periglacial environments, and thought to form through a combination of processes including frost heave and the upfreezing of clasts (Washburn, 1956; Feuillet et al., 2012) – and thermal contraction crack polygons, including various subtypes such as ‘ice-wedge’, ‘sand-wedge’, ‘composite-wedge’ and ‘sublimation’ (Marchant et al., 2002). Polygonal features have also been observed to form through the dehydration of volatile-rich material generally in arid conditions – termed desiccation polygons or desiccation cracks (Neal et al., 1968) – and through the polygonal weathering of exposed surfaces of boulders and rock outcrops (Williams and Robinson, 1989). Due to the large range of potential formation mechanisms it is important to pinpoint characteristics unique to each polygon type to aid with identification, this is particularly key for their use as morphological (or perhaps process) analogues for features observed on Mars.

On Mars, polygonal surface features have been observed that range in diameter from metres to tens of kilometres (e.g. Pechmann, 1980; Seibert and Kargel, 2001; Mangold, 2005; Morgenstern et al., 2007; Soare et al., 2008; Lefort et al., 2009; Levy et al. 2011). Systematic study of these landforms and comparison with terrestrial analogues can help gain information into the mechanism by which they formed, and so gain insight into past and present environmental conditions. We have identified polygonal clast-bounded networks around Lyot crater, Mars. These polygons are enigmatic in that the clasts that demarcate the polygon sides are up to 15 metres across, with an average polygon diameter of 130 metres. This is significantly larger than morphologically similar polygons observed on Earth or on Mars (e.g., Figure 7.1) which are found with maximum diameters of tens of metres (Washburn, 1956; Balme et al., 2009; Treml et al., 2010; Feuillet et al., 2012; Soare et al., 2016). Additionally, clastic polygons of this morphology and scale are – to our knowledge – unique to the ejecta blanket located around Lyot and so are of particular
interest. Their distinctive morphology and location implies that there is a unique material and/or process leading to their formation. Thus, a better understanding of these features could provide useful information about the environment around Lyot, as well as the material that they are composed of.

Figure 7.1: A) HiRISE (ESP_016985_2315) image of part of a clastic polygonal network observed to the north-east of Lyot crater within the outer ejecta blanket. Image credit: NASA/JPL/University of Arizona. B) An example of morphologically similar terrestrial polygonal features found on Tindastóll Plateau, Northern Iceland. The scale is a 25 cm square with 5 cm markers. Image credit: Alex Barrett. C) HiRISE (PSP_004072_1845) image of possible sorted patterned ground located in the Elysium Planitia region of Mars (Balme et al., 2009). Image credit: NASA/JPL/University of Arizona. D) HiRISE (PSP_005597_1250) image of possible sorted patterned ground in the Argyre region of Mars (Soare et al., 2016). Image credit: NASA/JPL/University of Arizona.

The primary aim of this paper is to present the first in-depth study of these clastic polygonal features using both qualitative observations and quantitative morphometric measurements derived from high-resolution remote sensing data. Secondly, these data will be compared to both terrestrial and Martian polygon datasets collected from other studies in order to assess possible formation mechanisms and, finally, to infer a working hypothesis for their origin.
7.2.2.1 Lyot Study Area

Lyot Crater (50°N, 30°E) is a ~215 km diameter, late-Hesperian-aged impact crater located north of Deuteronilus Mensae and immediately to the north of the dichotomy boundary. Lyot Crater exhibits the lowest points of elevation in the northern hemisphere with a maximum depth of ~3 km in the crater interior (~7 km below datum). It has a central peak within an inner peak ring, and an extensive ejecta blanket composed of hummocky outer ejecta extending to ~2.5 crater radii from the crater rim, and smoother, more continuous inner ejecta with a marginal scarp which extends to ~1 crater radius from the crater rim (Figure 7.2). The ejecta blanket is not well-preserved in the south and southwest due to superposition by deposits of the Deuteronilus Mensae region.

![Lyot Crater Study Area](image)

**Figure 7.2:** Lyot crater study area displayed using colourised Mars Orbiter Laser Altimeter (MOLA; Zuber et al., 1992) topographic data overlain on a MOLA hillshade. The crater peak, inner peak ring, crater rim, inner ejecta scarp and outer ejecta extent are indicated by black lines. Regions of clastic polygonal features are marked as white circles. Image credit: MOLA Science Team.
The impact event which formed Lyot crater is estimated to be late-Hesperian to early-Archean in age (~1.6 – 3.4 Ga; Greeley and Guest, 1987; Werner, 2008; Dickson et al., 2009). Large braided channels, extending >300 km beyond the ejecta margins to the north, west and east of Lyot, are suggested to be the result of groundwater release during the impact event (Harrison et al., 2010). There are also numerous small channels present within the crater interior and inner ejecta blanket that are attributed to more recent fluvial activity, possibly associated with obliquity-driven climate cycles (Dickson et al., 2009; Fassett et al., 2010; Hobley et al., 2014).

Landforms and landscapes morphologically similar to those formed by glacial and/or periglacial processes have also been located in and around Lyot crater. These include Viscous Flow Features (VFFs) thought to be analogous to glacial landforms on Earth (Dickson et al., 2009; Balme et al., 2013b; Hobley et al., 2014), mantling deposits thought to be formed by a dusty, ice-rich material (Hobley et al., 2014), and polygonal networks, as described above. Thus, it appears that the surface in and around Lyot crater has experienced the action of both ancient water sourced from underground, and recent water, probably sourced from the atmosphere (Dickson et al., 2009; Harrison et al., 2010; Hobley et al., 2014).

The clastic polygonal landforms (Figures 1 and 2) are located only in the outer ejecta within a band ranging from the north to south-east. The polygons are of particular interest as they are much larger than most terrestrial clastic polygons and those heretofore reported on Mars (e.g. Balme et al., 2009; Gallagher et al., 2011; Soare et al., 2016), and are located in a very specific area: this pattern could indicate a genetic link between their formation and the Lyot-forming impact event.

7.2.3 Terrestrial Polygon Types

There is a large variety of processes that result in the production of polygonal features on Earth at a range of scales. The most widespread decametre-scale types include sorted patterned ground, thermal contraction cracks and desiccation polygons. These polygonal landforms generally do not exceed polygon diameters of ~300 metres (Neal et al., 1968). In this section, we detail the morphological characteristics and possible formation mechanisms associated with each polygon type.

7.2.3.1 Sorted Patterned Ground

Sorted patterned ground (e.g., Figures 3 and 4) is composed of material sorted into coarse and fine domains which form distinctive geometric shapes including circles,
polygons and stripes (e.g., Washburn, 1956; Kessler and Werner, 2003; Treml et al., 2010; Feuillet et al., 2012). Patterned ground is most commonly found in cold-climate conditions where repeated freezing and thawing cycles occur, but the exact formation mechanism is still not well understood (Washburn, 1956; Treml et al., 2010; Feuillet et al., 2012).
Figure 7.3: A) Sorted stone circles from Brøgger Peninsular on the west coast of Spitsbergen, Svalbard, Norway. The pole is approximately 1.25 metres tall. Image credit: Matt Balme. B) Sorted stone polygons from the western side of Hafnarfjalt, W. Iceland. Image credit: Susan Conway.

Frost heave, a process by which clasts may be moved to the surface, is thought to be important in the development of sorted patterned ground. There are two proposed mechanisms by which heaving occurs (Washburn, 1956; MacKay, 1984; French, 2007). The first mechanism, termed “frost-pull”, involves the growth of ice lenses by downward freezing. This ‘grips’ the clast and the overall heave of the ground leads to it being moved upwards (MacKay, 1984; French, 2007). The second mechanism, termed “frost-push”, leads to the clast being forced upwards by the forming of ice beneath it due to the greater thermal conductivity of the clast compared to surrounding material (MacKay, 1984; French, 2007). Frost heave also leads to the expansion of soil perpendicular to the freezing front causing surface clasts to migrate to clastic borders (Kessler and Werner, 2003). Over time and repeated freeze-thaw cycles, this leads to the separation of fine material from coarse material leading to raised and potentially imbricated borders (Dahl, 1966; Kessler and Werner, 2003; Soare et al., 2016). Another recognisable feature of this process is the tilting of stones due to differential heave at the top and bottom of the clast (French, 2007).

The transition between different geometric sorted forms is the result of a variety of factors including slope gradient and lateral frost heave (Kessler and Werner, 2003; French, 2007; Feuillet et al., 2012). Of particular interest is the transition from circles and polygons to stripes. Stripes (Figure 7.4) are composed of parallel lines of alternating coarse and fine domains oriented down slope (Washburn, 1956). Sorted polygons merge into stripes through a transition gradient of 3° to 7° (Washburn, 1956). According to Goldthwait (1976), polygons and nets form on slopes of 2° to 4°, ellipses form on 3° to 6° slopes and stripes occur on slopes of 4° to 11°.
Sorted polygons generally range in diameter from 1 to 3 metres, with maximum diameters of 10 metres (Washburn, 1956; Treml, 2010; Feuillet et al., 2012). It has been suggested that the larger polygonal structures (>3 metres in diameter) have a polygenetic origin in which pre-existing desiccation or thermal contraction cracks are exploited (French, 2007; Treml et al., 2010). Due to the presence of clasts at the polygon boundaries, polygons of this type are generally denoted as “low-centred”, meaning that the polygon edges are topographically higher than the polygon centre. Another indicator of this polygon type is that the size of the clasts within the polygon borders increases with the size of the polygon, as has been observed on Earth (Washburn, 1956; Goldthwait, 1976; Bertran et al., 2010). Goldthwait (1976), for example, suggests a ratio between the polygon diameter and mean clast size of between 1:5 and 1:10.

7.2.3.2 Thermal Contraction Crack Polygons

Thermal contraction cracks (Figure 7.5) are the most widespread polygonal feature found in permafrost regions (Black, 1976; French, 2007). They can be separated broadly into the subtypes ice-wedge, sand-wedge, composite-wedge and sublimation polygons.
Figure 7.5: Figure adapted from Figure 6, Marchant and Head (2007). A) (Top to bottom) shows a photograph of sublimation polygons from the Antarctic Dry Valleys (FOV ~50 m) and a block diagram indicating how they might develop. B) (Top to bottom) shows a photograph of sand-wedge polygons from the Antarctic Dry Valleys (FOV ~100 m) and a block diagram indicating how they might develop. C) (Top to bottom) shows a photograph of ice-wedge polygons from the Antarctic Dry Valleys (FOV ~50 m) and a block diagram indicating how they might develop.

Thermal contraction of ground materials occurs as a result of low (below-zero) temperatures and rapid cooling, resulting in the contraction and expansion of the ground (Lachenbruch, 1962; French, 2007). Cracks form when the tensile stress due to contraction on cooling exceeds the tensile strength of the material (Lachenbruch, 1962; Haltigin et al., 2012). This is particularly prevalent in ice-cemented material as the expansion coefficient of ice is far higher than that of most silicates (Greene, 1963) and the tensile strength relatively low compared to rocks. Therefore, ice-cemented material is more likely to result in significant thermal contraction cracking where cold sub-zero air temperatures are present and where temperature losses are rapid (Lachenbruch, 1962; French, 2007).

Thermal contraction cracks form polygonal networks with average diameters in unconsolidated sediments of 10’s of metres; in consolidated bedrock, they form smaller polygons of maximum size of 5 to 15 metres (French, 2007). Theoretically, the intersection of a crack with another pre-existing crack should tend towards orthogonal angles of intersection (Lachenbruch, 1962; Plug and Werner, 2001). This implies that for angular intersections of 120° to occur, cracks must develop at a series of points almost
simultaneously (French, 2007). It has also been observed in the Dry Valleys of Antarctica that young polygons have a range of sizes and tend to be relatively large with orthogonal intersections. More mature networks, however, tend towards intersections of 120° with smaller, more regularly sized polygons (Sletten et al., 2003). This indicates that the maturity of a polygonal network may be indicated by the angle of intersection and the regularity of polygon size. The various thermal contraction polygon subtypes result from different subsurface properties and environmental conditions leading to an alternate evolution of the original fracture (Black, 1976; Levy et al., 2008a).

Ice-wedges form as a result of the refreezing of water derived from melting snow, and/or the build-up of hoar frost within the fractures (Leffingwell, 1915; Black, 1976, French, 2007). This fills the fracture and prevents re-closing, and also forms a zone of weakness that reopens yearly, building up a wedge of ice over a period of tens to thousands of years (Leffingwell, 1915; Lachenbruch, 1962; Plug and Werner, 2001). The formation of the ice wedge can cause deformation of the surrounding material, resulting in raised rims on either side of the fracture (Lachenbruch, 1962; Plug and Werner, 2001, French, 2007, Haltigin et al., 2012). French (2007) also suggests that deformation could be the result of material being pushed laterally away from the polygon centre and piling up at the margins. Ice-wedge polygons (Figure 7.5C) require moisture to form and so are less likely to occur in very arid environments (Black, 1976; French, 2007).

Sand-wedge polygons (Figure 7.5B) occur in cold, arid conditions (Black, 1976; Sletten et al.; 2003; French, 2007). Ice is found as pore ice within the sediment at a value of less than 30% by volume (Marchant and Head, 2007). They form when wind-blown sediment infills the fracture, preventing full closure (Sletten et al., 2003; French, 2007). Like the ice-wedge case, sand-wedges cause a build-up of stresses resulting in deformation of the ground on either side of the fracture, thus forming ridges (Sletten et al., 2003). As in the case of ice-wedge polygons, sand-wedge polygons will reopen in cold conditions, and further infill of surface wind-blown material leads to the growth of the wedge, causing more defined rims to form (Sletten et al., 2003). In some cases, layers of both sand and ice will be added to the wedge due to variations in humidity, leading to the formation of composite-wedge polygons (Black, 1976; Levy et al., 2008b).

It is difficult to differentiate between ice-wedge, sand-wedge and composite-wedge polygons from surface morphology alone, as the surface expression could be similar in each case (Black, 1976, Sletten et al., 2003). All of these polygon types express an initially “high-
centred” appearance, meaning that the polygon borders are demarcated by troughs surrounding the polygon centres. Over time, as the surface material is forced upwards on either side of the fractures, the polygons take on a low-centred appearance and in some cases gain ‘double-rim’ margins (Black, 1976; French, 2007). An additional effect of these processes is the realigning of clasts adjacent to the fracture (Black, 1976). Clastic material near to the fracture may slump into the depression (Levy et al., 2010), which could lead to a clastic border. It is worth noting that if the wedge is composed of ice, removal of this ice (i.e. through thaw or sublimation) results in the slumping of surface material into the fracture, leading to collapse structures and resulting in a high-centred appearance (Washburn, 1956; Black, 1976; Sletten et al., 2003; Levy et al., 2010; Ulrich et al., 2011). Ice-wedge polygons have a typical diameter range of 10 to 40 metres with maximum diameters of over 100 metres occasionally seen (Washburn, 1956; Black, 1976; Washburn, 1980). The diameters of sand-wedge and composite-wedge polygons are similar, due to similar formation mechanisms.

Sublimation polygons (Figure 7.5A) occur in locations where there is excess ice in the shallow subsurface in an environment with no wet active layer (Marchant and Head, 2007). On Earth, sublimation polygons are uncommon and occur in the Dry Valleys, Antarctica, where they are underlain by massive ice (Marchant and Head, 2007). As cracks form in the underlying ice, debris overlying the fracture can collapse into the depression, resulting in coarse-grained debris at polygon borders (Marchant et al., 2002; Marchant and Head, 2007; Levy et al., 2010; Haltigin et al., 2012). This in turn increases the rate of sublimation beneath, leading to the development of deep troughs and the typically high-centred appearance of this polygon subtype (Marchant et al., 2002; Sletten et al., 2003; Levy et al., 2006; Marchant and Head, 2007; Levy et al., 2010). The location of ice in the subsurface significantly impacts polygon morphology, as does the aspect of the slope that they are forming on (Marchant et al., 2002; Levy et al., 2008a). Sublimation polygons in Beacon Valley, Antarctica, have diameters ranging from 6 to 35 metres (Marchant et al., 2002; Levy et al., 2008a). Sublimation polygons tend to be smaller in diameter than sand-wedge polygons where the two occur in similar environmental conditions (Levy et al., 2006; Levy et al., 2010).

7.2.3.3 Desiccation Polygons

Desiccation polygons (Figure 7.6) are commonly found in arid playa environments where the ground is extremely hard and flat (Neal et al., 1968; Loope and Haverland, 1988).
They generally form in lacustrine deposits composed of fine grained aeolian material, which are able to accommodate a significant amount of volatile material (Neal et al., 1968; Loope and Haverland, 1988; El Maarry et al., 2010; El Maarry et al., 2012; El Maarry et al., 2014).

Desiccation cracks form as a result of volume loss due to the evaporation or diffusion of volatiles such as ice or water as summarised by El Maarry et al. (2012) and El Maarry et al. (2014). This volume loss causes a change in surface tension between grains, resulting in a build-up of stress (El Maarry et al., 2010; El Maarry et al., 2012). Once this stress exceeds the strength of the ground material, a fracture can form (Neal et al., 1968; Weinberger, 2001; El Maarry et al., 2010). Fractures take months to years to develop as seasonal flooding only permeates shallowly into hard playa material (Neal et al., 1968). Once they form, desiccation fractures are generally irregular, jagged and less than 30 centimetres wide (Neal et al., 1968). Young fractures are discontinuous and material may slump into them (Neal et al., 1968). Secondary fractures develop at right angles to the initial fracture, leading to polygonal patterns with orthogonal intersections in plan-view (Neal et al., 1968; Loope and Haverland, 1988; Weinberger, 2001). Non-orthogonal intersections can occur either as a result of the development of a series of fractures simultaneously in a homogenous material (El Maarry et al., 2014), or the maturation of a network from the repeated opening and closing of fractures due to recurrent wetting and drying (Neal et al., 1968; Loope and Haverland, 1988). Further contraction fracturing often subdivides large polygons into regular, smaller polygons giving a nested appearance (El Maarry et al., 2014).
Figure 7.6: A) Image showing large desiccation cracks from Coyote Lake, California. After Figure 1A, El Maarry et al. (2012). B) Image showing desiccation stripes, ~ 400 metres in length, from the Indian Springs Playa, Nevada. After Figure 8, El Maarry et al. (2010).
Desiccation polygons occur across a range of scales. Primary desiccation fracture polygons rarely exceed diameters of 1 metre (Lachenbruch, 1962; Loope and Haverland, 1988; El Maarry et al., 2014), although much larger polygons have been observed in some playa basins (e.g. Figure 7.6A). These polygons can have diameters of between 15 to 100 metres, or even up to 300 metres (Neal et al., 1968). As polygon diameter is related to the thickness of the layer subject to fracturing, the stressed region needs to be thick to produce larger polygons (Neal et al., 1968; Groisman and Kaplan, 1994; El Maarry et al., 2014). It is thought that large desiccation polygons occur as a result of intense evaporation due to a period of increased aridity, combined with lowering of the ground-water level as a result of geological and/or human activity (Neal et al., 1968; El Maarry et al., 2012). The fissuring process might be aided by gradual subsidence or sudden earthquakes (Neal et al., 1968).

Desiccation polygons typically exhibit a pentagonal to hexagonal appearance (Neal et al., 1968; Weinberger, 2001; El Maarry et al., 2014). They can grade into stripes or parallel fractures as a result of constriction in narrow zones (Figure 7.6B; Neal et al., 1968). Desiccation polygon centres are often depressed relative to the polygon lip, and at intersections a collapse hole is often visible (Neal et al., 1968). Over time, wind-blown material will fill these fractures leading to a surface “stain” even when the topographic signature of the fracture is not visible (Neal et al., 1968).

7.2.4 Data and Methods

7.2.4.1 Data

Observations of landforms and landscapes in and around Lyot crater, indicate that clastic polygons are recognisable as organised, rough-textured features with clasts demarcated by dark shadows in 6 m/pixel context camera (CTX) images (Figure 7.7; Malin et al., 2007). Although CTX data is suitable for the mapping of clastic polygonal networks, it is not suitable for an in-depth study of these polygons, since the clasts which demarcate polygon edges are generally beneath the resolution of the data. Due to this, higher resolution images from the High Resolution Imaging Science Experiment (HiRISE) have been chosen for use in this study.
Figure 7.7: CTX image (B17_016484_2346) showing a section of a clastic polygonal network visible as organised, rough textured, polygonal features. The presence of clasts is inferred both by comparison of similar features in HiRISE images and from the dark shadows that are cast by the polygon margins. Image credit: NASA/JPL-Caltech/MSSS.

HiRISE data has a ground pixel size of 0.25 to 1.3 m/pixel allowing metre-scale objects to be resolved (McEwen et al., 2007), although the spatial coverage for this dataset is low on the global scale. In addition, where HiRISE stereopairs are available, digital elevation models (DEM}s) and orthoimages can be produced (Kirk et al., 2008). This allows information such as slope and aspect to be extracted from the data.

Six areas or ‘strips’ were identified as having HiRISE image coverage suitable for use in this study (Figure 7.8). These images cover clastic polygonal networks to the north and northeast of Lyot crater, with five strips clustered in a similar location. One of the strips has a HiRISE stereopair available (ESP_033059_2345 and ESP_032980_2345); where other images are absent of repeated stereo cover, CTX stereopairs (B19_016985_2338 and B21_017697_2303, B19_016906_2346 and D15_033059_2344) have been used instead (Figure 7.8). The method of producing DEMs from stereopairs is described by Kirk et al. (2008), and we have followed that approach here. DEMs used within this study were generated using SOCETSET software. The HiRISE DEM so produced has a grid spacing of 1 metre, whereas the CTX DEMs have grid spacings of 18 metres. For analysis of aspect and
slope, the DEMs were smoothed using a moving-window running mean to remove the effects of small topographic features, with a window size of 100 metres for the HiRISE DEM and 400 metres for the CTX DEMs. Slope and aspect maps were derived from the smoothed HiRISE and CTX DEMs. All morphometric measurements and post-DEM processing was performed using ArcGIS 10.1 software.

**Figure 7.8:** The red box within the inset map indicates the extent of the main image. The background is a MOLA hillshade. The strips are labelled according to the naming convention used in this study. Black lines indicate the crater rim, inner ejecta scarp and outer ejecta extent. Image credit: MOLA Science Team.

### 7.2.4.2 Morphometric Analyses

To collect morphometric data, the observed polygons must be digitised and analysed. Our method of analysis is based upon that of Ulrich et al. (2011), who used a similar technique to digitise thermal contraction fracture networks. Clastic polygonal networks were manually digitised onto HiRISE data using ArcMap 10.1 software using a Cassini projection centred on the 30° East meridian. Only polygonal networks in which the polygons could be reliably interpreted as clastic were mapped. This was done by digitising polygon margins down their centre-lines with the start and end points corresponding to the intersection of other polygon edges or a lack of further clasts. If all of the lines formed
a continuous, closed feature then this was considered to be a clastic polygon. All the
candidate clastic polygons visible in each strip were examined and digitised.

The overlaying of mantling deposits on the polygons can make it challenging to
identify whether clasts are present, as well as obscuring the polygon geometry. If there was
a high level of uncertainty surrounding the identification of a clastic polygon edge it was
left unmarked.

Once each strip was completely digitised, morphometric parameters were calculated
and extracted using a tool written in the Python scripting language for use with ArcMap.
The key parameters extracted are listed in Table 7.1. Figure 7.9 illustrates how the values
are extracted for a typical clastic polygon. In addition, the length and width of the five
largest clasts in a polygon were measured for fifteen randomly selected polygons from each
strip. These measurements were made manually within ArcMap, with the clasts selected
by observing which are the largest within the polygon at a high zoom level.

The parameters chosen were based upon those selected by Ulrich et al. (2011), as well
as other studies of polygonal networks (Neal et al., 1968; Yoshikawa, 2003; Burr et al.,
2005; Treml et al., 2010; Barrett, 2014). The digitised parameters were automatically added
as an attribute table attached to the digitised polygons and then exported so that statistical
values could be calculated (e.g., mean, minimum, maximum, first quartile, median, third
quartile, standard deviations and skewness values). All graphs were created using R (R Core
Team, 2013) within RStudio (RStudio Team, 2015).
Figure 7.9: HiRISE (ESP_016985_2315) image of a digitised clastic polygon from the Lyot study area with the scheme of the key extracted values displayed (listed in Table 7.1). The green shaded area indicates the minimum bounding area of the polygon which takes the form of a simplified polygon. This is used to extract length, width, orientation and the number of sides. The polygon area is taken as the area within the blue digitised lines, and the perimeter is taken as the lengths of each of the blue lines added together. Other values such as size and circularity are calculated from the extracted values. Image credit: NASA/JPL/University of Arizona.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Data Source</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Number of sides</em></td>
<td>HiRISE</td>
<td>The number of sides are extracted from simplified polygons.</td>
</tr>
<tr>
<td><em>Length (m)</em></td>
<td>HiRISE</td>
<td>The length of the longest axis of the simplified polygon.</td>
</tr>
<tr>
<td><em>Width (m)</em></td>
<td>HiRISE</td>
<td>The length of the minor axis of the simplified polygon.</td>
</tr>
<tr>
<td><em>Perimeter (m)</em></td>
<td>HiRISE</td>
<td>The length of all of the polygon sides added together.</td>
</tr>
<tr>
<td><em>Area (m²)</em></td>
<td>HiRISE</td>
<td>The area enclosed within the polygon.</td>
</tr>
<tr>
<td><em>Orientation (°)</em></td>
<td>HiRISE</td>
<td>The orientation of the longest polygon axis with values of between 0 and 180° measured from true north.</td>
</tr>
<tr>
<td><em>Size (m)</em></td>
<td>HiRISE</td>
<td>$= \sqrt{\frac{4A}{\pi}}$ where A is the polygon area.</td>
</tr>
<tr>
<td><em>Circularity</em></td>
<td>HiRISE</td>
<td>$= \frac{4\pi A}{P^2}$ where A is the polygon area and P is the polygon perimeter. 0 indicates an elongate ellipse and 1 indicates a circle.</td>
</tr>
<tr>
<td><em>Distance from crater centre (km)</em></td>
<td>HiRISE</td>
<td>The distance measured from Lyot crater centre to the centre of the polygon.</td>
</tr>
<tr>
<td><em>Intersection Type</em></td>
<td>HiRISE</td>
<td>A count of the number of sides meeting at each polygon vertex.</td>
</tr>
<tr>
<td><em>Intersection Angle (°)</em></td>
<td>HiRISE</td>
<td>The angle between the sides at an intersection.</td>
</tr>
<tr>
<td><em>Clast Length (m)</em> (for 15 polygons per strip)</td>
<td>HiRISE</td>
<td>The length of the longest axis of the five largest clasts.</td>
</tr>
<tr>
<td><em>Clast Width (m)</em> (for 15 polygons per strip)</td>
<td>HiRISE</td>
<td>The length of the minor axis of the five largest clasts.</td>
</tr>
<tr>
<td><em>Clast Size (m)</em> (for 15 polygons per strip)</td>
<td>HiRISE</td>
<td>$= \sqrt{\frac{4A}{\pi}}$ for the five largest clasts where A is the clast area.</td>
</tr>
<tr>
<td><em>Mean slope (°)</em></td>
<td>Smoothed HiRISE or CTX DEM</td>
<td>The mean slope across the area of the polygon.</td>
</tr>
<tr>
<td><em>Mean aspect (°)</em></td>
<td>Smoothed HiRISE or CTX DEM</td>
<td>The mean aspect across the area of the polygon.</td>
</tr>
</tbody>
</table>
Table 7.1: Morphometric parameters and topographic properties extracted and calculated for each polygon network digitised. The parameters selected are based upon those used by Ulrich et al. (2011).

7.2.5 Observations and results

Six strips of HiRISE data were surveyed for clastic polygonal networks, with a total of 3588 polygons digitised, of which 3197 occur within available, high resolution topographic data (HiRISE or CTX DEMs). During the digitisation process, clastic polygonal features were also scrutinised for morphological features that might provide insight into their formation process. This section is therefore divided into qualitative observations and quantitative morphometric data.

7.2.5.1 Polygon Observations

Clastic polygonal features are recognisable on both CTX and HiRISE data as comparatively rough-textured, patterned areas, surrounded by dark shadows cast by the clasts. In HiRISE images, individual clasts can be easily seen. The polygons are located within areas of terrain with kilometre scale hummocky relief. They are constrained to the higher topography at the upper parts of the hummocks; depressed areas are marked by a lack of polygonal features. However, it should be noted that many of the depressions found near to the polygonal networks contain infilling, smooth, mantling deposits, which could obscure polygonal features (Figure 7.10).

The networks themselves appear to change shape in relation to differences in topography, with clastic polygons “bending” around depressions (Figure 7.10A) and occasionally becoming drawn out into small boulder fields in constricted regions (Figure 7.10B). Equally, clastic polygons can be seen to form oriented patterns on and around higher topography features (Figure 7.10B). Towards the boundary of the network, the edges become discontinuous and isolated polygons are observed, many of which appear nearer circular in shape (Figure 7.11A). In some locations, the polygonal networks do not grade into discontinuous forms, but instead become draped by mantling deposits which obscure the clasts (Figure 7.11B).
Figure 7.10: A) HiRISE (ESP_026466_2345) image of clastic polygons “bending” around a large depressed region infilled with a smooth, mantling unit. B) HiRISE (ESP_016985_2345) image of clastic polygons oriented around a higher topographic feature. Clasts to the east of the high topographic feature have been drawn out as a result of constriction between the depressed region and the area of higher topography. Image credit: NASA/JPL/University of Arizona.
Figure 7.11: A) HiRISE (ESP_019685_2315) image of the edge of a clastic polygonal network which appears to become discontinuous, with isolated, more circular, forms being present with wider clastic borders. B) HiRISE (ESP_017829_2345) image of clastic polygons which appear to have become draped by a mantle deposit. Image credit: NASA/JPL/University of Arizona.
Figure 7.12: CTX (B20_017552_2335) image of a large boulder field present within the outer ejecta blanket of Lyot crater. The clastic material is recognisable as a low albedo rough texture found on areas of higher topography. Clastic material bends around depressions which often contain smooth, mantle material. Image credit: NASA/JPL-Caltech/MSSS.

In addition to the areas of clastic polygonal ground, large boulder fields occur within the outer ejecta blanket of Lyot Crater, which have a similarly distinctive topographic relationship with interspaced mantle-filled depressions (Figure 7.12). No large boulder fields have been observed near to the polygon nets studied, although the landscape related to the polygons contains a large volume of clastic material, both within the polygons, and as a light covering of small clasts throughout the region interpreted from shadows cast by material below the resolution limit. There are some small boulder fields bordering polygonal networks but there is often a cleared boundary area between the end of the polygonal network and the start of the boulder field (Figure 7.13).
Figure 7.13: HiRISE (ESP_017829_2345) image of the edge of a clastic polygonal network where a small boulder field can be observed. There is a cleared area located between the end of the polygonal network and beginning of the small boulder field. Image credit: NASA/JPL/University of Arizona.

The clastic polygons are often irregular in size and form, although they generally approximate hexagonal and pentagonal shapes. In many large polygons, discontinuous lines of clasts partition the interior, suggesting either that the larger polygon formed from merging smaller polygons, or that an existing large polygon was partitioned into smaller areas after it formed (Figure 7.14). In such examples, the clastic line is often draped with mantle material indicating that it could continue underneath the mantling deposit. Alternatively, Soare et al. (2016) suggest that sorting could become truncated early on during the process of development, leaving the landform in adolescence. The vast majority of polygons appear to emerge from superposing mantling material. This can make identification of features associated with many terrestrial polygons, such as fractures along polygon sides, difficult.
Figure 7.14: HiRISE (ESP_016906_2345) image showing a large clastic polygon which contains discontinuous internal clastic lines. Due to a change in albedo of the mantle and bumps where boulders would be expected, it is indicated that smaller polygons are contained in the interior which have become overlain with mantle material. Image credit: NASA/JPL/University of Arizona.

Figure 7.15: A) HiRISE (ESP_017829_2345) image showing a variety of different clastic features including large angular clasts aligned in rows and smaller rounded clasts which are generally located...
further from polygon edges. B) HiRISE (ESP_016985_2315) image showing clastic polygons with double rims of parallel angular clasts, occasionally with a slight gap between them, which is characteristic of many polygon edges. Image credit: NASA/JPL/University of Arizona.

The polygon margins are composed of clasts which range in size and form from small and approximately circular boulders, just at the image resolutions, to large and ridge-like clasts of up to 15 metres in diameter. Small boulders can also be found within some polygons (Figure 7.15A). The polygon borders sometimes have a “double-rimmed” appearance, where there are two lines of clasts roughly parallel to each other, sometimes with a slight gap between them (Figure 7.15B). The clasts which form the polygon edges are commonly angular with square to rectangular shapes (Figure 7.15). They often line up end to end to form the polygon sides and many give the appearance that they have formed from the fracturing of larger, more elongate clasts. The topographically high clasts give the polygons a low-centred appearance (Figure 7.16). No fractures or troughs have been observed to demarcate the polygon borders, although the presence of probably younger overlying mantle makes this difficult to constrain. Polygon margins are high standing and clasts do not appear to overlie or be contained within fractures or troughs (Figure 7.16). Examination of many topographic-profiles (made from the best 1 m-resolution HiRISE DEM, and avoiding any possibly noisy areas) across the polygon margins show no evidence of the clasts sitting within troughs at the margins.
Figure 7.16: A) Anaglyph (ESP_032980_2345_ESP_033059_2345) showing part of a clastic polygonal network. Scale is approximate. Image credit: NASA/JPL/University of Arizona. B) 3D view of part of a clastic polygonal network using an orthoimage generated from HiRISE data (ESP_032980_2345) and the corresponding 1m/pixel DEM. There is a vertical exaggeration of 1.2. Figure created using ArcScene 10.1. Image credit: NASA/JPL/University of Arizona.

In summary, the clastic polygons located around Lyot crater are generally hexagonal to pentagonal in form and occur on a range of scales (see Section 3.2). They are constrained to areas of higher topography, though polygons occurring in depressed regions might be obscured by mantling units. Polygon morphology is observed to have been affected by topographic features causing them to be orientated around areas of high or low topography and become elongate in form in constricted regions. Polygon edges are commonly composed of angular clasts which can exhibit a “double-rimmed” appearance. No fractures or troughs have been observed demarcating polygon edges. Finally, we have not observed polygon networks and boulder fields being spatially adjacent within the outer ejecta blanket. This is in contrast to Earth where sorted patterned ground commonly occurs in a pre-existing clastic layer such as glacial till or blockfields (e.g. Goldthwait, 1976; Wilson and Sellier, 1995; Grab, 2002).

7.2.5.2 Morphometric data

The key morphometric data gathered from 3588 polygons are presented in Tables 7.2, 7.3 and 7.4. These tables provide summarised statistical values for each parameter of interest.

The clastic polygons across the six strips are relatively uniform in shape, with no statistically significant difference between the mean values recorded for the individual strips. Clastic polygons are commonly 5 to 6 sided with a mean polygon size of 130 metres. The distribution of polygon sizes in each strip is broad, indicated by the quartile ranges in Figure 7.17, but as a whole the populations are fairly similar for each strip. Strips 1 and 4 appear to show a more confined normal distribution when compared to the other areas. Polygon circularity values are generally constrained to values of between 0.5 and 0.9 with a mean value of ~0.7 (Figure 7.18). This indicates that polygons tend to be equidimensional rather than elongate in shape.
Figure 7.17: Boxplots displaying polygon size values for all of the strips analysed. The boxes represent the interquartile range of each strip. The bold horizontal line is the median of each strip. The whiskers represent the upper and lower quartile. The outliers are located 1.5 times the interquartile range above the upper quartile.
Figure 7.18: Boxplots displaying polygon circularity values for all of the strips analysed. The boxes represent the interquartile range of each strip. The bold horizontal line is the median of each strip. The whiskers represent the upper and lower quartile. The outliers are located 1.5 times the interquartile range above the upper quartile.
Table 7.2: Morphometric parameters summarised for all polygons analysed within the study.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Length (m)</th>
<th>Width (m)</th>
<th>Size (m)</th>
<th>Circularity</th>
<th>Intersection Angle (°)</th>
<th>Underlying Slope (°)</th>
<th>Clast Length (m)</th>
<th>Clast Width (m)</th>
<th>Clast Size (m)</th>
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</thead>
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<td>Mean</td>
<td>179</td>
<td>117</td>
<td>130</td>
<td>0.712</td>
<td>116</td>
<td>2</td>
<td>5</td>
<td>3</td>
<td>5</td>
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<tr>
<td>Maximum</td>
<td>676</td>
<td>450</td>
<td>435</td>
<td>0.956</td>
<td>268</td>
<td>12</td>
<td>15</td>
<td>10</td>
<td>12</td>
</tr>
<tr>
<td>Minimum</td>
<td>19</td>
<td>11</td>
<td>15</td>
<td>0.226</td>
<td>19</td>
<td>0</td>
<td>1</td>
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<td>1</td>
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<td>3588</td>
<td>3588</td>
<td>3588</td>
<td>21267</td>
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<td>450</td>
<td>450</td>
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<tr>
<td>Median</td>
<td>172</td>
<td>111</td>
<td>126</td>
<td>0.727</td>
<td>114</td>
<td>2</td>
<td>5</td>
<td>3</td>
<td>4</td>
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<tr>
<td>1st Quartile</td>
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<td>79</td>
<td>91</td>
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<td>93</td>
<td>1</td>
<td>3</td>
<td>2</td>
<td>3</td>
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<tr>
<td>3rd Quartile</td>
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<td>149</td>
<td>164</td>
<td>0.793</td>
<td>139</td>
<td>3</td>
<td>7</td>
<td>4</td>
<td>6</td>
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<td>Standard Deviation (SD)</td>
<td>78.053</td>
<td>10.799</td>
<td>56.179</td>
<td>0.114</td>
<td>34.397</td>
<td>1.413</td>
<td>2.196</td>
<td>1.571</td>
<td>1.993</td>
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<td>0.699</td>
<td>-0.645</td>
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<td>0.816</td>
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<td></td>
<td></td>
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<td></td>
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<td>4-ray</td>
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<td>6.7</td>
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</tr>
</tbody>
</table>

Table 7.3: Intersection types that occur within the polygonal networks summarised across all strips.

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<thead>
<tr>
<th>Number of Sides</th>
<th>Number</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>3-side</td>
<td>236</td>
<td>6.8</td>
</tr>
<tr>
<td>4-side</td>
<td>593</td>
<td>17.0</td>
</tr>
<tr>
<td>5-side</td>
<td>912</td>
<td>26.1</td>
</tr>
<tr>
<td>6-side</td>
<td>781</td>
<td>22.4</td>
</tr>
<tr>
<td>7-side</td>
<td>503</td>
<td>14.4</td>
</tr>
<tr>
<td>8-side</td>
<td>258</td>
<td>7.4</td>
</tr>
<tr>
<td>9-side</td>
<td>144</td>
<td>4.1</td>
</tr>
<tr>
<td>10-side</td>
<td>64</td>
<td>1.8</td>
</tr>
<tr>
<td>Total</td>
<td>3491</td>
<td>100.00</td>
</tr>
</tbody>
</table>

Table 7.4: Number of sides per polygon.

The polygon intersection values show a distribution centred around 120° (Figure 7.19). The highest percentage of intersections (Table 3) are 3-ray (93.1%), with only 6.7% of intersections recorded as 4-ray. This indicates that the polygons tend towards equiangular nets, as opposed to orthogonal nets.
histograms of polygon network intersection angle

Figure 7.19: Histograms displaying polygon intersection angle values for each of the strips analysed. The median for each strip is marked as a red line.

7.2.6 Analysis

7.2.6.1 Key relationship between parameters – hypothesis testing

Sorted patterned ground on Earth formed by freeze-thaw processes displays clear relationships between circularity and average slope values (Washburn, 1956; Goldthwait, 1976), aspect and polygon orientation (Washburn, 1956), and between polygon size and clast size (Washburn, 1956; Goldthwait, 1976; Bertran et al., 2010). On Earth, circularity is inversely related to underlying slope (i.e. sorted polygonal forms become elongated downslope on steeper slopes). Also, terrestrial observations indicate that larger polygons tend to have larger clast sizes in their borders (Washburn, 1956; Goldthwait, 1976; Bertran et al., 2010). We therefore investigate whether there are any relationships between polygon circularity and average slope, polygon orientation and average aspect, and polygon size and clast size, which might support a periglacial origin for the clastic polygons in Lyot.

There is no evidence for any relationship between polygon circularity and underlying slope (Figure 7.20). There is also no evidence of any relationship between polygon orientation and underlying aspect, signifying that polygons are randomly oriented (Figure 7.21). This indicates that the shape of polygons does not vary with slope, and therefore polygons do not appear to grade into elongate forms downslope as slope increases. This
could be the result of a small amount of data available for polygons on slopes of greater than 6°; according to Goldthwait (1976), polygons and nets form on slopes of 2° to 4°, ellipses form on 3° to 6° slopes and stripes occur on slopes of 4° to 11°. Finally, there is no correlation between polygon size and clast size (Figure 7.22). In fact, the smallest polygons can contain the largest clasts and vice versa. These three null results, taken together, argue against the Lyot clastic forms having a periglacial, freeze-thaw origin: they do not show any of the relationship found in terrestrial freeze-thaw sorted polygons and circles.

Figure 7.20: Scatterplots displaying the relationship between the circularity of polygons and their mean slope. There is no suitable 3D data available for strip 2 and therefore no slope data.
Figure 7.21: Scatterplots displaying the relationship between the orientation of polygons and their mean aspect. There is no suitable 3D data available for strip 2 and therefore no orientation data.

Figure 7.22: Scatterplot displaying the relationship between polygon size and clast size.

7.2.6.2 Comparison Datasets

Having obtained measurements and observations of these polygonal features, we can use these data to test hypotheses for how they formed. However, to gain additional insight into their possible formation mechanisms we also gathered comparative data for other polygonal patterned grounds, both from Earth and Mars. The various polygon datasets
have been collated from the literature, and cover a variety of different polygon types including thermal contraction crack polygons, sorted patterned ground and desiccation polygons (i.e. not just positive-relief margin, clast-bounded polygons). We have done this in order to consider the widest possible suite of possible formation mechanisms. For example, it could be imagined that negative-relief margin (i.e., trough or fracture-bounded) polygons could become infilled by clasts or debris and then inverted by differential erosion to form features similar to those seen here. Hence, we need to consider many types of polygonal ground. However, it has been difficult to locate morphometric measurements of similar detail to those conducted in either this study or that of Ulrich et al. (2011). Accurate intersection angle and clast size data has been particularly challenging to locate. As a result, many parameters do not have comparison data available.

First, we consider the basic plan-view shape and size of the Lyot polygons. Figure 7.23 shows mean polygon size values plotted against mean circularity for each strip. The comparison data include small thermal contraction crack polygons from Earth and Mars (Ulrich et al., 2011), and sorted patterned ground from Earth and Mars (Barrett, 2014). The figure indicates that, although Lyot polygons are far larger, their circularity values are within the range of both sorted patterned ground and thermal contraction crack polygons. Sorted patterned ground values show a larger spread compared to the other polygon types and, in general, the mean values for circularity from Lyot are closer to the values indicated for thermal contraction crack polygons.
Figure 7.23: Scatterplot of comparison data displaying the relationship between circularity and polygon size. The thermal contraction crack polygon values for Earth and Mars are taken from Ulrich et al. (2011). The sorted patterned ground values for Earth and Mars are adapted from Barrett (2014). For further information about these comparison data refer to Table 7.5. One standard deviation on the mean is used for the error bar values.

Next, we examine whether larger Lyot polygons contain larger marginal clasts, as might be expected for sorted periglacial patterned ground on Earth. Figure 7.24 shows polygon size plotted against clast length. The datasets used as comparisons are all taken from sorted patterned ground. These include data from Iceland, Earth and around Lomonosov Crater, Mars (Barrett, 2014). Polygon data for known patterned ground displays a characteristic positive correlation where clast length increases with polygon size. The data for Lyot crater does not sit along this line, indicating that it does not follow this relationship. In fact, the clasts observed are similar in mean length to the larger measurements from Lomonosov Crater, although the polygon size values are far greater for the Lyot examples.

Figure 7.24: Scatterplot of comparison data displaying the relationship between clast length and polygon size. Skagi, Tindastoll and Lomonosov values are adapted from Barrett (2014). One standard deviation is used for the error bar values.

To test whether polygons of any formation mechanism are analogous in size and shape to those observed around Lyot crater exist on Earth or Mars, polygon size and lengths for our Lyot data and a variety of other datasets are compared (Table 5). Table 7.5 shows that
the polygons found at Lyot crater are within the range of size values for large thermal contraction cracks from Utopia Planitia and the South Polar Layered Deposit on Mars, and also within the size ranges displayed by large desiccation polygons observed occasionally in playa environments on Earth. Sorted patterned grounds have size values that are far lower, as are the sizes of thermal contraction crack polygons on Earth. This might indicate that thermal contraction polygons on Mars are able to form with larger diameters than those on Earth, or that larger thermal contraction polygons are more poorly preserved on Earth.
<table>
<thead>
<tr>
<th>Location</th>
<th>Source</th>
<th>Polygon Type</th>
<th>Size (m)</th>
<th>SD</th>
<th>Length (m)</th>
<th>SD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lyot Crater, Mars</td>
<td>This study</td>
<td>To be determined</td>
<td>130.5</td>
<td>56.18</td>
<td>179.2</td>
<td>78.05</td>
</tr>
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<td>Skagi, Iceland, Earth</td>
<td>Barrett, 2014</td>
<td>Sorted patterned ground</td>
<td>0.3</td>
<td>0.02</td>
<td>0.3</td>
<td>1.02</td>
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<td>Tindastoll, Iceland, Earth</td>
<td>Barrett, 2014</td>
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<td>4</td>
<td>0.88</td>
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<td>1.68</td>
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<td>High Sudetes, Central Europe, Earth</td>
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<td>Sorted patterned ground</td>
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<tr>
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<td>Ulrich et al., 2011</td>
<td>Thermal contraction cracks</td>
<td>29.7</td>
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<td>14.9</td>
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<td>Dry Valley, Antarctic, Earth</td>
<td>Yoshikawa, 2003</td>
<td>Thermal contraction cracks</td>
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<td>Desiccation Polygons</td>
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<td>Possible thermal contraction cracks</td>
<td>137.7</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Athabasca Valles, Mars</td>
<td>Burr et al., 2005</td>
<td>Possible thermal contraction cracks</td>
<td></td>
<td></td>
<td>~25</td>
<td>~10</td>
</tr>
<tr>
<td>Giant Polygons, Mars</td>
<td>Yoshikawa, 2003</td>
<td>Km-scale polygons</td>
<td>7236.6</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

**Table 7.5**: Comparison data displaying the average sizes and lengths for various polygons taken from the literature. Where an average value is not available a range has been used.
The comparison of the Lyot clastic polygon data with data from other studies of polygonal features indicates that the Lyot crater polygons are comparable in size and circularity to some thermal contraction crack and desiccation polygons, but not to sorted clastic polygons. More specifically, they are similar in size and shape to thermal contraction crack polygons located on Mars, as the terrestrial examples tend not to reach as large a size. Desiccation polygons can also occur at a similar scale to those which occur around Lyot, and the literature indicates that they possess a similar plan view to thermal contraction cracks. The plots analysed demonstrate that the Lyot forms have different morphological values than other sorted patterned ground: sorted patterned grounds are more circular than the polygons located around Lyot crater, although the spread of data could encompass those at Lyot, and they are far smaller in size. This could be a result of the sorted patterned ground data selected for use in this study, but we note that there are almost no published observations of sorted patterned ground with diameters greater than ~10-20 m. A further relationship, that between clast size and polygon size, also indicates that the Lyot polygons do not follow the typical “sorted patterned ground” relationship, with clasts too small and polygons too large.

7.2.7 Discussion

7.2.7.1 Comparison of the Lyot Clastic Polygons and Periglacial Sorted Patterned Ground

Clastic polygonal networks around Lyot crater morphologically resemble sorted patterned ground in that they have clastic material and low centres (Figure 7.1B). However, their large size is unusual. Sorted patterned ground rarely contains polygons or circles greater than a few metres in diameter, although maximum diameters of around 10 metres have been observed on Earth (Washburn, 1956; Treml, 2010; Feuillet et al., 2012) and possible examples on Mars of around 23 to 25 metres in diameter have been observed (Balme et al., 2009; Barrett, 2014; Soare et al., 2016). The typical polygon found at Lyot crater is ~130 metres in diameter, more than 10 times the size of those found on Earth, and more than 5 times the size of the other examples found on Mars. This would indicate an exceptionally large sorting depth of between ~33 to 43 metres based on a polygon diameter to sorting depth ratio of 3 to 4 (Ballantyne and Harris, 1994; Treml et al., 2010). A sorting depth of this scale is far larger than any previously observed and seems unlikely. This indicates that the formation of these polygons through freeze-thaw cycling alone is questionable. It has been observed that large polygons can form by the amalgamation of smaller sorted polygons over time (Kessler and Werner, 2003). This could potentially be
indicated by the discontinuous clastic lines observed in the large polygon interiors (Figure 7.14). On the other hand, burial by mantling units, as opposed to the elimination of small polygons, is another possibility — and could suggest that polygon sizes are somewhat overestimated. However, even taking this internal partitioning into account, and noting that it is not seen in all of the observed polygons, it seems unlikely that the difference in size between the Lyot clastic polygons and previous observations of more typical sorted polygonal ground is due to observation of only the largest polygons in the network, while smaller, internal ones are somehow always buried.

Terrestrial observations also indicate that there is a relationship between polygon diameter and clast size, with a ratio ranging from 1:5 to 1:10 (Goldthwaite, 1976). This positive correlation is not observed in the case of the clasts demarcating Lyot polygons. Furthermore, according to the ratio provided by Goldthwait (1976), the clasts in the Lyot polygons should be between 13 and 26 metres in diameter, but even the largest clasts observed around Lyot polygons are smaller than this lower limit (Table 1). It should also be noted that there is no evidence of the imbrication of clasts, which can occur as a result of frost sorting processes (Dahl, 1966; Kessler and Werner, 2003; French, 2007, Soare et al., 2016). Also, the observations of double lines of clasts, and of clasts that appear to have formed from the fracture of larger, elongate ridges of material do not match a periglacial sorting model (Figure 7.15).

Finally, periglacial sorted patterned ground has a distinctive relationship with topography in which geometric forms change from polygons on slopes of 2 to 4°, to ellipses on slopes of 3 to 6°, and finally to stripes on slopes of 4 to 11° (Goldthwait, 1976). Therefore it is expected that sorted patterned ground on Mars will behave in a similar way, and that larger slopes will lead to a lower value for circularity indicating elongation. Such a relationship has been observed at other locations on the surface of Mars (e.g. Gallagher et al., 2011). Although a qualitative relationship between topography and the polygons is indicated from observations, there is no positive correlation between circularity and slope, and high circularity polygons have been recorded on slopes as large as 12°. Also, no clastic stripes have been observed in this study.

Overall, the morphometric analysis and qualitative observations do not support the hypothesis that freeze-thaw sorting processes are responsible for formation of the clastic polygonal features around Lyot. Both Martian and terrestrial data indicate that there should be a stronger relationship between clast size and polygon diameter, and between
polygon form and underlying slope and aspect than is observed (e.g. Goldthwait, 1976; Gallagher et al., 2011; Barrett, 2014). It should be noted that little quantitative data describing the relationship between polygon elongation and slope angle is currently available on Mars, although this relationship has been observed qualitatively. Similarly, there is also little comparison data available quantifying the relationship between clast size and polygon diameter on Mars. Nevertheless, the combined observations and measurements and their comparison with terrestrial and martian datasets suggest that a periglacial sorting origin is not the best working hypothesis to explain the Lyot clastic polygons.

7.2.7.2 Comparison of Lyot Clastic Polygons and Thermal Contraction Crack Polygons

Thermal contraction crack polygons vary in morphology and size depending upon subsurface properties and the environment in which they form. The Lyot crater study area contains landforms and landscapes indicative of a cold environment with brief, possibly climate-driven, periods of fluvial activity. Fractures would be affected by topography and this could explain the oriented polygonal features observed around topographic features. This suggests that the thermal contraction of ice-cemented soils could be a potential mechanism to form patterned ground in this area – especially if a mechanism exists that could allow a “wedge-type” polygon to evolve into a “clast-bounded” polygon. The formation of these wedge-type polygons around Lyot is discussed here, and their morphology and morphometric properties are compared to those of the Lyot clastic polygons. We discuss whether thermal contraction features could evolve to form the clast-bounded polygons observed later.

Typical diameters for terrestrial ice-wedge polygons range from 10 to 40 metres with maximum diameters of over 100 metres (Washburn, 1956; Black, 1976; Washburn, 1980; Maloof et al., 2002). On Mars, possible thermal contraction polygons of a similar size to the clastic polygons around Lyot crater, with diameters ranging from 30 to 170 metres (Figure 7.25; Yoshikawa, 2003; Lefort et al., 2009), have been observed in Utopia Planitia. Hence, there is evidence that thermal contraction cracking on Mars can produce polygonal ground of the same spatial scale as the Lyot clastic polygons – although a simple thermal contraction crack mechanism would not explain their clastic boundaries. It has been also observed that thermal contraction polygons potentially subdivide into smaller polygons over time (Greene, 1963; Black, 1976), this might be analogous to the internal partitioning observed inside some of the larger Lyot clastic polygons (Figure 7.14).
Taking each subtype separately, sand-wedge and composite-wedge polygons appear more likely to produce the gross polygon morphology and scale observed at Lyot. Sublimation polygons are rare on Earth, and require underlying massive ice or excess ice (as opposed to ice-rich sediments or soils) directly beneath the layer in which the polygons form. While this might have been the case here, we note that sublimation polygons are usually both smaller than wedge type polygons, and form broader dome-like features with more poorly defined marginal troughs. Hence we think that they are less likely to be good analogues for the Lyot clastic network patterns than wedge-type polygons.

Sand-wedge and composite-wedge polygons can form low-centred polygons as a result of the forcing of material upwards on either side of the initial fracture. This can lead to the realigning of clasts and the production of a double rimmed appearance (Black, 1976; French, 2007), as is seen at Lyot (Figure 7.15). Although ice-wedge polygons can also produce this surface morphology, one might expect ice exposed to the Martian surface environment to sublime, leading to the formation of pits or other signs of collapse and eventually a high-centred appearance, as described for other parts of Mars (Levy et al., 2011), unless it were shielded from the surface by a layer of desiccated material (e.g., Lefort et al., 2009). Ice-wedge polygons also indicate the presence of a more humid environment, to allow for the growth of an ice-wedge, whereas sand-wedge polygons occur in a more arid environment where ice occurs in pore spaces within the sediment (Pewe, 1959; Black, 1976; French, 2007). Although Lyot has experienced periods of fluvial activity, indicating the past presence of water (Dickson et al., 2009; Fassett et al., 2010, Hobley et al., 2014), and ice-rich material is located nearby in the form of mantling units, it might be expected that ice deposited in the fractures would sublime without sufficient cover, leading to the formation of troughs. It appears more likely that wind-blown sediment might infill exposed fractures resulting in a sand-wedge. Alternatively, obliquity-driven changes in climate (Laskar et al., 2004) might have resulted in variations in humidity and pressure leading to composite-wedge or ice-wedge polygon formation.

The polygonal networks around Lyot possess similar intersection angles to terrestrial mature thermal contraction polygonal networks, tending towards 120° with, on average, 5-sided polygons (French, 2007). The Utopia Planitia polygons also tend towards equiangular intersections of ~120° (Lefort et al., 2009), although they appear more regular than the Lyot networks (Figure 7.25). Thus they are somewhat similar in both polygon size and network shape to the Lyot polygons.
In both terrestrial thermal contraction polygons, and the polygons present in Utopia Planitia, troughs bound the polygons. Troughs are not visible around the clastic polygons at Lyot.

Figure 7.25: HiRISE (PSP_002070_2250) image of possible thermal contraction crack polygons located in Utopia Planitia (45°N, 88°E). The polygons have somewhat similar sizes and morphologies to the network shape seen in the polygons located around Lyot crater. The terrain in this area contains scalloped depressions (Séjourné et al., 2011) and a surficial layer of small clastic material. Image credit: NASA/JPL/University of Arizona.

Hence, while it can be shown that there are examples of Martian thermal contraction polygons similar in size and overall shape to the polygonal networks seen in Lyot, and that the action of either a sand-wedge or composite-wedge polygon formation mechanism appears plausible in this current environment, the clastic appearance of the Lyot polygons provides a very clear contrast to the usual morphology of thermal contraction polygons on Earth or elsewhere on Mars. It is possible that the gravitational slumping of overlying clastic material into the fractures could be responsible for allowing a clastic-bounded polygon to evolve from a thermal contraction crack polygon, but in the Lyot study area there is no relationship between boulder fields and polygons, there is no evidence for troughs at the polygon margins, and there is generally an absence of boulders within the polygons.
between the clastic margins: the clasts seem too well-confined to the polygon boundaries
to have formed by gravitational slumping from what was originally a boulder-rich terrain.

We therefore provide another possible, but somewhat speculative, alternative
formation hypothesis: if a thermal contraction crack network formed, and was then infilled
by later materials, then this infilling material could become indurated or cemented within
the fractures. This material could then be revealed in positive relief if there were general
downwearing of the landscape, perhaps due to climatic changes and enhanced aeolian
erosion. These lines of inverted, erosion resistant fracture-fill could then themselves
degrade to produce clasts. This explanation is consistent with some of the detailed
observations of the Lyot polygons, such as linear clasts, and double lines of clasts that might
have formed from material forming on either side of a centrally fractured sand-wedge or
composite-wedge. Furthermore, it is also broadly in keeping with the overall polygon size
and network shape.

7.2.7.3 Comparison of Lyot Clastic Polygons and Desiccation Polygons

Having compared the Lyot clastic polygons to thermal contraction fracture polygons,
we now consider whether they could have formed as, or evolved from, desiccation
polygons. Desiccation polygons have been observed at a wide range of scales on Earth:
although they rarely exceed 1 metre in diameter, rare examples of up to 300 metres across
exist (Neal et al., 1968). This sets them within the size range of the clastic polygons around
Lyot crater. Therefore, a process similar to that suggested for thermal contraction cracks
above, by which infill of fractures, induration/cementation of fill, and inversion of the fill
forms clastic polygons, might be invoked.

However, to produce desiccation polygons of this scale, the stressed region would
need to be thick to produce deep fractures, and the environment would need to have first
been very humid, then have experienced a period of intense aridity combined with
lowering of the ground-water level (Neal et al., 1968; El Maarry et al., 2012). Desiccation
polygons are generally associated with remnant lacustrine deposits in a playa environment
(Neal et al., 1968; Loope and Haverland, 1988; El Maarry et al., 2010; El Maarry et al., 2012;
El Maarry et al., 2014). This indicates that sustained fluvial activity would be necessary to
form large desiccation polygons. On Mars, such a body of water could be a crater lake
sustained by a hydrothermal system, as thermal energy would need to be provided to
sustain a body of water on the surface of Mars (El Maarry et al., 2010; El Maarry et al.,
2012; El Maarry et al., 2014). The clastic polygonal features around Lyot appear to be
geologically recent and so must have formed far later than the impact event which formed the crater, so any hydrothermal activity, or remnant heat from the impact would have long-since dissipated.

Although there is geomorphological evidence consistent with the presence of fluvial activity on the inner ejecta blanket, there is no evidence of standing bodies of water occurring on the outer ejecta blanket of Lyot, and the clastic polygons occur on areas of higher topography rather than depressions where standing water would be present. Given the geographical setting, and the fact that the clastic polygons appear to form on the upper parts of hummocks, it seems unlikely that desiccation cracking is the origin of the clastic polygonal ground around Lyot.

7.2.7.4 Other mechanisms

Polygonal grounds of somewhat similar network morphology to the clastic polygons located around Lyot crater include the thermal contraction of cooling lava (e.g. Peck and Minakami, 1968), the subaqueous contraction of sedimentary gel, termed syneresis (e.g. Dewhurst et al., 1999), the formation of polygonal fault systems via the intersection of a series of normal faults within marine basins (e.g. Tewksbury et al., 2014), and the polygonal weathering of sedimentary rocks (e.g. Williams and Robinson, 1989). We now briefly consider whether any of these mechanisms could have created, or evolved to form, the Lyot clastic polygons.

The thermal contraction of cooling lava can produce polygons up to decametres in size (Peck and Minakami, 1968; Grossenbacher and McDuffie, 1995; Hetényi et al., 2012). These polygons, also referred to as columnar joints, have a high centred appearance and tend towards a hexagonal shape (Toramaru and Matsumoto, 2004). Polygons of this type are far smaller than those observed around Lyot crater and possess a different typical morphology. There are also no lava-like flows associated with Lyot polygons, no possible sources of lava visible, and the distribution on top of small hummocks, and around the crater, cannot be explained by this polygon type. Thus the thermal contraction of lava can be eliminated as a potential hypothesis.

Syneresis is a process that occurs in fine-grained sediments such as mudstones within a subaqueous environment (Pratt, 1998; Coker et al., 2007). Polygons attributed to syneresis vary in diameter from crack patterns of centimetres in scale (Pratt, 1998) to polygonal fault systems of decametres to kilometres in scale (Dewhurst et al., 1999; Coker et al., 2007; Moscardelli et al., 2012). Typically, such polygons are found in slope or basin-
floor depositional settings (Moscardelli et al., 2012). Lyot polygons are low-centred, as opposed to high-centred, and occur at a different range of scales to that typical of syneresis polygons. There is also a lack of evidence indicating a subaqueous basin environment, and because of this it is expected that syneresis would be unlikely to have occurred. Thus, it is suggested that a hypothesis involving syneresis can be eliminated.

Syneresis is also the primary mechanism suggested for the formation of polygonal fault systems formed via the intersection of a series of normal faults. Such fault systems occur in deep marine basin environments, with typical polygon diameters of kilometres (Dewhurst et al., 1998; Watterson et al., 2000; Moscardelli et al., 2012; Tewksbury et al., 2014). Polygons around Lyot crater average 130 metres in size, which is significantly lower than polygons formed by polygonal fault systems. There is also a lack of evidence for a deep marine environment around Lyot crater at this time. Due to these factors it is suggested that polygonal faulting can be eliminated as a hypothesis.

Polygonal weathering occurs on the exposed surfaces of boulders and rock outcrops composed of massive sandstone (Williams and Robinson, 1989). The cracks are pentagonal to hexagonal in shape and can be flat, concave or convex (Williams and Robinson, 1989). Most such polygons are between 10 and 20 centimetres in diameter (Williams and Robinson, 1989), with maximum polygon sizes recorded of metres in scale (Chan et al., 2008). Small polygons can be contained within large polygons giving a nested appearance (Chan et al., 2008). The polygons around Lyot crater are typically over 100 times larger than the largest polygonal features formed by weathering on Earth. The Lyot polygons also do not display a nested appearance, although small nested polygons could perhaps be obscured by later mantling deposits. It is therefore suggested that the polygonal weathering of sedimentary rocks can also be eliminated as a possible formation hypothesis for the Lyot clastic polygons.

7.2.7.5 Possible formation mechanisms for the clastic polygons around Lyot Crater

Based upon the previous discussion it is suggested that thermal contraction crack polygons, in particular sand-wedge or composite-wedge polygons, provide the best morphometric and morphological analogue to the polygonal network patterns of the Lyot clastic polygons. On Mars, the best size and shape analogue is found in Utopia Planitia, proposed to have formed through thermal contraction cracking (Yoshikawa, 2003; Lefort et al., 2009).
However, thermal contraction cracking does not explain the concentration of clasts at the boundaries of the polygon margins. There are a number of suggested mechanisms by which clastic material can concentrate in thermal contraction polygon boundaries. These include freeze-thaw cryoturbation, as in sorted patterned ground, gravitational slumping, and CO$_2$ frost ratcheting, whereby boulders become locked in a layer of CO$_2$ ice and are forced outwards as thermal contraction and expansion occurs (Orloff et al., 2013). The size of the clasts found in the Lyot polygons are almost certainly too large to have been moved via a freeze-thaw or CO$_2$ ratcheting process, so we reject this as a mechanism.

We now consider gravitational slumping of a pre-existing population of clasts. This might occur if clasts were ejected during the Lyot impact event and later concentrated into the thermal contraction fractures by gravity (e.g., Levy et al., 2010). In this case we would expect there to be a pre-existing boulder field of large clastic material around the polygonal features, as clasts the size of those found at Lyot are unlikely to have travelled far before slumping into the fractures. Taking this into consideration we would expect: (i) evidence of an association of the polygonal networks with boulder fields and (ii) for clasts of a similar morphology to those observed within the polygon margins to be present away from these margins, within the polygons centres. These relationships have not been observed: the large boulder fields are located away from the clastic polygonal networks, and, where a small boulder-rich area is present, we find a cleared margin between the edge of the clastic network and boulder-rich area (Figure 7.13). Large angular clasts are not observed located away from polygon margins, only smaller, more circular clasts are seen (Figure 7.15). Also, gravitational slumping does not account for the large ridge like clasts that have been observed in distinctive double rim patterns (Figure 7.15).

Another indicator of gravitational slumping is for the clastic material to be associated with fractures or located within troughs (Levy et al., 2010; Barrett et al., 2017). We have seen no evidence of either fractures or troughs associated with the clastic material, polygonal margins are high standing and clasts do not appear to be located within depressions (Figure 7.16). It is also not certain how angular boulders of the scale of those observed at Lyot would be moved via a process of gravitational slumping as the slopes seen in this study are low, with a maximum angle of 12° (Table 2). Finally, a recent study (Barrett et al., 2017) has indicated that fracture control mechanisms (such as gravitational slumping) of clasts into patterns could largely be a result of chance, and that fractures did not seem to substantially influence the arrangement of clasts at most sites where both
clasts and fractures are present. Therefore, we reject the idea that gravitational slumping of a pre-existing population of boulders into thermal contraction cracks was the process by which the clasts forming the Lyot polygons were concentrated.

We therefore return to the mechanism suggested previously: the clasts are the result of the infilling of pre-existing polygonal fractures with wind-blown sediments and/or ice, followed by cementation or induration of this fill, and then differential erosion, which exposes the network. Finally, weathering the indurated fill material formed the pattern of blocks seen today. In this case, the network is the trace of the exposed sand-wedge or composite-wedge network itself. This mechanism explains many of the observations and morphological features of the polygons, although the large size of the polygons is still difficult to fully explain, given that many are larger than even the thermal contraction crack polygons identified in Utopia Planitia (Yoshikawa, 2003; Lefort et al., 2009). Furthermore, the suggested mechanism in which fill material is indurated, then fractured, to form the clasts does not appear to have been observed elsewhere on the surface of Mars or on Earth; terrestrial clastic polygons are generally the result of gravitational slumping or freeze-thaw processes. This indicates that either the method for clast formation is unique, or the environment must have been unique such that a previously observed mechanism acted on such a scale. Furthermore, there is no clear reason why fracture fill should become indurated or cemented here, but not in other locations on Mars. Further investigation is clearly required to more fully understand how these enigmatic features have formed.

7.2.7.6 The affect of the spatial distribution of clastic polygons and “primary” formation hypotheses

A key characteristic of the Lyot polygons is their spatial distribution: their specific location on the Lyot ejecta blanket, at a narrow range of radial distances from the crater rim, indicates a genetic link with the crater itself. It seems unlikely that the polygons formed in this pattern by chance. This indicates either that this region of the ejecta hosts a unique material type, and/or a unique process has acted on the materials here. Given the radial distribution of the polygons, it seems the explanation must lie with the composition of the ejecta – and hence with the type of material ejected from a particular depth during the impact event. For example, if a vertically constrained, sub-surface layer of a specific composition was penetrated by the Lyot-forming impact, ejecta from this layer might be deposited in a radial band at a certain distance from the crater. It might be speculated that a water-rich layer could thus result in a region of ejecta with high water or ice content being
emplaced which, in turn, could be susceptible to certain processes that led to the unique polygonal morphologies seen here today.

Alternatively, the spatial link between the crater and clastic polygons could suggest a completely different formation hypothesis to those proposed above: namely that the polygons are primary, structural features picked out by erosion. In this scenario, and like many impact events (Rodríguez et al., 2005), the formation of Lyot resulted in extensive fracturing of the surface. Radial and concentric fault systems are often found around impact craters (Rodríguez et al., 2005). It is possible that such a fault system could then have been exploited by hydrothermal fluids, resulting in the formation of vein networks. Later erosion might result in the exposure of these veins as a high relief feature. An issue for this hypothesis, however, is that fracture networks around impact craters do not possess an organised polygonal appearance. Furthermore, the fracture network would be expected to be more continuous around the crater rather than within only a specific area of the outer ejecta blanket, and it also seems unlikely that such fractures would penetrate through the outer, poorly consolidated ejecta to the surface.

Another “primary” mechanism for the formation of polygonal fractures or patterns in the ejecta of Lyot is that of fracturing of impact melt ponds, as seen in several craters on the Moon and Mercury (e.g. Xiao et al. 2014). While some examples of such fracturing are superficially similar to the patterns seen at Lyot, they are generally much larger (> 100s of metres in diameter) and also form in well-defined, low standing melt ponds. This is not the case in Lyot, as the polygons appear to be forming on the tops and margins of hummocks in the outer ejecta. Thus, while primary formation mechanisms might be invoked, we think they are unlikely, and that the most likely formation mechanisms for the creation of the Lyot clastic polygons are secondary, and occurred after the outer ejecta was emplaced. We therefore suggest that the relationship between the distribution of the polygons and the site of the Lyot crater is due to the composition of the outer ejecta material as opposed to a structural relation.

7.2.8 Conclusion and Further Work

Enigmatic clastic polygonal networks located around Lyot crater were studied using morphometric analysis and qualitative morphological observations to provide information by which a possible formation mechanism might be inferred. Amongst the large population of observed polygons on Mars, the Lyot polygons are unique on Mars in that they have clastic margins and average diameters of 130 metres, far larger than clastic polygons
observed on Mars or Earth. The polygons are generally five or six sided with average intersection angles of 120°. Polygons tend toward equidimensional rather than elongate forms, with geometric shape largely independent of slope or aspect. The clastic blocks which compose the polygon edges are angular, with an average maximum clast size of 5 metres. Clast size does not increase with polygon size. These results taken together indicate that polygons are unlikely to be the result of freeze-thaw processes.

The morphometric data we collected were compared to various terrestrial and Martian polygon datasets to observe any trends or associations. Quantitative analysis using high-resolution remote-sensing data, based upon the method of Ulrich et al. (2011), and comparison to existing polygon datasets demonstrates that the Lyot polygons are comparable in size and shape to some desiccation polygonal fracture networks on Earth and some thermal contraction polygon networks on Mars. Qualitative observations indicate the Lyot polygons are similar to large-scale sorted patterned ground or thermal contraction cracks, but with important differences, such as their very large diameters, their large clast size independent of polygon diameter, their double-rimmed clastic borders and their elongated border clasts. Based upon these data, and a literature review of other geological polygonal features, thermal contraction cracks are proposed as the most likely mechanism by which the initial Lyot fracture networks formed.

Quantitative analysis of the morphometric data, and careful qualitative analysis of the morphology and context of the Lyot clastic polygons, seems to rule out simple gravitational slumping of pre-existing boulders into fractures as the mechanism by which initially negative relief polygon boundaries became erosion-resistant, positive relief boundaries.

We describe an alternative mechanism in which thermal contraction cracks are infilled with wind-blown sediment, and potentially ice, which forms resistant-fill material. The polygons are then buried by later mantle deposits which are eroded/ablated away leading to exhumation, fracturing of the resistant fill and the formation of angular clastic borders. We note that this mechanism explains many of the characteristics of the Lyot clastic polygons, but that the induration/cementation mechanism remains to be described.

The distribution of the clastic polygonal networks indicates a relationship between the clastic polygonal networks and the composition of the outer ejecta material. However, several lines of morphological and contextual evidence argue against a primary, structural or impact-melt derived mechanism for the formation of the Lyot clastic patterned ground.
This suggests that the formation of clastic polygons is related to the material exhumed from a specific depth during the Lyot crater impact event. Such material could have been ice-rich and susceptible to certain processes leading to the formation of the exceptionally large clastic polygons observed. Thus, further work will involve the numerical modelling of this impact event to identify the depth from which such material might be exhumed, and whether this is related to the depth of the Martian cryosphere in this area.

We conclude that, although large amounts of data have been gathered that describe the setting, morphometry and morphology of these enigmatic clastic polygons, and many formation hypotheses described and tested, no single hypothesis can satisfactorily explain their origin. Our favoured hypothesis is that the polygons formed by inversion of polygonal thermal contraction crack networks by infill, cementation/induration and differential erosion.

### 7.3 Supporting work

Since this paper was published, I have conducted new studies of 130 polygons located in Utopia Planitia, Mars (45°N, 110°E), to provide data to further test whether these polygons could be a suitable analogue for those located at Lyot crater. Figure 7.26 indicates that Utopia polygons are similar in size and circularity to those distributed around Lyot crater. This supports the hypothesis that Lyot polygons initially formed as a result of thermal contraction. Thus, polygons in Utopia Planitia might be a suitable Martian analogue.

![Comparison Graph for Circularity vs. Polygon Size](image)

**Figure 7.26:** Modified from Figure 23 (Brooker et al., 2018). Scatterplot of comparison data displaying the relationship between circularity and polygon size with the addition of Utopia Planitia data.
values to the data for possible Martian thermal contraction polygons (blue circle). The thermal contraction crack polygon values for Earth and Mars are taken from Ulrich et al. (2011). The sorted patterned ground values for Earth and Mars are adapted from Barrett (2014). One standard deviation on the mean is used for the error bar values.

7.4 Conclusions

This chapter presents the first in-depth study of large (~130 km diameter) clastic polygonal landforms that, to our knowledge, are unique on Earth and Mars. These polygons are commonly five or six-sided, with intersection angles of ~120°, and a mean clast size of 5 metres. Qualitative observations were taken using the best quality 0.25 metres/pixel HiRISE data, and quantitative measurements were conducted, based upon the method of Ulrich et al. (2011). These data were then compared to data from a variety of polygonal landforms from both Earth and Mars.

Based upon this work, I conclude that the Lyot clastic polygonal landforms are unlikely to result from freeze-thaw processes and desiccation, and are most likely representative of the thermal contraction of an ice-rich material. However, this mechanism does not explain the clastic margins, for which a new mechanism is proposed. This mechanism involves the infill of polygonal fractures with wind-blown sediment, and potentially ice, which forms resistant-fill material. The polygons are then buried by later mantle deposits which are eroded/ablated away leading to exhumation, fracturing of the resistant fill and the formation of angular clastic borders.

This unique landform is found in a specific location within the outer ejecta blanket, and might be representative of ancient subsurface ice deposited from a specific depth as a result of the exhumation of material during the Lyot impact event. A thick layer of such ice-rich material might enable the exceptionally large thermal contraction polygons observed to form. Thus, clastic polygons around Lyot crater could represent the surface expression of subsurface ancient ice.

To investigate this further, I performed impact modelling of the formation of Lyot crater, and these results are presented in Chapter 8. This impact modelling helps reveal the depth from which the material in which the polygons occur was exhumed from. It can then be analysed whether this depth corresponds to material that is likely ice-rich, and further investigate if the Lyot crater forming impact might have released ancient water onto the Martian surface.
Chapter 8: Impact Modelling of Lyot Crater

8.1 Introduction

This chapter presents work on modelling of the impact event that formed Lyot crater using a hydrocode called iSALE. The primary focus of the modelling is to provide estimates of the depth from which material deposited in the ejecta blanket originally came from. Such estimates can be compared with the theorised depth of the cryosphere in this area identifying: (i) if the ejecta blanket might contain this cryosphere material, and, (ii) if the cryosphere might have been penetrated by the impactor. As previous work, presented in Chapter 7, identified clastic polygonal features that potentially indicate the deposition of an ice-rich layer, it can be tested if such features could represent the surface expression of deposited cryosphere material, and thus represent the location of ancient groundwater exposed on the Martian surface. Furthermore, through the addition of a thermal gradient, the models provide data to test if theoretical pore ice during different stages of the impact event might have been melted (or vaporised). Thus, it can be tested if subsurface ice deposits were melted and exhumed onto the surface as liquid water.

8.1.1 Fundamentals of Impact Cratering

Impact craters on planetary bodies form as a result of the hyper-velocity impact of projectiles (French, 1998; Pierazzo and Collins, 2004). The formation of an impact crater is the result of an intense shock wave that radiates through both the projectile and surface (also referred to as target) from the point of impact, which compresses the material leading to high pressures reaching hundreds of GPa (Melosh, 1989; French, 1998; Pierazzo and Collins, 2004). Once the shock wave reaches the end of the projectile and target, it is reflected back as a rarefaction wave which typically has a higher velocity than the shock wave (Melosh, 1989; Pierazzo and Collins, 2004). This reflected wave leads to decompression of the material and its release back to lower pressures, furthermore, the particles behind the wave maintain some velocity which ultimately leads to the opening of the crater (Melosh, 1989; Pierazzo and Collins, 2004). Impact cratering is a rapid and continuous process which is divided into three stages, beginning the instant the projectile hits the target (Melosh, 1989; Kenkmann et al., 2014). These stages are: contact and compression, excavation, and, modification (Gault et al., 1968; Melosh, 1989; French, 1998; Kenkmann et al., 2014).
The contact and compression stage is the shortest stage, a maximum of seconds in duration, beginning when the projectile first hits the target and ending once the projectile has unloaded from high pressure (i.e. until compression ends; Melosh, 1989; French, 1998; Kenkmann et al., 2014). The projectile penetrates 1 – 2 times its own diameter into the target (French, 1998), compressing the material and accelerating it to a large fraction of the projectile velocity (Melosh, 1989). Conversion of the kinetic energy of the projectile to internal energies of the projectile and target, results in the generation of shock waves which cause changes in the pressure, temperature and density of the material (Melosh, 1989; Kenkmann et al., 2014), at such a rate that it is mathematically treated as a discontinuity in material characteristics (Pierazzo and Collins, 2004). Materials melt and vaporise during this stage as a result of the high pressures and temperatures involved (Melosh, 1989). The shock wave compression is irreversible, whereas decompression is reversible, leading to a net temperature increase and residual particle velocity (Melosh, 1989; Kenkmann et al., 2014). The duration of this stage depends upon the projectile’s diameter, composition and impact velocity (Melosh, 1989; Kenkmann et al., 2014).

During the excavation stage, the impact crater is opened up as a result of interactions between the shock waves and the original ground surface (Melosh, 1989; French, 1998; Kenkmann et al., 2014). After the contact and compression stage, the material is not completely decelerated, the final particle velocity component is typically one-third to one-fifth of the peak particle velocity during shock wave compression (Melosh, 1989; Kenkmann et al., 2014). As the energy of the shock wave is spread out over a larger volume of material, the shock wave decays in strength to a stress wave which travels at, or above, the bulk speed of sound in the rock (Melosh, 1989; Kenkmann et al., 2014). The stress wave mobilises the target material leading to the excavation flow which opens up the crater (Melosh, 1989). The presence of the free surface induces a tangential component of the material velocity which complements the radial component, this deflects the particle trajectories towards the surface; thus, material is pushed into and away from the expanding crater (Pierazzo and Collins, 2004). In the lower levels of the target (lower than the crater), material moves mainly downward and outward, whereas at higher levels, material moves upward and outward; thereby opening the transient cavity – a bowl-shaped depression (French, 1998). The innermost material is ejected first, and to higher altitudes than the material ejected further from the crater centre (Melosh, 1989; Kenkmann et al., 2014). The excavation flow stops when the remaining kinetic energy is
unable to displace the target material against the force of its own weight – for large gravity-dominated impacts – or against the cohesive strength of the material – for smaller strength-dominated impacts (Melosh, 1989; Pierazzo and Collins, 2004; Kenkmann et al., 2014). This leads to the furthest extent of the transient crater and the end of the excavation stage (Melosh, 1989; Pierazzo and Collins, 2004; Kenkmann et al., 2014). Although longer than the first stage, the excavation stage typically lasts a maximum of minutes in duration (French, 1998).

The final stage of impact cratering is the modification stage. This stage does not have a clear end as it can include processes such as geological mass movement, erosion, and sedimentation (French, 1998). During this stage, the direction of flow is reversed leading to the collapse of the transient crater, primarily driven by gravity (Kenkmann et al., 2014). Elastic rebound of the underlying rock can also have an influence on the modification stage (Melosh, 1989). This modification results in a shallower crater that is stable in a gravity field, depending on the degree of this modification, the crater is classified as either simple or complex (Pierazzo and Collins, 2004; Kenkmann et al., 2014). Crater modification depends upon the impact energy, gravity and strength of the target material (Kenkmann et al., 2014).

Simple craters have a bowl-shaped morphology, and are commonly less than a few kilometres in diameter (French, 1998; Kenkmann et al., 2014). If the projectile was relatively coherent and of significant size, large impact craters form. If this resultant crater is above a threshold size, the initial transient crater is not gravitationally stable, which results in the formation of complex craters with central peaks, peak-rings, terraced walls, and/or flat floors (Melosh, 1989). The transition from a simple bowl-shaped morphology to a complex crater morphology, averaged globally, occurs at ~8 kilometres on Mars, though in the high polar latitudes this transition occurs at ~11 kilometres (Robbins and Hynek, 2012).

8.1.2 Why Model Impact Events?

Large impact events occur infrequently; thus, although impact craters are observed throughout the Solar System, we have yet to directly observe the formation of a large impact crater. This leads to inevitable limitations in our understanding of the mechanics of impact cratering. One way to overcome this, and thereby fill the gaps in our knowledge, is to conduct laboratory impact experiments (e.g. Gault and Wedekind, 1977, 1978; Gault and Greeley, 1978; Greeley et al., 1980). Although large amounts of useful information can
be gathered through the study of these experiments, it is difficult to extrapolate the results to factor in the scale of large impact events (Schmidt and Holsapple, 1982; Holsapple, 1993). Furthermore, key cratering mechanics, such as the mode of failure, could differ when considering large impact events (e.g. Melosh et al., 1992).

During the formation of large craters (> 1 km in size), gravitational processes dominate the later stages of crater modification, and extreme pressures and temperatures are involved (Schmidt and Holsapple, 1982; Melosh and Ivanov, 1999; Pierazzo and Collins, 2004). Such conditions are either impossible or not easy to reproduce in a laboratory. Additionally, numerical impact models, achievable through the use of hydrocodes, are powerful tools by which the process of impact cratering can be analytically represented, thus a wide range of conditions and scenarios can be tested (Collins, 2002; Collins et al., 2013). Thus, numerical modelling of impact events has become an important tool in the study of impact cratering, especially when combined with laboratory impact experiments and remote sensing studies of impact craters.

A hydrocode is a computer code used to model fluid flow by employing classical continuum mechanics (Collins, 2002; Collins et al., 2013). This involves the separation of the computational region into a grid, and for each timestep equations of motion which express the conservation of mass, momentum and energy, are solved for each cell in the grid (Collins, 2002; Pierazzo and Collins, 2004; Ivanov, 2005). The forces that act on the grid each time step are determined using: the Newtonian laws of motion, the equation of state, and the constitutive model (Collins, 2002; Pierazzo and Collins, 2004). Newton’s laws of motion are represented as a set of differential equations using the principles of the conservation of momentum, mass and energy (Collins, 2002; Pierazzo and Collins, 2004). The equation of state accounts for compressibility effects by relating pressure to density and internal energy (Collins, 2002; Pierazzo and Collins, 2004). Finally, the constitutive model describes the response of the material to deformation by relating stress to a combination of strain, strain rate effects, internal energy and damage (Collins, 2002; Pierazzo and Collins, 2004). For further information about hydrocodes refer to Benson (1992).

8.1.3 Introduction to iSALE

The impact event will be modelled using a hydrocode called iSALE (impact Simple Arbitrary Langrangian Eularian) which is based upon the SALE hydrocode (Amsden et al., 1980). In particular, iSALE-Dellen has been used, which is a more recently released version
of iSALE (Collins et al., 2016). Models have been run predominantly in 2-dimensions, though some have been run in 3-dimensions, using iSALE-2D and iSALE-3D respectively.

The iSALE-2D shock physics code (Wünnewann et al., 2006), uses the SALE hydrocode solution of Amsden et al. (1980) as a base. This hydrocode has been modified multiple times to include an elasto-plastic constitutive model, fragmentation models, various equations of state (EoS), and multiple materials (Melosh et al., 1992; Ivanov et al., 1997). Other improvements include a modified strength model (Collins et al., 2004), porosity compaction model (Wünnewann et al., 2006; Collins et al., 2011), and, dilatancy model (Collins, 2014). The 2D code assumes axial symmetry and a vertical impact.

The iSALE-3D shock physics code (Elbeshausen et al., 2009; Elbeshausen and Wünnewann, 2011) uses a solver described in Hirt et al. (1974). This code includes a strength model (Melosh et al., 1992; Ivanov et al., 1997; Collins et al., 2004) and a porosity compaction model (Wünnewann et al., 2006; Collins et al., 2011). Elbeshausen et al. (2009) describe the development history of iSALE-3D. These hydrocodes have both been validated against experimental data from laboratory-scale impacts (Pierazzo et al., 2008; Davison et al., 2011; Miljkovic et al., 2012) and benchmarked against other hydrocodes (Pierazzo et al., 2008).

8.2 Modelling Approach

Before the impact crater was modelled, a number of key input parameters needed to be decided upon. These included: the target and projectile composition, the crustal thickness, the size and velocity of the projectile, and a suitable resolution for which models were run at. Furthermore, the method of collecting data regarding ejecta exhumation needed to be considered, which involved the generation of a grid of tracers which were used to track the position of the ejected material, as well as to record key information such as the tracer ballistics values (discussed in section 8.2.3.3). A final important consideration was the geometry of the mesh used for the simulations (Figure 8.1), specifically the size of the high resolution zone. This zone is the central region of the grid that has constant cell heights and widths; beyond this region there are extension zones, where cell width increases in the same manner – defined by an extension factor – away from the high-resolution zone.
Other than the model resolution, the parameters were selected from observations of Lyot crater and by taking values from the literature. The resolution was then selected by conducting a series of models in which the resolution was varied. The rim-to-rim diameter of each modelled crater was then extracted and plotted within a scatter graph of resolution against crater diameter. The point at which the difference between the extracted crater diameters was negligible or small (i.e. the point of convergence), defined the lowest resolution that achieved a suitable representation of the impact event. This ensures the model has sufficient resolution but has the shortest computer run time necessary to model the impact. A series of test model runs, ensured that the produced crater has a morphology and diameter that suitably represents Lyot crater. The various selected parameters, and the reasons for their selection, are presented in this chapter section.

8.2.1 Projectile Properties

8.2.1.1 Projectile velocity

A velocity of 10 kms$^{-1}$ was used for the impact models. This is based upon impact probability estimates from Ivanov (2001) and Le Feuvre and Wieczorek (2011), and previous numerical modelling studies for Mars (e.g. Senft and Stewart, 2010; Ivanov et al., 2010).
**8.2.1.2 Projectile diameter**

The projectile diameter was calculated using Equation 4 from Johnson et al. (2016). To use this equation, a series of values were taken from the literature and observations of Lyot crater.

\[ D_{\text{fin}} = 1.52 \left( \frac{\rho_{\text{imp}}}{\rho_{\text{targ}}} \right)^{0.38} D_{\text{imp}}^{0.88} v_{\text{imp}}^{0.5} g^{-0.25} D_{SC}^{-0.13} \sin^{0.38} (\theta) \]  

(Eq. 8.1)

where \( D_{\text{fin}} \) is the final crater diameter, \( \rho_{\text{imp}} \) is the density of the impactor, \( \rho_{\text{targ}} \) is the density of the target, \( D_{\text{imp}} \) is the impactor diameter, \( v_{\text{imp}} \) is the impactor velocity, \( g \) is the gravitational acceleration, \( D_{SC} \) is the final rim diameter at the simple-to-complex transition, and \( \theta \) is the impact angle with respect to the target surface.

To calculate \( D_{\text{imp}} \) the other values needed to be selected, and were chosen from the literature and observations of Lyot crater. The final crater diameter of Lyot crater is \(~215\) km taken from measurements of the rim-to-rim diameter in ArcMap 10.1. The value of \( \left( \frac{\rho_{\text{imp}}}{\rho_{\text{targ}}} \right) \) used is \( 1 \), as crater diameter has only a weak dependence on the density ratio and the material used for the projectile and target in the models was the same for computational expedience. The chosen velocity of the impactor was \( 10 \) km\( s^{-2} \) (as decided in section 8.2.2.1). The gravitational acceleration value for Mars is \( 3.71 \) km\( s^{-2} \). According to Robbins et al. (2012), the transitional simple-to-complex diameter for Mars varies depending on the latitude. When separated by terrain types, the polar terrain craters have a transition at \(~7.9\) km, while northern plains, southern highlands and volcanic terrains have a transition at \(~7.0\) km (Robbins et al., 2012). Furthermore, Garvin and Frawley (1998) find a natural break between simple and complex craters, with an optimum transition at \( 7 \) km. Therefore, a value of \( 7 \) km was used in this equation. The impact angle value used was \( 90^\circ \), as the models were conducted in 2D, so all impacts in this case are modelled at a right angle to the target surface. Inputting these values, a \( D_{\text{imp}} \) value of \(~21\) km was calculated.

**8.2.2 Target Properties**

**8.2.2.1 Target (and projectile) composition**

A simple target composition was selected with two layers: one representing the crust and another representing the mantle. The mantle was modelled using dunite as the material. The crustal material was more challenging to select: both a basalt and granite material models were considered. The closest compositional analogue for the martian crust among the material models available was basalt (Pierazzo et al., 2005). However,
Simulations using the better-calibrated granite material model to represent the crust produced a final crater structure more consistent with a Lyot-scale crater on Mars than simulations with a basalt material model (Figure 8.2). Figure 8.2 compares two models for Lyot crater which have the same input parameters, except for a difference in the crustal material. These plots show that in the case of basalt the final crater does not have a peak-ring and a central peak, whereas, the model with granite is more likely to have both a central peak and peak-ring. As Lyot crater has both a peak-ring and central peak, this indicates that granite produces a model crater with a morphology closer to the observed one. The likely reason for this is that the basalt equation of state has a much lower bulk sound speed than the granite equation of state, which reduces the effectiveness of dynamic weakening during impact (Collins, personal communication, 2018).

**Figure 8.2:** Comparison of the output from the two impact models which have the same input parameters other than a difference in the selected crustal material, where \( r \) represents the horizontal distance from the axis of symmetry and \( z \) represents the vertical distance with 0 defined as the original target surface position. The red lines represent a grid of the tracers. This grid can be used to estimate the movement of material during crater formation. Purple represents the crustal material and yellow represents the mantle material. The model using granite appears to have both a central peak and peak-ring indicated by the apparent distortion of the grid upwards and sideways. Conversely, the model using basalt appears to have just a central peak indicated by the upwards distortion of the grid. These plots were created using pySALEPlot, created by Tom Davison.

### 8.2.2.2 Crustal thickness

Crustal thickness was selected by using estimations from a number of studies. Figures 6 and 7 in Neumann et al. (2004) provide thickness estimates of \(~30\) km at the latitude and
longitude of Lyot crater. Figure 1C in Zuber et al. (2000) provides an estimate of 20 - 30 km. Figure 2 from Zuber (2001; Figure 8.3) gives an estimate crustal thickness of ~30 km. Therefore, the maximum and minimum crustal thickness estimates for the location of Lyot crater are 20 km and 30 km respectively. A constant crustal thickness of either 20 or 30 km is assumed as the thickness at the location of Lyot crater does not theoretically vary by more than a few kilometres (Figure 8.2). A value of 30 km is more likely based upon values within the literature presented above.

8.2.2.3 Thermal Parameters

After the initial models used to select the resolution and test the crustal composition, a final model was run with a geothermal gradient. Using this model, the possible phase state of any pore water/ice present in the target material could be considered (i.e. would it be vaporised, melted or still present as ice?). Furthermore, the accuracy of the model was improved, as the presence of a geothermal gradient is more representative of the martian subsurface. The chosen geothermal gradient was 5 K/km with a lithospheric thickness of 200 km. These values were based on choosing an intermediate lithosphere thickness and thermal gradient between thermal profiles in Ivanov et al. (2010), who constructed assumed thermal profiles for early and modern Mars (Figure 8.4). As Lyot crater is theorised to have formed during the late-Hesperian/early-Amazonian (Dickson et al., 2009; Russell and Head, 2002; Harrison et al., 2010), these should be suitable values for this time period which is between early and modern Mars.
Figure 8.4: Taken from Figure 2, Ivanov et al. (2010). The assumed thermal profile for Mars with an “asthenosphere” at 150 km constructed from models of Mars thermal evolution: (1, 2) “modern” Mars after Nimmo and Tanaka (2005) and Spohn et al. (2001); (3) “early” Mars after Nimmo and Tanaka (2005).

The chosen surface temperature for the models is 200 K. This is based upon modelled mean annual surface temperatures at 50°N, which range from 195 – 200 K for obliquities ranging from 0 to 60° (Haberle et al., 2003).

8.2.3 Other Considerations

8.2.3.1 Model resolution

Model resolution was selected by running simulations at a variety of different resolutions – measured in cells per projectile resolution (CPPR) – starting with a resolution of 5 CPPR (i.e. a grid spacing of ~2 km) and gradually increasing the resolution until the difference in the output rim-to-rim diameter of the model crater does not vary significantly (Figure 8.5). At this point the model resolution is selected. This ensured that the crater could be best represented with the lowest computer run time necessary. Models were run up to a time of 2000 seconds after the initial projectile contact with a time step of 0.005 seconds – 5 different resolutions were tried, with a maximum resolution of 20 CPPR. The maximum resolution of the converging values was selected, which has an average computational run time of 4 days.
8.2.3.2 High-resolution zone

The high-resolution zone of the models was selected to be tall enough to capture the maximum height of the ejecta, wide enough to capture the outer ejecta extent, and deep enough to capture part of the upper mantle along with the crust. This leads to a high-resolution zone of 480 x 350 km.

8.2.3.3 Tracers

To ensure that the ejecta material could be tracked effectively tracers were used. This is because the impact model employs a Eulerian description, which uses a fixed grid through which the material in the model moves – so each cell within the grid represents the state of a spatial location through each timestep. The tracer particles use a Langrangian description and can be imagined as massless particles which move with the material, and therefore track the state of the material it is placed within throughout the modelled impact event. Thus, the tracer particles can be used to track the changing state of the material, as
well as track the location of the material as it passes through model time. Tracers were placed in the region directly beneath the maximum potential crater radius within the crust and impactor, this led to a tracer area of ~215 x 30 km in the target. The tracers had a spacing of 0.2 cells, leading to a total tracer number of ~330000.

Occasionally, tracers can be lost from the model space as a result of leaving the top or sides of the gridded model space (i.e. mesh). This leads to the loss of information about what the final position of the material being tracked was. To ensure that this information was not completely lost, a parameter called tracer ballistics (TrB) was used within the 2D model runs. This parameter leads to a theoretical surface (EJECT_V) being used within the model. If a tracer particle crosses this theoretical surface, a series of tracer fields are stored including: the time it crossed the line (TBt), the radial distance at which it crossed the line (TBr), and the vertical (TBy) and horizontal (TBx) components of velocity at the point it crossed the line. These fields can be used to manually calculate the final position of the tracer particles (and therefore ejected material), through use of the equations of motion (assuming a parabolic trajectory and no atmospheric drag). These calculations are discussed further in the analysis section.

8.2.3.4 Impact angle

The ejecta blanket around Lyot crater appears to be asymmetrical with a distinct lack of ejecta to the southwest of the impact crater (Figure 8.6). This lack of ejecta could result from the superposition of the younger deposits of Deuteronilus Mensae on top of the ejecta deposited in this region, or it could indicate that ejecta was not deposited in this region. If the asymmetry results from a lack of ejecta deposition in the southwest, then this might indicate that the impact which formed Lyot crater was oblique (Kenkmann et al., 2014).
Figure 8.6: MOLA topographic map showing Lyot crater with the outer ejecta extent, inner ejecta extent, inner peak-ring and central peak marked. The ejecta blanket is asymmetrical with a lack of visible ejecta in the southwest.

Figure 8.7 shows further possible evidence for an oblique impact. The figure shows that Lyot crater has a less well-defined crater rim, and a lower elevation peak-ring, in the southwest (yellow-orange profiles in the lower elevation profile plot). Oblique impacts can lead to craters with a depressed rim and steepened inner slope uprange of the impact (Kenkmann et al., 2014).
Figure 8.7: Taken from Figure 18, Pan et al. (2018). On the left, is a MOLA topographic map of Lyot crater with the crater rim and peak-ring marked as dashed black lines. Solid black lines represent cross-sections of Lyot crater. On the right, the corresponding cross-sections of Lyot crater are displayed.

If the Lyot-forming impact event had a significantly oblique angle (<15°), it would be expected to have a crossrange butterfly ejecta pattern with both uprange and downrange forbidden zones where ejecta is not deposited (Figure 8.8; Kenkmann et al., 2014). Furthermore, the central peak would be offset uprange from the centre of the crater, which is not observed for Lyot crater (Kenkmann et al., 2014). Impact angles of <45° would still result in asymmetric ejecta, though in this case there should be one forbidden zone uprange (Figure 8.8; Kenkmann et al., 2014). As Lyot crater has one possible forbidden zone it is likely that the Lyot-forming impact had a slightly oblique angle of ~45 - 20°, though it is most likely the impact was 45° as this is a more common impact angle (Ivanov, 2001). The uprange zone is likely the southwestern region of Lyot crater, and therefore, the impact would have originated from this direction.

The effect of the impact angle on the location of ejecta cannot be modelled within 2D, therefore, 3D models are run to better understand the cause of the asymmetrical ejecta. These models are preliminary models only, as the 3D run times are significant – taking approximately two weeks to run. These preliminary models are presented in section 8.6.
Figure 8.8: Taken from Figure 11, Kenkmann et al. (2014), showing examples of martian impact craters which form at different impact angles.

8.3 Results

A number of different models were run to ensure that the parameters selected were suitable to form a model crater with a morphology and diameter similar to Lyot crater. Models were conducted with two end member scenarios for crustal thickness, these are 20 km and 30 km. Once these models were analysed, a final 2D model run was completed using the most likely crustal thickness (based upon the literature), with the addition of the geothermal gradient.

Within this section, the results of the key impact model runs are presented. All of the models in this section were run with a resolution of 20 CPPR. Unless stated, the parameters are as listed in the previous chapter section (section 8.2). All of the plots in this chapter have been created with the pySALEPlot tool written by Tom Davison.

8.3.1 Lyot Crater (30 km granite crust)

This model, presented in Figure 8.9, uses a granite crust of 30 km thickness, impacted by a spherical granite impactor with a velocity of 10 km/s. The distortion of the grid of tracers indicates that the final morphology of the model crater appears to have both a central peak and an inner peak-ring, with a final rim-to-rim diameter of ~207 km and maximum depth of ~4 km. This is considered to be a suitable representation of Lyot crater which has a rim-to-rim diameter of ~215 km and a maximum depth of ~3.5 km (considering -3500 to represent the average surface elevation beyond the crater rim, Figure 8.7). The final ejecta blanket extends beyond the high-resolution zone, with ejecta material being lost from the top of the mesh. The ejecta blanket has been completely deposited by timestep 130 (650 seconds after the impact).
Figure 8.9: High resolution zone of the 20 CPPR model with a granite crust of 30 km thickness, where $r$ represents the horizontal distance from the axis of symmetry and $z$ represents the vertical distance with 0 defined as the original target surface position. The red lines represent a grid of the tracers. This grid can be used to estimate the movement of material during crater formation. Purple represents the crustal material and yellow represents the mantle material. The model crater appears to have both a central peak and peak-ring indicated by the apparent distortion of the grid upwards and sideways.

8.3.2 Lyot Crater (20 km granite crust)

This model, presented in Figure 8.10, uses a granite crust of 20 km thickness, impacted by a spherical granite impactor with a velocity of 10 km/s. The final morphology of the model crater appears to have both a central peak and an inner peak-ring, with a final rim-
to-rim diameter of ~223 km and maximum depth of ~6 km. This crater has a larger and deeper morphology than would be expected for Lyot crater. The impactor diameter is similar to the thickness of the crust which is expected to lead to material from greater depth being excavated during the impact event – though no mantle material is observed being excavated in this scenario. The excavation flow in this scenario also appears to have a steeper angle than the 30 km crust model. The final ejecta blanket extends beyond the high-resolution zone, with ejecta material being lost from the top of the mesh. The ejecta blanket has been completely deposited by timestep 135 (675 seconds after the impact).

**Figure 8.10:** High resolution zone of the 20 CPPR model with a granite crust of 20 km thickness, where $r$ represents the horizontal distance from the axis of symmetry and $z$ represents the vertical distance with 0 defined as the original target surface position. The red lines represent a grid of the
tracers. This grid can be used to estimate the movement of material during crater formation. Purple represents the crustal material and yellow represents the mantle material. The model crater appears to have both a central peak and peak-ring indicated by the apparent distortion of the grid upwards and sideways.

8.3.3 Lyot Crater (30 km granite crust with geothermal gradient)

This model, presented in Figure 8.11, uses a granite crust of 30 km thickness, with the addition of a geothermal gradient of 5K/km and a lithospheric thickness of 200 km, impacted by a spherical granite impactor with a velocity of 10 km/s. The 30 km crustal model was selected as the model to which a geothermal gradient would be added as the final model crater appears most representative of Lyot crater’s morphology, and the literature most commonly indicates a 30 km thick crust at the location of Lyot crater.

The final morphology of the model crater appears to have both a central peak and an inner peak-ring that is more pronounced than the model with no geothermal gradient. The final model crater has a rim-to-rim diameter of ~206 km and maximum depth of ~5 km. The model crater is not significantly different in size when compared to the model crater with no geothermal gradient. The final ejecta blanket extends beyond the high-resolution zone, with ejecta material being lost from the top of the mesh. The ejecta blanket has been completely deposited by timestep 130 (650 seconds after the impact).
Figure 8.11: High resolution zone of the 20 CPPR model with a granite crust of 30 km thickness and a geothermal gradient of 5K/km with a lithospheric thickness of 200 km, where \( r \) represents the horizontal distance from the axis of symmetry and \( z \) represents the vertical distance with 0 defined as the original target surface position. The red lines represent a grid of the tracers. This grid can be used to estimate the movement of material during crater formation. Purple represents the crustal material and yellow represents the mantle material. The model crater appears to have both a central peak and peak-ring indicated by the apparent distortion of the grid upwards and sideways.

8.4 Analysis

In this section the key models presented in section 8.4 are analysed, focussing on the maximum depth from which material was exhumed during the impact event, and where
this material is revealed on the surface in the ejecta blanket. To conduct this analysis the ejecta trajectories needed to be manually calculated to ensure that the results of the analysis are accurate. This is because some of the tracers are lost from the top of the mesh during the model run which leads to inaccurate final ejecta positions and the loss of some of the tracers. By manually calculating the final positions, all of the information about the tracer particles can be retained. The method by which the final x-positions of the tracers were calculated is presented in section 8.4.1.1.

To analyse the locations of the material from various depths within the ejecta blanket, a plot showing the thickness of impactites against the horizontal location of these impactites is created. This enables the thickness of ejecta, and locations of exhumed material from specific initial depths to be analysed. The method by which these plots were created is presented in section 8.4.1.2.

The final type of analysis conducted is made possible through the addition of a geothermal gradient to the 30 km crustal thickness model. This model enables the estimation of the state of the pore water present within the crustal material (ice or liquid water), by assuming this theoretical pore ice would have the same temperature and pressure values as the host material. The method by which the plots were created is based upon plots presented in Pierazzo et al. (2005). The plot assumptions and analysis are presented in section 8.4.4.1.

8.4.1 Ejecta Analysis Methods

8.4.1.1 Calculation of the final x-position of the ejecta

The calculation of the final horizontal positions of the tracers is presented in this section. The tracers can be used to represent the locations of packets of material during the impact event, and therefore can be used to track the ejected material trajectories.

To manually calculate the trajectories of the ejecta, tracer ballistics parameters are used. This places a theoretical surface (named EJECT_V) at a user-defined height (h) above the target surface. When a tracer crosses EJECT_V, the time it crosses (TBt), radial distance at which it crosses (TBr), and vertical (TBy) and horizontal (TBy) components of velocity at the point it crosses are recorded as tracer fields in the model output. Using these fields, the final x-position of the tracer can be calculated assuming parabolic flight and no atmospheric drag (Figure 8.12). The calculations are outlined below – these calculations have been written into a Python script which is used to batch calculate the final tracer x-
positions. These final x-positions positions are output as an array which can be used to create the plots shown in the following analysis sections.

• \textbf{EJECT\textunderscore V} is a theoretical surface used within the model. This parameter has been set to slightly above the final model crater rim height (h).

• \(u_x\) and \(u_y\) are the horizontal and vertical components of the projectile’s velocity. It is assumed that \(u_x\) will remain constant due to no atmospheric drag.

• \(h\), \(u_x\) and initial \(u_y\) are known values.

• \(t_0\), \(t_1\), \(t_2\) and \(t_f\) represent the time the tracer crosses \textbf{EJECT\textunderscore V}, the time the tracer reaches its maximum height, the time the tracer re-crosses \textbf{EJECT\textunderscore V} and the time the tracer reaches the original surface respectively.

• \(z_1\) represents the total horizontal distance that the tracer travels, where \(r_0\) represents the radius at which the tracer crosses \textbf{EJECT\textunderscore V}, and \(r_f\) represents the final tracer position. \(s_1\) is the horizontal distance the tracer travels from when it crosses \textbf{EJECT\textunderscore V} to when it re-crosses this theoretical surface. \(s_2\) is the horizontal distance the tracer travels from when it re-crosses \textbf{EJECT\textunderscore V} till it reaches its final position on the surface.

\textbf{Figure 8.12:} Schematic diagram showing the theoretical parabolic flight of a tracer particle and the various parameters used in the ballistics calculations presented in this section.

To calculate the travel time to the peak of the parabola (\(t_1\)), the vertical component of velocity (\(u_y\)) is used as follows:

\[ v = u + at \quad \text{(Eq. 8.2)} \]
\[ \therefore 0 = u_y - g_M t_1 \quad \text{(Eq. 8.3a)} \]
\[ \therefore t_1 = \frac{u_y}{g_M} \quad \text{(Eq. 8.3b)} \]

where \(g_M\) is the value for gravitational acceleration on Mars.
To calculate the travel time for the tracer to reach the vertical position of EJECT_V (t₁), the travel time to the peak of the parabola (t₂) is used as follows:

\[ t_2 = 2t_1 \quad (\text{Eq. 8.4}) \]

To calculate the distance the tracer travels to reach the vertical position of EJECT_V (s₁), the horizontal component of velocity (uₓ) is used as follows:

\[ s = ut + \frac{1}{2}at^2 \quad (\text{Eq. 8.5}) \]

\[ \therefore s_1 = u_x t_2 \quad (\text{Eq. 8.6}) \]

To calculate the time it takes the tracer to reach the target surface (tᵣ), using the vertical component of velocity (uᵧ) within Eq. 8.5 and the quadratic equation (Eq. 8.8) as follows:

\[ h = u_y(t_r - t_2) + \frac{1}{2}g_M(t_r - t_2)^2 \quad (\text{Eq. 8.7a}) \]

\[ \therefore \frac{1}{2}g_M(t_r - t_2)^2 + u_y(t_r - t_2) - h = 0 \quad (\text{Eq. 8.7b}) \]

\[ (t_r - t_2) = \frac{-b \pm \sqrt{b^2 - 4ac}}{2a} \quad \text{where} \quad a = \frac{1}{2}g_M, \quad b = u_y \quad \text{and} \quad c = -h \quad (\text{Eq. 8.8}) \]

Only the positive output from Eq. 8.8 for \((t_r - t_2)\) is retained.

To calculate the distance the tracer travels between reaching the vertical location of EJECT_V and the target surface (s₂) using the horizontal component of velocity (uₓ) within Eq. 8.5 as follows:

\[ s_2 = u_x(t_r - t_2) \quad (\text{Eq. 8.9}) \]

Using the two calculated distances (s₁ and s₂), the horizontal distance that the tracer particle travelled from the point it crossed EJECT_V till it reached the original surface (z₁) can be calculated as follows:

\[ z_1 = s_1 + s_2 \quad (\text{Eq. 8.10}) \]

To calculate the x-position of the final tracer position (rᵣ) the calculated distance (z₁) and the initial x-position at which it crossed EJECT_V (r₀) can be used as follows:

\[ r_f = r_0 + z_1 \quad (\text{Eq. 8.11}) \]

The calculated horizontal positions are then output as an array to be used within various analysis plots presented in the following analysis sections.

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8.4.1.2 Calculation of impactite thickness

The thickness of material deposited within a particular horizontal range can be calculated by approximating this range as a “ring” around the crater (represented as the grey area in Figure 8.13). The sum of the volume of tracers that land within this horizontal range can then be divided by the calculated area for this horizontal range to give an estimate of the thickness – this thickness estimate is a lower limit as the ejecta is initially more compact and will bulk up over time (Figure 8.13). This method is based on the method used by Collins et al. (2008), who created figures showing the effective thickness of impactites exposed to different peak shock pressures against the final radial position of these impactites. Within my analysis, the thickness values of impactites from each initial depth range was calculated, by binning the tracers that originate from particular initial depth ranges and separately calculating the thickness of each initial depth range for each horizontal distance. These values can then be plotted together in a stackplot showing each initial depth range as a separate colour in the plot.

**Figure 8.13:** Schematic plan-view diagram showing how the area of a horizontal range around a crater can be calculated. The crater is assumed perfectly spherical and is represented in this diagram as the shaded grey area.

\[
\text{Area}_1 = \pi r_1^2 \\
\text{Area}_2 = \pi r_2^2 \\
\text{Area of blue shaded ring} = \text{Area}_1 - \text{Area}_2
\]
8.4.2 Lyot Crater (30 km granite crust with no geothermal gradient)

Figure 8.14 displays plots that analyse the depths from which the material present within the ejecta blanket was exhumed from. Figure 8.14A shows that ejecta that was deposited at the furthest horizontal distance from the model crater centre originated from the regions ~20 km from the symmetry axis. Figure 8.14B shows that the deepest material present within the ejecta blanket came from a depth of 10 - 12 km. This material is only visible up to a distance of ~220 km. The radial distance range at which the clastic polygonal landforms are typically visible is shown as a grey shaded region on the plot. This model indicates that the material present on the surface in this region is theoretically exhumed from a depth of 6 - 8 km, though a small amount of material exhumed from 10 - 12 km is present at the starting edge of the horizontal range.
Figure 8.14: A) Plot of initial tracer positions coloured by the final horizontal position of the tracer. The black line represents the profile of the surface and impactor. Values of 0 either represent that the tracer was not ejected, or the tracer was vaporised during the impact event. B) Stackplot showing the theoretical thickness of impactites as a factor of radial distance, coloured by the initial depth from which they were exhumed. The shaded grey region represents the radial range at which clastic polygonal features are observed within the Lyot ejecta blanket.

8.4.3 Lyot Crater (20 km granite crust)

Figure 8.15 displays plots that analyse the depths from which the material present within the ejecta blanket was exhumed from. Figure 8.15A shows that ejecta that was deposited at the furthest horizontal distance from the model crater centre originated from the regions ~20 km from the symmetry axis, this does not differ significantly from the 30 km model. Figure 8.15B shows that the deepest material present within the ejecta blanket came from a depth of 12 - 14 km, and far more material is present from a depth of 10 - 12 km, when compared to the 30 km model. The deepest material is visible up to a distance of ~200 km from the crater centre, whereas the material exhumed from 10 - 12 km is present up to ~280 km from the crater centre.

Compared to the 30 km model presented in section 8.5.2, the material present in the ejecta blanket is exhumed from deeper and a larger volume of material is deposited – as indicated by thickness values almost two times larger than those recorded in the 30 km model. This model indicates that the material present on the surface in the region where the clastic polygonal features are observed was theoretically exhumed from a depth of 6 - 12km, although the vast majority is from the 6 - 8 km depth bin. This indicates that in a 20 km crustal thickness scenario the clastic polygonal features might be forming in a region of material excavated from deeper in the crust. None of the excavated material comes from the mantle as there is no material deposited from depths above 14 km.
Figure 8.15: A) Plot of initial tracer positions coloured by the final horizontal position of the tracer. The black line represents the profile of the surface and impactor and the boundary between the crust and mantle material. Values of 0 either represent that the tracer was not ejected, or the tracer was vaporised during the impact event. B) Stackplot showing the theoretical thickness of impactites as a factor of radial distance, coloured by the initial depth from which they were exhumed. The shaded grey region represents the radial range at which clastic polygonal features are observed within the Lyot ejecta blanket.

8.4.4 Lyot Crater (30 km granite crust with geothermal gradient)

Figure 8.16 displays plots that analyse the depths from which the material present within the ejecta blanket was exhumed from. Figure 8.16A shows that ejecta that was deposited at the furthest horizontal distance from the model crater centre originated from
the regions ~20 km from the symmetry axis. Figure 8.16B shows that the deepest material present within the ejecta blanket came from a depth of 10 - 12 km. This material is only visible up to a distance of ~220 km. This model indicates that the material present on the surface in the region where the clastic polygonal landforms are observed is theoretically exhumed from a depth of 6 - 8 km, though a small amount of material from 10 - 12 km is present at the starting edge of the horizontal range where they are observed. These values and plots do not differ significantly from the 30 km crust run without the geothermal gradient.

Figure 8.16: A) Plot of initial tracer positions coloured by the final horizontal position of the tracer. The black line represents the profile of the surface and impactor. Values of 0 either represent that the tracer was not ejected, or the tracer was vaporised during the impact event. B) Stackplot
showing the theoretical thickness of impactites as a factor of radial distance, coloured by the initial depth from which they were exhumed. The shaded grey region represents the radial range at which clastic polygonal features are observed within the Lyot ejecta blanket.

8.4.4.1 Analysis of the stable phase of water

Using the model with the geothermal gradient, estimates of the stable phase of water can be made by analysing the temperatures and pressures present underneath the crater. This analysis assumes that any pore ice/water has the same temperature and pressure as the host material. The values used for phase transitions are based upon Pierazzo et al. (2005), who estimate the boundaries using the phase diagram for pure water (Figure 8.17). Figure 8.17, taken from Pierazzo et al. (2005), shows that at high pressures of above ~0.012 GPa, liquid water can exist up to temperatures of ~650 K. Based upon this phase diagram, boundaries can be estimated at which changes of state will theoretically occur given the high pressures present during an impact event. These boundaries are: ~273 K for the transition from ice to liquid water and ~650 K for the transition from liquid water to water vapour.
Figure 8.17: Taken from Figure 5, Pierazzo et al. (2005). Phase diagram for pure water. At pressures of above ~0.012 GPa, the boiling point of water is increased, therefore, liquid water can exist up to the critical point (~650 K, Pierazzo et al., 2005).

Figure 8.18 shows profiles for the transient crater (Figure 8.18A) and the collapsed crater (Figure 8.18B), with material coloured by the theoretical phase state of water based on temperature. Pressure contours are also overlain as black lines which indicate that the pressure beneath the crater at both cratering stages is commonly no lower than 0.05 GPa. Figure 8.18A demonstrates that at the transient crater stage, any pore water/ice present in the material would be vaporised in the crater centre. The ejected material is generally at a temperature which could indicate liquid water or potentially ice. Figure 8.18B shows that, at the collapsed crater stage, pore ice/water within the crater centre (up to ~50 km from the axis of symmetry) would theoretically be vaporised. The rest of the crater interior is at a temperature suitable for the presence of liquid water. Furthermore, the plot indicates that liquid water could be present in the material on top of the ejecta blanket up to ~200 km from the axis of symmetry. The rest of the uppermost crustal material has a temperature which would indicate that the state of the pore material would be ice. Both models indicate that the transition from pore ice to pore water might be located at ~14 km.
Figure 8.18: Plots displaying the theoretical stable phase of water. Blue represents ice, yellow represents liquid water, and red represents water vapour. Pressure contours are displayed as overlain thin black lines. Thick black lines represent the crater profile and the boundary between the crustal and mantle material. **A)** Profile of transient crater coloured by the theoretical phase state of pore ice/water (assuming the temperature and pressure are the same as the host material). The pressure contours have been smoothed using a gaussian filter with a sigma value of 1.4. **B)**
Profile of collapsed crater coloured by the theoretical phase state of pore ice/water (assuming the
temperature and pressure are the same as the host material).

8.5 Preliminary 3D Models

More recent work has involved the modelling of the impact event in 3D to consider
the possibility of an oblique impact. Due to issues with the code this work has been
conducted late in the thesis, and as the computational run time of 3D models are long (~2
weeks), there has not been sufficient time to perform in-depth analysis of the results in the
manner that was conducted on the 2D models. The results presented here are therefore
preliminary; though still reveal interesting information about the potential impact angle of
the Lyot-forming impactor.

Parameter selection for the 3D models has slight differences from the 2D scenario due
to the addition of the 3rd dimension. Primarily, the high-resolution zone was altered to a
smaller size than the 2D models, this was to ensure that the computational runtime was
not significantly increased beyond 2 weeks. The new high-resolution zone was ~309 x ~120
x ~101 km, with the impact point offset by 30 km. This leads to axes of -180 - 120 km, 0 -
120 km, and -50 - 50 km, for the x, y and z-axis respectively. To account for the impactor
angle leading to the partitioning of part of the velocity into a horizontal component the
velocity was increased. The chosen value was 14 kms^{-1}, as this value ensures that ~10 kms^{-1}
of the velocity is partitioned into the vertical component.

The composition of the material was kept constant between the 2D and 3D models,
so the modelled crust was granite which overlies a dunite mantle. The crustal thickness
was selected based on the value most commonly estimated for the location of Lyot crater
within the literature, as well as the model crater that appeared most representative of the
gross crater morphology from the 2D models. Therefore, the selected thickness was 30 km.
The models were run with a geothermal gradient as used in the final 2D model (i.e. 5K/km
with a 200 km thick lithosphere). Impact angles were chosen to represent the most likely
impactor angle, and the minimum potential impact angle, based on observations of Lyot
crater.

The plots in this section were created using a triple plot code developed by Thomas
Davison, for use with pySALEPlot.
8.5.1 45° Impact Angle

As discussed in section 8.2.3.4, the most likely angle of the Lyot-forming impact event was \( \approx 45° \). Figure 8.19 shows that ejecta deposition occurs downrange of the impact with no obvious ejecta curtain uprange of the impact event.

![Figure 8.19: Side-view of the formation of an impact crater for a Lyot-forming scenario with a 45° impact angle. The uprange direction is to the left of the plots.](image)

Figure 8.20 shows that there is a lack of significant ejecta deposition in the downrange region directly behind the formed crater. This is morphologically similar to the ejecta blanket of Lyot crater, which appears to lack significant ejecta directly behind the crater to the southwest. If this lack of visible ejecta is a result of no (or little) deposition in this region, then this provides evidence that the impact angle could have been \( \approx 45° \), and that the impactor hit the surface from the southwest.
Figure 8.20: Plan-view of the formation of an impact crater for a Lyot-forming scenario with a 45° impact angle. The uprange direction is to the left of the plots.

The final model crater appears to have both a central peak and peak-ring (Figure 8.21). The elevation of the crater is lower in the eastern (downrange) region of the crater interior, with a minimum elevation of -8 km, and the final rim-to-rim diameter is ~231 km (Figure 8.21). This is deeper and wider than Lyot crater, however, the model does provide evidence that the elevation of a crater from a 45° impact angle would be steeper in the downrange interior, which is a feature observed within Lyot crater. The downrange crater rim is also lower in elevation with a less well-defined rim which is another morphological similarity to Lyot crater.
**Figure 8.21:** Final model crater with the side-view (top), plan-view (middle) and crater gradient/profile (bottom) displayed as separate plots. The crater profile plot has the crater diameter (in reference to the initial target surface) and rim-to-rim diameter displayed. It is observed that the crater is deepest in the downrange interior region with steep gradients of up to $\pm 0.3$ on the crater walls.
8.5.2 30° Impact Angle

An impact angle of 30° was used within the second 3D model. This scenario was chosen to test a more oblique impact angle, but not an angle significantly oblique as to lead to two forbidden zones in the surrounding ejecta blanket (see section 8.2.3.4). Figure 8.22 shows that ejecta deposition occurs downrange of the impact with no obvious ejecta curtain uprange of the impact event. The angle of this ejecta blanket is slightly steeper than the previous 45° impact angle model, though not significantly steeper.

![Figure 8.22: Side-view of the formation of an impact crater for a Lyot-forming scenario with a 30° impact angle. The uprange direction is to the left of the plots.](image)

Figure 8.23 shows that there is a lack of significant ejecta deposition in the downrange region directly behind the formed crater, however, this forbidden zone is larger than in the 45° scenario, extending across into the northeast region around the crater. This is a larger region of little (or no) ejecta deposition than the ejecta blanket of Lyot crater. The central peak of Lyot crater in this scenario is offset uprange by ~20 km when compared to the 45° impact angle model. Such an offset is not observed in Lyot crater, which has a central peak that is relatively central.

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Figure 8.23: Plan-view of the formation of an impact crater for a Lyot-forming scenario with a 45° impact angle. The uprange direction is to the left of the plots.

The final model crater appears to have both a central peak and peak-ring (Figure 8.24). The elevation of the crater is lower in the eastern (downrange) region of the crater interior, with a minimum elevation of ~8 km, and a final rim-to-rim diameter of ~240 km (Figure 8.24). This is deeper and wider than Lyot crater. The region of lower elevation is less pronounced than in the 45° impact angle scenario, though the crater itself is wider. The downrange crater rim is also lower in elevation with a less well-defined rim.
Figure 8.24: Final model crater with the side-view (top), plan-view (middle) and crater gradient/profile (bottom) displayed as separate plots. The crater profile plot has the crater diameter (in reference to the initial target surface) and rim-to-rim diameter displayed. It is observed that the crater is deepest in the downrange interior region with steep gradients of up to $\pm 0.3$ on the crater walls.
8.6 Discussion

The work in this chapter provides estimates of the maximum excavation depth of Lyot crater, and the potential phase state of the excavated material. In a 30 km crust scenario, the maximum depth of excavation is up to 12 km and this material is present up to a maximum of 220 km away from the crater centre. For a 20 km crust, the maximum depth is up to 14 km and this material is present up to a maximum of 200 km from the crater centre, though the material exhumed from 10 - 12 km is present up to a maximum of 280 km from the crater centre. These estimates for the maximum depth of excavation agree relatively well with the maximum excavation depths calculated by Russell and Head (2002), who state that material would be excavated and ejected from depths of 11 km, and Pan et al. (2018), who state that excavation depths can reach 13 - 14 km. Based on these excavation depths, it can be estimated whether the cryosphere was penetrated and the depth from which the material where the clastic polygons are located came from.

According to Russell and Head (2002), the predicted depth of the cryosphere at the latitude of Lyot crater for a variety of geothermal heat fluxes is: 0.4 km (45 mW/m²), 3.9 km (30 mW/m²), and 16.0 km (15 mW/m²). Clifford et al. (2010) predict that the Amazonian cryosphere extends to depths of up to ~13 km at this latitude for a surface heat flux of 30 mW/m² and an ice-cemented crustal thermal conductivity of 3 mW/m² (Weiss et al., 2018). Further estimates of cryosphere thicknesses are provided by David Weiss, who calculated the thickness as the depth of the 273 K isotherm using the 1D steady-state heat equation (Weiss, personal communication, 2016). They found values ranging from 3.1 - 18.6 km using heat fluxes of 60 - 10 mW/m² (Weiss, personal communication, 2016).

Based on these cryosphere estimates, the impact models indicate that the depth of excavation is sufficient to penetrate the predicted cryosphere in this region; unless the heat flux is significantly low. This agrees with the work of Russell and Head (2002) who also concluded that theoretically the Lyot-forming impact event should have penetrated the cryosphere. In this scenario the large outflow channels around Lyot crater, described by Harrison et al. (2010) and Weiss et al. (2017), could result from the artesian release of water. This formation mechanism was discounted by Russell and Head (2002) and Weiss et al. (2017), as they do not find significant evidence of the ponding of groundwater in the interior of Lyot crater. As stated by Harrison et al. (2010), and observed in the mapping of Lyot crater, such evidence could be hidden by extensive mantling deposits within the crater interior. Both Russell and Head (2002) and the impact models presented in this chapter
indicate that the cryosphere at the latitude of Lyot crater would have been penetrated by the impact event. Furthermore, the crater interior would have been covered with a sheet of impact melt (e.g. Newsom, 1980; Melosh, 1989; Kring, 1995), and cryospheric material would largely have been ejected from the crater relatively early in the impact process. Thus, I do not think that a lack of evidence for the ponding of water in the interior of the crater is sufficient to rule out a deep groundwater origin for these channels.

Regardless, the impact models presented here indicate that water would have been exhumed onto the surface during the impact either within the ejecta blanket, or through the artesian release of water. Furthermore, Figure 8.17 indicates that temperatures and pressures were sufficient for a relatively significant volume of liquid water to have been stable on the surface up to 200 km from the crater centre. This correlates approximately with the location of the inner ejecta scarp and the location of the deepest exhumed material, beyond this distance, the stable phase would be ice. This indicates that ancient water could have been deposited within the inner ejecta blanket of Lyot crater; though no evidence of such activity shortly after the impact is observed within the inner ejecta blanket, as the fluvial channels in this region are dated to significantly after the impact event (Chapter 6).

A further point of analysis was the depth from which the material in which the clastic polygonal features, presented in Chapter 7, originally came from. Analysis of the phase state of water revealed that the predominant theoretical stable phase where the polygon-forming material would have been deposited was ice. Analysis of the ejecta exhumation depths in this region indicates that this material was originally from a depth of 6 - 8 km. This indicates that the material deposited in the region where the polygons formed would have been located within the cryosphere, unless the heat flux at the location of Lyot crater is relatively high. As such, modelling indicates that the polygons might have formed within material that contained ancient ice from the cryosphere, rather than as a result of ancient groundwater, unless the geothermal heat flux was relatively high. However, according to analysis of the stable phase of pore ice/water in deposited material, liquid water present within the inner ejecta blanket could have subsequently flowed away from the crater into the outer ejecta blanket. This water could have refrozen where the polygons formed leading to a thick deposit of ice, which might explain the large diameter of the polygons that formed in this location.
It must be noted that these values are estimates based upon the impact models presented here, which do not take into consideration all of the processes involved in the emplacement of ejecta. Furthermore, the impact models are simple, considering only two layers – the crust and mantle – and using a crustal composition that is not necessarily typical of Mars, which has a predominantly basaltic crust (Zuber, 2001; Neumann et al., 2004; McSween et al., 2009). Therefore, the work presented in this chapter provides a generalisation of the Lyot-forming impact event – this is sufficient to provide first order estimates of the way in which Lyot crater formed, and the depth from which ejecta might have been exhumed, but cannot be taken to be completely accurate. These models could be improved by modification of the crustal composition to make it more accurate to Mars and the location of Lyot crater, and, the consideration of different thermal gradients and lithospheric thicknesses to test how the stable phase of pore water/ice is affected.

The 3D models presented in section 8.5 delve into testing whether the impact that formed Lyot crater could have been oblique. These models are preliminary, as there was not sufficient time to conduct a suitable analysis of the results, or fine tune the model parameters to form an impact crater more similar in diameter to Lyot crater. Based on the preliminary models, the results indicate that the Lyot-forming impact was most likely angled at 45°, as this leads to a crater that has a forbidden zone downrange of the impact, yet still has a central peak that is not considerably offset from the centre. The 30° impact angle scenario has a larger forbidden zone and an offset central peak that is not observed for Lyot crater. This indicates that the Lyot-forming impact event was unlikely to result from a significantly oblique impact, rather from a more probable 45° impact. In this scenario, the lack of ejecta in the southwest is a result of the ejecta not being deposited in this region, rather than the later deposits of Deuteronilus Mensae overlying the ejecta blanket.

8.7 Conclusions

A series of 2D and preliminary 3D models have been run using iSALE-Dellen (Collins et al., 2016), a shock physics code based upon the SALE hydrocode (Amsden et al., 1980). The 2D models use end member scenarios for crustal thickness (20 km and 30 km). Once the final model crater morphologies were analysed, the most suitable model was selected and a geothermal gradient was added. Using the models, maximum ejecta exhumation depths were extracted. These exhumation values – combined with estimations of the phase state of pore ice/water within the exhumed material – were used to study if the cryosphere at the location of Lyot crater could have been penetrated, and the possible location of ancient
groundwater, specifically to test if this location correlates with the clastic polygonal landforms studied in Chapter 7.

These analyses indicate that the maximum exhumation depth, using the most likely crustal thickness of 30 km, is 10 - 12 km. Similarly to Russell and Head (2002), this indicates that the cryosphere at the location of Lyot crater is likely to have been penetrated during the Lyot-forming impact – assuming that the geothermal heat flux at the time of the impact was not below ~30 mW/m². The location of this material is within the inner ejecta blanket, and also corresponds to material that would be at sufficient temperatures and pressures for pore water/ice to be stable as liquid water. These results indicate that pore water/ice within deposited material beyond the inner ejecta scarp – which includes the location of the clastic polygonal features (~200 - 350 km) beyond the crater rim – would be stable as ice rather than water. The material in which the clastic polygons form, corresponds to an exhumation depth of ~6 - 8 km (using a crustal thickness of 30 km). Thus, the clastic polygonal features studied in Chapter 7 potentially formed within ancient ice-rich material exhumed from the cryosphere.

Further work presented in this chapter involved preliminary 3D models, which were used to study the potential impact angle of the Lyot-forming event. Based on these models, the impact most likely had an angle of 45°, as this leads to a downrange forbidden zone similar in size to that observed around Lyot crater, but no significant offset of the central peak. The model using an angle of 30° leads to a larger forbidden zone than observed at Lyot crater, and a relatively significant central peak offset. It should be noted, that the final model craters are larger in diameter and deeper than Lyot crater. This is due to a lack of sufficient time to improve the model parameters such that the final model crater is more similar to Lyot crater.

The results in this chapter provide a guide which can be used to estimate the way in which Lyot crater formed, but the models have limits and drawbacks that should be considered (e.g. the model does not consider all the processes involved in the emplacement of ejecta). Further work could involve the improvement of the 2D models to consider different thermal gradients and lithosphere thicknesses, and a more accurate crustal composition. Furthermore, the 3D models could be analysed in a similar manner to that conducted on the 2D models, and additional models could be run to create a final model crater more similar to Lyot crater.
Chapter 9: Synthesis

In this chapter, results from the previous research chapters are combined to piece together the history of water within and around Lyot crater, and to help discuss the wider implications of my work in the context of Mars during the Amazonian. Furthermore, the astrobiological potential of the site is assessed to ascertain whether it could be a key location for future scientific exploration — either robotic or human. As each segment of research includes a discussion section, rather than reiterating previous discussions, this chapter synthesises the work into a coherent story set against the backdrop of current martian science.

9.1 The History of Water In and Around Lyot Crater

The research presented within this thesis, combined with previous studies of Lyot crater and its surroundings, has demonstrated that this location has experienced considerable modification by liquid water at points in its history. Even before the formation of the impact crater, the location where it formed — the northern latitudes of Mars — is theorised to have contained an extensive martian ocean (e.g. Baker et al., 1991; Parker et al., 1993; Head et al., 1999; Clifford and Parker, 2001; Di Achille and Hynek, 2010). In the following chapter, the research results from my work are discussed alongside martian literature to piece together the history of water in and around Lyot crater.

9.1.1 Before and During the Impact

9.1.1.1 The northern ocean hypothesis and its implications for Lyot

Although the research in this thesis does not focus on the history of the location where Lyot crater has formed prior to the impact event, the previous history of the area will have a bearing on the formation of the target material that was impacted into. Thus, a brief discussion of the prior history of the region is useful to understand the material that could be excavated during the impact event. As the history of this location is complex, this discussion focusses on crucial information when considering water-related activity in the northern latitudes of Mars, the northern ocean hypothesis.

The northern plains of Mars are theorised to have hosted a large ocean, named “Oceanus Borealis” by Baker et al. (1991). This ocean potentially covered a third of the planet during the Noachian (Baker et al., 1991; Parker et al., 1993; Head et al., 1999; Clifford and Parker, 2001; Di Achille and Hynek, 2010), and could have existed transiently into the Amazonian (Baker et al., 1991; Parker et al., 1993; Webb, 2004). The presence of this ocean
is inferred from a combination of geomorphological features including: possible relict shorelines (Parker et al., 1989; Parker et al., 1993; Head et al., 1999; Webb, 2004; Perron, 2007), the abrupt termination of many valley networks at the northern plains (Parker et al., 1993; Dohm et al., 2000; Di Achille and Hynek, 2010), the debouching of large outflow channels onto the northern plains (Baker et al., 1991; Parker et al., 1993; Fairén et al., 2003), and an equipotential surface delineated by the distribution of ancient deltas within and along the margins of the northern plains (Di Achille and Hynek, 2010).

Assuming Mars had a warmer, wetter climate during the Noachian and hosted a large ocean, then the ocean would have begun to freeze as the climate of Mars shifted to a colder climate and crustal heat flow declined (Clifford and Parker, 2001; Kreslavsky and Head, 2002b). In the highland crust, a freezing front would have developed that acted as a cold-trap for subsurface water (Clifford and Parker, 2001). As ice condensed behind the freezing front, this would have created a sink for groundwater leading to thickening of the cryosphere (Clifford and Parker, 2001). Clifford and Parker (2001) estimate that 400 million years after the Noachian, the volume of water assimilated by the cryosphere doubled. This led to a lower supply of groundwater, which might account for the observed decline in outflow channel activity (Clifford and Parker, 2001; Kreslavsky and Head, 2002b). Furthermore, some of the northern ocean is potentially present as massive ice deposits within the subsurface of the northern latitudes (Clifford and Parker, 2001; Kreslavsky and Head, 2002b). This is important when considering the material that the Lyot-forming impact event released onto the surface.

This summary of the northern ocean – and fate of the ocean – as presented primarily in Clifford and Parker (2001), has interesting implications for Lyot crater. If relict ice from the northern ocean is present in the northern latitudes, this could mean that some of the material exhumed by the Lyot-forming impact could be from this theorised ocean. As life on Earth might have originated in the oceans at sites of hydrothermal activity (e.g. Corliss et al., 1981; Baross and Hoffman, 1985; Martin et al., 2008), then potentially the martian ocean could have hosted similar suitable environments for the origination of life. Carter et al. (2015), Pan et al. (2017) and Pan and Ehlmann (2018), found signatures of chlorite-prehnite on the outer rim of Lyot crater, which is indicative of hydrothermal alteration. This signature could result from the hydrothermal alteration of crustal material prior to the impact event (or shortly after), which were then exposed by the impact event (Pan and Ehlmann, 2018). If this resulted from hydrothermal activity prior to the impact event, could
this activity have been during the time of the northern ocean? If so, this might have implications for the development of life in this location. Regardless, locations where water was present on the martian surface are the most likely locations for preserved material of astrobiological interest (McKay et al., 1992; Westall et al., 2015). Thus, the material exhumed by the Lyot forming impact, taking into account a maximum exhumation depth of 10 - 12 km (assuming a 30 km crustal thickness and a 5 K/km geothermal gradient; section 8.4.4) and the model result that some of this material was at temperatures and pressures where ice or liquid water was stable (section 8.4.4.1), might have contained material (ice, water and associated sediments) from an ancient northern ocean.

9.1.1.2 Are the outflow channels indicative of groundwater release?

Although the outflow channels around Lyot crater were not studied directly in this thesis, the modelling work presented in Chapter 8 presents further details about the material that was released during the impact event, and whether the cryosphere was punctured. This has implications for the formation of the outflow channels. Previous work by Harrison et al. (2010) and Weiss et al. (2017), focussed on how the outflow channels might have formed; both studies concluded – based on Russell and Head (2002) – that if the outflow channels resulted from the release of water during the impact event, then there should be accompanying evidence for the significant ponding of water in the crater interior, which they do not find. The impact models, presented in Chapter 8, shed new light on whether water would pool in the crater interior, and whether or not the cryosphere was penetrated.

Analysis of the stable phase of water in section 8.4.4.1 shows that material in the centre of Lyot crater would likely have vaporised during the impact event. Furthermore, this region of the crater would have been covered with a sheet of impact melt (e.g. Newsom, 1980; Melosh, 1989; Kring, 1995), and cryospheric material would largely have been ejected from the crater relatively early in the impact process. Thus, under these conditions it is not surprising that no evidence of the ponding of water is observed in the crater interior. Moreover, the later modification of the crater interior by fluvial and aeolian processes, and the emplacement of extensive mantling units and viscous flow features (presented in Chapter 5), would obscure any evidence of ponding were it present. Therefore, the lack of evidence for ponded water in the interior is not sufficient to discredit the hypothesis of groundwater release for the formation of the outflow channels surrounding the Lyot ejecta.
The crucial question remains, was the cryosphere penetrated by the Lyot-forming impact event? If the answer is potentially yes, then the simplest explanation for the outflow channels would be that they represent the release of groundwater as a result of the puncturing of this constraining layer. Prior work by Russell and Head (2002), showed that cratering mechanics predict that the event should have penetrated the cryosphere. The impact models presented in Chapter 5 corroborate this, further providing evidence that the cryosphere would likely have been penetrated by the impact event. Thus, it is suggested that the impact event did penetrate the cryosphere, and the outflow channels represent evidence of this release of ancient groundwater. Further to this, the models demonstrate that material deposited in the inner ejecta blanket would have been at temperatures and pressures such that liquid water would be stable. Therefore, deposited cryospheric material might have been melted in this region and liquid water could have flowed from the inner ejecta margins. This could have influenced the location of where the clastic polygonal features (presented in Chapter 7) formed. Potentially, this water refroze in this region leading to a thick ice-rich deposit in which the polygons could have formed. It is unlikely the water resulting from melting was the source of the outflow channels as these features are located beyond the outer ejecta margins.

Further to this potential release and deposition of water, the heat from the impact event would likely lead to hydrothermal activity (Newsom, 1980; Osinski et al., 2001; Abramov and Kring, 2007; Schwenzer et al., 2012). Potentially, the chlorite-prehnite signature discussed in Chapter 5, could be a result of this activity (Pan and Ehlmann, 2018), rather than indicating pre-impact materials. This has important implications when considering the astrobiological potential of Lyot crater, as hydrothermal systems could provide a habitable environment for life on Mars (Newsom et al., 2004; Abramov and Kring, 2005; Osinski et al., 2012).

9.1.1.3 Summary of water activity prior to and during the Lyot impact event

- The northern latitudes could have been host to an extensive ocean during the Noachian. As martian climate shifted to colder and more arid conditions, the ocean would have frozen. By the time of the Lyot-forming impact event it would have largely been assimilated into the crust and into a thickened cryosphere. Potentially, there was hydrothermal activity within the crustal material before the impact event.
- The Lyot-forming impact occurred either during the late-Hesperian or early-Amazonian, and exhumed cryosphere material onto the surface. During the impact event the...
The cryosphere was penetrated, leading to ancient water being released, which formed the outflow channels to the north of Lyot crater. Cryospheric material was also deposited onto the surface during this impact event from a maximum depth of 10 - 12 km. Material in the inner ejecta blanket might have melted, and possibly flowed beyond the inner ejecta margins before refreezing in the region where the clastic polygonal features formed. The impact event likely resulted in further, localised hydrothermal activity.

**9.1.2 Water-Related Activity Since the Impact**

**9.1.2.1 Deposition of the pitted mantle**

The next phases of water-related activity are presented in Chapters 5 and 6. Morphological mapping of the interior in Chapter 5 indicates that, after the impact event that formed the crater and the accompanying potential hydrothermal activity, the next stage of water-related activity was the deposition of the pitted units (Map Sheet 1). As summarised in section 5.6.1.2, the pitted units are interpreted as a mantling deposit. Such deposits typically result from the airfall of ice-rich material as a result of climate cycling due to changes in obliquity (e.g. Head et al., 2003). The pitted morphology of the units probably resulted from the sublimation of ice; thus, the material was deposited and then the climate shifted from net deposition to net sublimation of ice. The pitted units are dated as being mid-Amazonian in age (Dickson et al., 2009), therefore the activity presented in this chapter occurred during and after the mid-Amazonian.

Two of the mapped pitted units in Map Sheet 1 feature km-scale polygonal features (section 5.5.3): one unit features depressed “pristine” polygons, and the other features inverted pseudo-polygons. It is uncertain as to how they formed. They are located in the crater interior around the central peak in a region that could potentially have experienced hydrothermal activity (Pan and Ehlmann, 2018). Proposed mechanisms for the formation of the “pristine” polygons include: the surface expression of fractures beneath the unit formed during the impact event, or, the thermal contraction of a thick-ice-rich layer deposited around the central peak. The pseudo-polygons are composed of inverted ridges that are either discontinuous or approximate circular and polygonal shapes. These landforms could result from the differential compaction of the mantling material as a result of the underlying topography (McGill and Hills, 1992). Alternative hypotheses include that they result from the infill of fractures formed during the impact event or via thermal contraction by the hydrothermal circulation of fluids forming a mineralised vein network.
which is revealed through downwasting, or by coarse-grained aeolian material which becomes cemented and is downwasted to form high-standing ridges.

9.1.2.2 The formation of fluvial landforms

One of the primary aims of this thesis was the study of fluvial landforms, located (by grid mapping) within the inner ejecta and crater interior (presented in Chapter 6). These landforms comprise small channels and fan deposits found at the termini of many of these channels. Morphological mapping (Chapter 5) demonstrated that these landforms are incised into the pitted floor units and, rarely, the crater units, and thus can be reliably concluded to have formed after these units. This provides the upper limit for channel formation: the mid-Amazonian.

The channels are interpreted as forming subaerially, resulting from the melting of ice deposits, such as (i) the more degraded mantling units, or potentially (ii) viscous flow features formed during a prior episode of glaciation (as discussed in Chapter 6). This melting could result from a combination of climate change driven by obliquity cycles, combined with a unique micro-environment resulting from the low elevation of Lyot crater. Dickson et al. (2009) state that temperatures at the latitude of Lyot crater would have been above 273 K for significant time periods at higher obliquities, and even under present-day conditions, the pressure in the crater interior is probably above 6.1 mb (the triple point of water). This has important implications for the local climate during the Amazonian, as it implies that conditions within Lyot crater were such that relatively significant channel networks were able to form. Furthermore, not only were conditions sufficient for channel formation, there is evidence of the ponding of water within closed basins.

Standing bodies of water are inferred using a number of lines of evidence. Firstly, some fan deposit morphologies – the complex, broad and elongate fans – indicate the involvement of a subaqueous component, as they are morphologically similar to deltas and fan deltas (section 6.6.4). In particular, the broad and elongate fans commonly have steepened margins which could represent the prograding of an alluvial fan into a standing body of water. The next line of evidence is the presence of lower channel sections beneath a number of the fan deposits. These lower channel sections could correspond to outlet channels resulting from the fill and breach/spill of a ponded body of water within a closed basin. This hypothesis is supported by a fill analysis that convincingly demonstrates that the fan deposits are found at the edges of closed basins, with the potential outlet channels located at the topographically lower end of these basins. In this scenario the standing
bodies of water result from the filling of the closed basins as a result of the gravity-driven flow of water within the channels. This is an expected outcome if water from the channels terminates in a closed basin. The possibility of standing bodies of water within Lyot crater has important implications for habitability on Mars, as lakes provide a suitable environment for life (e.g. Cabrol et al., 2007), and therefore would be an ideal target for rover and/or manned missions (Cabrol and Grin, 2003). This will be discussed further in sections 9.2 and 9.3.

The timing of this activity is uncertain, and has not been studied within this thesis. Complex aggradation is generally not observed, except for within one system of inverted channels located in the southwest of the crater interior (see section 6.5.1.1). It is therefore estimated that there was only one period of fluvial activity and lake/fan formation, though whether this activity was long or short in duration is unknown. This one period of activity could indicate that there was a single point at which climate was sufficient in the Amazonian to lead to melting, potentially supported by a local increase in geothermal heat flux (e.g. Gallagher and Balme, 2015; Butcher et al., 2017); or an external event, such as an impact event, triggered this fluvial activity (e.g. Segura et al., 2002).

9.1.2.3 The most recent water-related activity

Based upon morphological mapping in Chapter 5, and the landforms observed in Chapter 4, the most recent water-related activity involved the deposition of ice-rich material, such as the mantle units (section 5.6.1.3) and viscous flow features (section 5.4.6), and the thawing/sublimation of ice-rich deposits to form landforms such as gullies, sublimation pits, and scallops. The clastic polygonal landforms, presented in Chapter 7, are also likely to have formed since the cessation of the fluvial activity that formed the small channels presented in Chapter 6, though the exact timing of their formation is uncertain. This is assumed as the clastic polygonal landforms are stratigraphically lower than the most recent mantling deposits (e.g. Figure 7.14) indicating that they formed prior to at least the most recent period of mantle deposition. However, thermal contraction generally results from sharp drops in temperature which most likely occurred after fluvial activity would have ceased, therefore I estimate that they formed after fluvial activity ceased, but before the most recent episode of mantle deposition.

The mantle units represent the highest stratigraphic units, and are thus the most recently deposited units in Lyot crater. Such features are interpreted as forming in the same manner as the flat pitted units (i.e. resulting from climate change due to variations
in orbital parameters), but these units have not undergone as much sublimation/degradation, and are thus interpreted as forming more recently. Furthermore, multiple depositional episodes are indicated by the presence of layering in these units. Potentially, the mantling deposits within the outer ejecta blanket are representative of the deposition of airfall deposits during the same climatic excursions as those which deposited the units within the crater interior. If this is the case, then these same mantling units potentially overlay the clastic polygonal features in the outer ejecta blanket, indicating they formed before the most recent mantle units were deposited.

I proposed that the clastic polygonal features result from the thermal contraction of ice-rich material (Brooker et al., 2018) that was primarily deposited from a depth of 6 – 8 km during the impact event that formed Lyot crater. Potentially, some of the ice-rich material could also have resulted from the refreezing of melted cryospheric material which was deposited in the inner ejecta blanket. Thus, the polygonal landforms could represent the deposition of ancient ice and/or groundwater, providing a guide as to where such deposits could be studied. I proposed that the clastic margins resulted from the infill of the thermal contraction fractures with wind-blown sediments, and potentially ice, which formed a resistant-fill (Brooker et al., 2018). The polygons were then buried by mantle deposits which were later eroded/ablated away leading to exhumation and later fracturing of the resistant fill to form the angular clastic margins of the polygons.

The most recent evidence of water-related activity is represented by landforms that potentially result from thaw, such as the gullies found predominantly on the central peak, and crater rim, as these features could result from the melting of frost deposits on steep slopes (e.g. Head et al., 2008, Conway et al., 2011, de Haas et al., 2015). If gullies do form as a result of melting, then they represent the most recent water-related activity in Lyot crater. This is indicated by observations that material transported from the head alcove of the gullies is deposited on top of the mantling units. Therefore, they formed after the mantling units were deposited. Thermokarst terrain, described by Glines and Gulick (2018), could also represent a more recent period of thaw/sublimation. However, these topics have not been addressed in detail in this thesis.

9.1.2.4 Summary of water-related activity since the Lyot impact event

- Mantling units are deposited across the Lyot study area during a period involving the net-deposition of ice-rich material as a result of a variation of orbital parameters. These units are the pitted floor units. Kilometre-scale polygons are potentially the surface
expression of fractures in the crater interior, or result from the later thermal contraction of ice-rich material. The timing of their formation is uncertain, though some of the fractures appear to have been exploited by fluvial activity, thus they might have formed before the channels and associated fan deposits.

- Climate shifts led to the net-removal of ice, leading in turn to desiccation of the deposited ice-rich material, and the formation of sublimation pits that characterise the pitted floor units.
- Before the deposition of the more recent mantling units, the fluvial features formed. They resulted from the melting of ice-rich material during a period of high-obliquity, after the deposition of the pitted floor units. The melting of ice-deposits led to the incision of channels and the ponding of water in closed basins, some of which overtop/breach forming outlet channels. The fan deposits also formed during this period of fluvial activity.
- Fluvial activity ceased as climate cycles resulted in the deposition of different generations of mantling units and the currently observed viscous flow features. Before the more recent periods of mantle deposition, the clastic polygonal features formed in the ejecta blanket.
- The most recent potentially water-related features are the gullies found predominantly on the central peak and crater wall, and the thermokarst terrain described by Glines and Gulick (2018).

9.1.3 Are Both Recent Atmospheric and Ancient Groundwater Recorded at Lyot Crater?

Looking at the history of water in and around Lyot crater, as presented above, both recent atmospherically-derived water and ice, and ancient subsurface water and ice might be preserved at this site. The atmospherically-derived water and ice could be represented by the small channels, fan deposits, evidence of lakes, various generations of mantling units, viscous flow features, gullies and thermokarst terrain. Impact modelling indicates that the cryosphere was punctured during the impact event, possibly resulting in the formation of the outflow channels around Lyot crater. Additionally, cryosphere material (some which could have been melted during the impact event), would have been deposited in the ejecta blanket. Furthermore, the clastic polygonal features, present in the outer ejecta, could represent the surface expression of a particularly thick deposit of this cryospheric material. In summary, the work in this thesis has demonstrated that both the
actions of recent atmospheric water and ancient groundwater very likely acted on the landscape in and around Lyot crater.

9.2 Lyot Crater as a Site of Astrobiological Interest

Section 9.1 summarises the history of water at the Lyot crater study site. As current knowledge indicates that water is important for life (Davis and McKay, 1996; Franks, 2000; Westall et al., 2000; Rothschild and Mancinelli, 2001), the history of water within and around Lyot crater has important implications for life on Mars.

When considering life on Mars, it is often assumed that, should life have developed, it would most likely have been early in Mars’ history when it was potentially warmer and wetter, but could have “died out” (at least on the surface) as climate shifted to cold, arid conditions (e.g. McKay et al., 1992; McKay, 1997). If there was life on Mars, evidence of this life could still be preserved in materials from the same time period and locations where it originated (Westall et al., 2015). As observations indicate that the northern latitudes of Mars played host to a northern ocean, material from this ocean would be a prime candidate for signatures indicative of past life. As discussed in section 9.1, the Lyot crater-forming impact would probably have exhumed material from up to 10-12 km depth, and thus cryospheric material could be preserved in the ejecta blanket of Lyot crater. Therefore, material from the time of the northern ocean could be present on the surface around Lyot crater, which would be of astrobiological interest.

Although the surface of Mars is hostile for life that does not mean that extant life on Mars is impossible. If life on Mars originated early in the planet’s history, it is possible that as the climate shifted, life could have adapted to the changing climate leading to specialised organisms able to survive in conditions considered inhospitable (Nealson, 1997; Clark, 1998). This is not unexpected, as organisms have adapted to a wide variety of extreme conditions on Earth, such that even the most extreme environments contain life (Rothschild and Mancinelli, 2001; Wharton, 2007). Experiments have even demonstrated that a bacterial strain from the Antarctic Dry Valleys (Brevundimonas sp.) could theoretically survive for up to 1.2 Ma before experiencing a $10^6$ population reduction when at a depth of 30 cm in the current martian subsurface (Dartnell et al., 2010). Thus, life could still exist in locations where conditions were once suitable for it to develop. and climate change was gradual enough for adaptation to occur, as new more extreme niches became available.
Even if life did not adapt such that it could survive on the surface of Mars, it might have survived within more habitable environments, particularly in the martian subsurface where it would be better sheltered from radiation. Alternatively, life could have developed at different times in the planet’s history if the conditions considered suitable for life were met, so life did not necessarily develop and then need to adapt to a significant climate shift. Conditions considered suitable for life include: water, protection from radiation and oxidation, and energy from volcanic activity, hydrothermal systems etc. (e.g. Cabrol and Grin, 2003; Westall, 2015).

Based on the criteria discussed above, Lyot crater would be a suitable location for life to have developed, and potentially survived to the present day. This is demonstrated by the abundance of ice-rich material, recent fluvial activity, potential standing bodies of water, and evidence of hydrothermal activity. Furthermore, the impact event that formed Lyot could have released water constrained beneath the cryosphere onto the surface, alongside material from the cryosphere itself (Chapter 8). Life might have originated in the subsurface of Mars (Maher and Stevenson, 1988), particularly where groundwater exists deep in the crust where radiogenic heating leads to temperatures above the freezing point of water (Clifford, 1993; Clifford et al., 2010). Thus, the Lyot-forming impact event might have flushed microorganisms to the surface (Westall et al., 2000), and provide the most recently excavated cryosphere material on the surface of Mars in a location where it could be studied.

9.3 Implications of this Work

The major implication of this work is that Lyot crater is a location on Mars where there has been relatively extensive water-related activity in a period of martian history largely inferred to have been arid and cold. Of particular note is the evidence for water pooling within closed basins leading to the development of channel and fan systems with lower reaches formed by the breach of basins. These systems indicate that during at least one point in the Amazonian, the climate was sufficient for ice-rich deposits to melt in this location. Furthermore, these lakes represent potential habitable environments.

The further implication is that Lyot crater could preserve material of astrobiological interest, potentially in a location where it could be reached by rovers. This material of astrobiological interest is inferred from both the conditions the crater has experienced, and the fact that material released and exhumed by the impact event provides a sample from a region of interest too deep for conventional methods to reach. For these reasons
and those outlined in section 9.1, I argue that Lyot crater is a site of high priority for further investigation, potentially building to a future mission to the location by either rovers or humans.

Lyot crater could potentially provide a suitable site for a manned mission to Mars. Evidence of surface/near-surface ice deposits is abundant allowing the ready extraction of water, a desirable resource for a human outpost (Stoker et al., 1992; Hoffman and Kaplan, 1997; Portree, 2001; Criswell et al., 2005). Furthermore, locating a habitat in the bottom of a crater could somewhat reduce the cosmic radiation that humans would be exposed to (Stoker et al., 1992). Through human exploration work, attempting to ascertain if life arose within Lyot crater could be conducted in earnest, forming the framework for exploration and answering a key question of planetary exploration programs: is there life on Mars? (Stoker et al., 1992; Hoffman and Kaplan, 1997; Drake et al., 2010).
Chapter 10: Conclusions and Further Work

10.1 Summary

This thesis describes the first in-depth study of Lyot crater, revealing a complex history of water and ice. The first stage of work involved reconnaissance grid mapping, presented in Chapter 4, which located features indicative of water/ice across the entire Lyot study area. These grid mapping results were used to identify landforms and surface types to include in a detailed geomorphological map and those which would be studied further in later chapters. Once these landforms and surface types were identified, the interior of Lyot crater was selected for detailed mapping to provide context for fluvial and glacial activity recorded in this region. This study, presented in Chapter 5, focussed on creating the first detailed geomorphological map of Lyot crater’s interior, which was used in turn to derive a stratigraphic history.

The next stages of work used observation, morphological analysis, and comparison to terrestrial and martian analogues to study landforms identified as key to the history of water in and around Lyot crater. The first landforms studied, presented in Chapter 6, were fluvial landforms within the crater interior and inner ejecta blanket. This work built upon the geomorphological mapping and presented science outcomes from this map. It involved the collection of additional data to test previously-presented hypotheses for the formation of the small channels and fan deposits. The other landforms studied were clastic polygonal networks found in the outer ejecta of Lyot crater, presented in Chapter 7. These were studied because their distribution indicates a genetic link to the impact crater, potentially representing the location of deposited ancient groundwater or ice. They are, to my knowledge, unique on both Earth and Mars and this constitutes the first study of such features. In Chapter 8, impact modelling is presented to test whether the impact event that formed Lyot crater could have released ancient groundwater onto the Martian surface, and to analyse the exhumation depth and stable phase state of the material deposited in the ejecta blanket. Furthermore, the exhumation depth of the material in which the polygons have formed is analysed, to test whether cryospheric material and/or ancient groundwater was involved in their origin.

All lines of evidence were then synthesised to build the history of water in and around Lyot crater, and discuss the implications of this research within the context of current martian science.
10.2 Conclusions

10.2.1 Summary of the Conclusions

My work has revealed new information about the history of water in and around Lyot crater, which in turn has important implications for water and habitability in the Amazonian. The important conclusions are summarised below, ordered by the constructed stratigraphic history of Lyot crater:

- Impact modelling indicates that the impact event that formed Lyot crater probably had an incidence angle of ~45°, and exhumed material from a maximum depth of 10 - 12 km. This demonstrates that the cryosphere should have been penetrated by the impact event – agreeing with the work of Russell and Head (2002). Outflow channels are interpreted as resulting from the penetration of this constraining layer. The crater is dated to the late-Hesperian/early-Amazonian (Russell and Head, 2002; Dickson et al., 2009; Harrison et al., 2010).

- The impact event exhumed cryospheric material, some of which formed from the assimilation of the northern ocean (Clifford and Parker, 2001), which has important implications for the preservation of astrobiological material on or near the surface of Mars.

- During the impact event, temperatures and pressures within the inner ejecta blanket were sufficient for pore water/ice to be stable as liquid water. Beyond this margin, the stable phase in the ejecta blanket would have been ice. Hydrothermal activity is likely to have occurred as a result of this impact event, and the crater units (shown in Map Sheet 1) are interpreted as representing the material modified by the impact event to form the central peak, inner peak-ring and crater rim.

- After a significant time lag, potentially of 100s of millions of years or more, the pitted units were deposited as an ice-rich airfall, when climate shifted to the net deposition of ice. After the deposition of this airfall mantling unit, the climate shifted to net sublimation resulting in the pitted morphology. Dickson et al. (2009) dated this unit to the mid-Amazonian.

- The inner ejecta blanket and interior of Lyot crater contain small channels and fan deposits (which formed after deposition of the pitted units) for which new data has been collected. Contrary to work conducted by Hobley et al. (2014), the channels are interpreted as forming subaerially via the melting of ice-rich deposits (either mantling deposits or an earlier generation of viscous flow features), forming during
a time of high obliquity and as a result of the unique microenvironment in Lyot crater – agreeing with work conducted by Dickson et al. (2009).

- The morphology of the fan deposits, and the fill analysis conducted on the area within the inner ejecta of Lyot crater, indicate that closed basins could have hosted standing bodies of water within the mid- to late-Amazonian. Such standing bodies of water represent possible habitable environments on Amazonian Mars.

- Before the most recent period of mantle deposition, the clastic polygonal features formed, the exact timing of their formation is uncertain. Morphometric data and observation of these landforms provide evidence, that when compared to morphological analogues, reveal these landforms are most likely the result of the modification of thermal contraction cracks. As thermal contraction generally results from sharp drops in temperature, it is estimated that they formed after the fluvial activity ceased but before the most recent mantle deposits were emplaced.

- Thermal contraction cracks form at a distance of 200-350 km from the crater rim in material exhumed from a depth of 6 - 8 km and could involve an ice component derived from the refreezing of melted pore water from the inner ejecta blanket (according to impact modelling). This material is thus reworked cryosphere material.

- The clastic polygonal networks are proposed to form as a result of the infill of the thermal contraction cracks with wind-blown sediment, and potentially ice, which became cemented or indurated leading to the formation of a resistant-fill. Differential erosion exposed the network, and weathering of the fill formed the angular clasts observed demarcating the borders.

- Mantling units were deposited on top of the older, more degraded units, and represent the latest periods of net ice deposition as a result of climate change driven differences in obliquity. Multiple generations of mantle are present, indicated by layering. Degradation of this mantle has revealed the fluvial channels and clastic polygons beneath them.

- During the deposition of the mantling units, snow and ice also accumulated on slopes forming what are probably debris-covered glaciers (mapped as viscous flow features). These features are interpreted as representing the most recent episode of glacial activity. Previous episodes might have existed which could have provided a source of melt for the channels, though this is uncertain and inferred purely from
the location of the channels beneath the extant viscous flow features. Thus the formation of viscous flow features could have been ongoing and sporadic.

- The most recent activity in and around Lyot crater was the formation of the gullies, potentially as a result of the melting of frost or ice deposits on steep slopes, and aeolian dune fields, which result from the reworking of sand-grade material likely deposited by the fluvial activity.

10.2.2 Key Conclusions

Presented below are the key conclusions of the thesis summarised as single sentence bullet points:

- The Lyot-forming impact likely penetrated the cryosphere and thus outflow channels could be groundwater-related.
- During the impact event, temperatures and pressures within the inner ejecta blanket were sufficient for pore water/ice to be stable as liquid water.
- Channels within the crater interior are not subglacial in origin, but instead form subaerially from the melting of ice-rich deposits.
- Closed basins within Lyot crater could have hosted standing bodies of water within the mid- to late-Amazonian.
- Clastic polygonal landforms around Lyot crater do not result from freeze-thaw processes, rather the modification of thermal contraction cracks.

10.3 Further Work

There are a number of avenues of further research resulting from the work contained within this thesis. Broadly, there are two routes of expansion: (1) further study of Lyot crater, and, (2) using the data and methods employed within this thesis to study other locations for points of further comparison. The latter line of research largely applies to the analysis of the clastic polygonal features conducted in Chapter 7, which has automated the extraction of large amounts of data from polyline shapefiles of polygonal networks. This analysis can be applied to other polygon networks on Earth, Mars and other planetary bodies, to construct a considerable dataset which can be used as a point of comparison for all future polygon studies.

Considering the further study of Lyot crater, this thesis has revealed many more questions. The main questions are presented below.
10.3.1 Was an ice layer(s) penetrated during the impact event and did this effect the morphology of Lyot crater?

Impact modelling of Lyot crater, revealed that the cryosphere in this region should have been penetrated by the impact event. However, these impact models are simplistic; being composed of two layers, and using a granite equation of state as opposed to basalt. Further work could focus on making these models more realistic by improving the basalt equation of state for use in the model, and including a more realistic crustal composition. In particular, it would be useful to include ice layers, and potentially a constraining water layer, within the model to see the effect that has on the morphology of Lyot crater. However, this would be challenging to achieve within the model for two main reasons. First, modelling ice in impact models is complex, so it is not often included (Senft and Stewart, 2008). Second, the resolution of the models limits the thickness of the ice layer that can be modelled. If these challenges are overcome, then the inclusion of ice in the models can provide valuable insights into the effect on the morphology of the crater, and particularly, the ejecta blanket. Previous work by Senft and Stewart (2008) has demonstrated that ice can significantly modify the resulting morphology of a crater; thus if included in the Lyot models, this might aid identification of features that could indicate whether ice was indeed present in the target material, and if water (and ice) was released.

10.3.2 What other evidence is there that outflow channels result from groundwater release during the impact event?

Although the outflow channels around Lyot crater have previously been studied by Harrison et al. (2010), this study assumed that a lack of ponding in the crater interior ruled out a groundwater origin. The work in this thesis indicates that the cryosphere would largely be ejected from the crater interior, and pore water/ice would be vaporised from this region. Furthermore, pervasive, young, mantling deposits are present across the crater interior that would obscure any ancient evidence of ponding in the interior. Therefore, this lack of evidence is not sufficient to rule out a groundwater origin. If the channels do indicate the release of groundwater during the impact event, then this could have important implications for astrobiological potential, as microorganisms from the subsurface could have been flushed to the surface (Westall et al., 2000). Thus, it would be useful to conduct another analysis of the outflow channels to see if there is evidence of a different origin, or at least to explore this hypothesis more.
10.3.3 What are the relative ages of landforms within the ejecta blanket of Lyot crater?

An obvious avenue of future work relating to Lyot crater is to expand the geomorphological map into the ejecta blanket, and eventually beyond the margins of the ejecta, to include features of interest such as the clastic polygonal networks and the extensive outflow channels to the north and east. Using this extended map, the stratigraphic history of Lyot crater could be expanded to include the landforms and surface textures present on the ejecta blanket. As part of this mapping, crater counting could be attempted to give approximate ages for the mapped units, to support the relative ages ascertained in this thesis.

10.3.4 How much water was involved in the formation of the channel and fan deposits, and what was the duration of this activity?

Interpretation of the channel and fan deposits indicate that they formed subaerially and that standing bodies of water could have been present in the crater interior. This is a striking conclusion, and one that deserves further study. Two important questions could be asked: how much water was required, and for how long could it have persisted to form the channel and fan deposits seen? These questions might be answered via a combination of analysis of the drainage area and widths/slopes of the channels (to estimate the volumes of water involved in their incision) and hydraulic modelling based on these estimates. Furthermore, terrestrial analogues can be used to provide a basis of comparison for the channels and fan deposits, to suggest the duration of fluvial activity and whether the volumes calculated are feasible. If the fluvial activity persisted for a significant time period, this has important implications for habitability.

10.3.5 Under what climatic conditions do ice deposits melt in and around Lyot crater, and under what climatic conditions are standing bodies of water stable in the crater interior?

A linked question to that presented in section 10.3.4, is under what climatic conditions do ice deposits melt in and around Lyot crater and standing bodies of water become stable? A study of the microclimatic conditions in Lyot crater could provide important evidence as to whether the subaerial channel-fan-lake hypothesis agrees with current understandings of the extrema in Mars’ recent climate, and potentially help distinguish the timing of the activity, or whether these observations indicate shortcomings in our understanding of the climate. To answer this question, I propose that climate modelling of Lyot crater be conducted using high-resolution models focussing specifically on the Lyot crater study region, and which explore a variety of orbital and other controlling parameters (e.g. total
CO₂ inventory, atmospheric dustiness, etc). In this way the conditions under which fluvial activity is initiated could be ascertained, and this would help distinguish whether or not Lyot crater does have a unique microenvironment suitable for the presence of water in the mid- to late-Amazonian.

10.3.6 Where do the other landforms not studied in this thesis fit in?

A final avenue of further work is the analysis of other landforms not studied within this thesis. These various landforms were identified in Chapter 4 as part of the grid mapping and provide further information about the history of Lyot crater and its surrounding area. These analyses would be conducted in a similar fashion to that conducted in Chapters 6 and 7, where observation and measurements are taken and compared to terrestrial and martian analogues to indicate how they might have formed and where they fit into the history of the area. Landforms of interest for further study include: potential thermokarst terrain (Glines and Gulick, 2018), km-scale polygonal terrain, viscous flow features and gullies in the crater interior, and the boulder fields in the outer ejecta blanket.
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Appendix A

This appendix shows the grid mapping results that were not included in Chapter 5 within this thesis. These plots were not included as they are either used as a form of quality checking, or because there are no significant patterns in the data.
Shapefiles
- Outer Ejecta Extent
- Inner Ejecta Extent
- Inner Peak Ring
- Crater Rim
- Central Peak

Relief Fill
- N
- 2
- 1
- P
- 0

Texture
- N
- 2
- 1
- P
- 0
Appendix B

This appendix presents additional observations about the fan deposits presented in Chapter 6.
<table>
<thead>
<tr>
<th>FID</th>
<th>Deposit Location (m)</th>
<th>CTX Image</th>
<th>Surface Type</th>
<th>Mantle?</th>
<th>Channel Morphology</th>
<th>Channel Width at Mouth (m)</th>
<th>Deposit Morphology</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-102,891.304 2,938,778.823</td>
<td>P03_002270_2303</td>
<td>Mantle. Channel also in flat pitted ground</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Depressed (inverted at very end before 'delta'), slightly sinuous, smooth interior</td>
<td>~84 (just before mantle, difficult to say when mantle unit involved)</td>
<td>Inverted, relatively long, slightly concave outwards. Appears to have a steep edge. Likely thinly covered by mantling unit.</td>
</tr>
<tr>
<td>2</td>
<td>38,726.488 2,998,132.223</td>
<td>P04_002494_2310</td>
<td>On flat pitted ground</td>
<td>Near to a shallow mantle unit</td>
<td>Short depressed sinuous channel. Relatively uniform width. Smooth interior. Slightly inverted end.</td>
<td>~96</td>
<td>Inverted, fan shaped, fractured. Crater at the top of the delta deposit. Concave outwards. Drop off at the end. Thin mantle covering at edges.</td>
</tr>
<tr>
<td>3</td>
<td>-39,477.249 2,867,432.937</td>
<td>D01_027719_2307</td>
<td>Flat pitted ground/mantle</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Depressed, runs alongside VFF, has smaller channel offshoots. Variable width. Inverted at end.</td>
<td>Not obvious</td>
<td>Inverted, relatively smooth. Drop off at end. Concave outwards. Channel facing edge has a jagged shape.</td>
</tr>
<tr>
<td>4</td>
<td>-46,790.349 2,879,259.620</td>
<td>D01_027719_2307</td>
<td>Smooth mantle, covered by aeolian surficial</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Shallow and depressed channel at the top, has smaller contributories.</td>
<td>~124</td>
<td>Inverted, covered by aeolian surficial material. Has an impact crater on the surface. Contains</td>
</tr>
<tr>
<td></td>
<td></td>
<td>material</td>
<td>Changes to an inverted deposit where the smooth mantle unit is. Shows evidence of multiple generations of channel.</td>
<td>evidence of multiple channels.</td>
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</tr>
<tr>
<td>5</td>
<td>-42,637.289 2,879,467.175</td>
<td>Smooth mantle, covered by aeolian surficial material</td>
<td>On top of/underneath (?) mantle unit Depressed channel, relatively sinuous, superposed by aeolian surficial material.</td>
<td>~344 Difficult to distinguish as covered by aeolian surficial material. Appears fractured and arrowhead shaped.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>48,816.235 2,942,968.248</td>
<td>On flat pitted ground Near to a mantle unit Depressed, sinuous channel with smooth interior and small offshoots. Variable width. Fan shaped deposit abuts a second channel.</td>
<td>Not obvious Slightly inverted, fan shaped deposit which is cut off against the second channel. Top of the unit is ruggedly textured. The deposit is not very steep compared to others.</td>
<td></td>
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</tr>
<tr>
<td>7</td>
<td>-49,881.967 2,882,555.608</td>
<td>Smooth mantle, covered by aeolian surficial material</td>
<td>On top of/underneath (?) mantle unit Short inverted channel section with multiple generations indicated. Originates from</td>
<td>~126 Inverted and truncate shaped deposit. Rough textured and appears fractured parallel to length though this is difficult to</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>31,614.423 2,907,326.344</td>
<td>G20_026031_2293</td>
<td>On mantle unit, near to flat pitted ground</td>
<td>On top of/underneath (?) mantle unit</td>
<td>another fan deposit.</td>
<td>Not obvious</td>
<td>Inverted, relatively long, slightly concave outwards. Appears to have a steeper edge. Top of the unit is ruggedly textured. Likely thinly covered by mantling unit. Unit lies on top of a fan shaped smooth mantle deposit which appears to be in a depression.</td>
</tr>
<tr>
<td>-----</td>
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<td>-------------------------------------------------</td>
</tr>
<tr>
<td>9</td>
<td>30,901.568 2,919,760.890</td>
<td>G20_026031_2293</td>
<td>On mantle unit, near to flat pitted ground</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Depressed, slightly sinuous, discontinuous, has small &quot;offshoots&quot;, relatively uniform fitness. Covered by mantle unit at end.</td>
<td>247</td>
<td>Inverted, relatively smooth but with a slight texture. Concave outwards. Potentially a slight drop off at the edge but it is covered by mantle.</td>
</tr>
<tr>
<td>10</td>
<td>56,888.583 2,971,827.242</td>
<td>P04_002494_2310</td>
<td>Flat pitted ground</td>
<td>Near small mantle units</td>
<td>Depressed, sinuous, branching, smooth interior. Variable width, generally narrows downstream. Inverted</td>
<td>Not obvious</td>
<td>Inverted, relatively rough texture, slightly steep gradient at the edge, concave outwards.</td>
</tr>
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</tr>
<tr>
<td>11</td>
<td>-44,104.241 3,028,540.910</td>
<td>P18_008098_2321</td>
<td>On mantle unit, near to flat pitted ground</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Two channels linked to the deposit, depressed, sinuous, discontinuous in parts, some branching of one of the channels, smooth interior, relatively uniform width, inverted at the end.</td>
<td>~56</td>
<td>Inverted, relatively smooth but with a slight texture. Two channels join with this delta. There is a steep drop off at the end and a second shallower drop off.</td>
</tr>
<tr>
<td>12</td>
<td>32,979.034 3,075,022.871</td>
<td>P04_002494_2310</td>
<td>On flat pitted ground, underneath or on top of mantle units</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Shallow and depressed channel with branching.</td>
<td>~229</td>
<td>Inverted, covered with a stippled texture (likely mantle). Has small impact craters on the surface. Has a sudden drop-off at the edge. Lobate in form.</td>
</tr>
<tr>
<td>13</td>
<td>-97,877.158 2,961,704.212</td>
<td>P04_002626_2304</td>
<td>Flat pitted ground/mantle</td>
<td>On top of/underneath (?), directly adjacent to mantle unit</td>
<td>Depressed (inverted at very end before 'delta'), sinuous, smooth interior</td>
<td>~95 (just before inversion)</td>
<td>Inverted, elliptical, fractured, slightly polygonised texture. Not a particularly steep edge.</td>
</tr>
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<tr>
<td>14</td>
<td>81,267.620</td>
<td>P17_007821_2306</td>
<td>Flat pitted ground/mantle</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Depressed shallow channel.</td>
<td>~87</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3,010,227.503</td>
<td></td>
<td></td>
<td></td>
<td>Inverted, lozenge shaped with a relatively smooth surface (covered by mantle). Has a steep drop off at the edge. Appears to be slightly dissected.</td>
<td></td>
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</tr>
<tr>
<td>15</td>
<td>75,693.581</td>
<td>P17_007821_2306</td>
<td>Flat pitted ground</td>
<td>Near mantle units</td>
<td>Depressed, slightly sinuous, smooth interior</td>
<td>~270</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3,039,928.712</td>
<td></td>
<td></td>
<td></td>
<td>Inverted, relatively smooth, slight drop-off at end, predominantly follows channel and lies underneath a younger more concave delta deposit.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>76,404.385</td>
<td>P17_007821_2306</td>
<td>Flat pitted ground</td>
<td>Near mantle units</td>
<td>Depressed, slightly sinuous, smooth interior</td>
<td>~230</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3,040,073.588</td>
<td></td>
<td></td>
<td></td>
<td>Inverted, relatively smooth, Drop off at end, Concave outwards. Potentially multiple generations. Lies on top of delta 1.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>68,627.810</td>
<td>P15_006819_2296</td>
<td>On flat pitted ground/textured mantle</td>
<td>On top of/underneath (?) mantle unit</td>
<td>Barely visible, slight depression, slightly sinuous.</td>
<td>Not obvious</td>
<td></td>
</tr>
<tr>
<td></td>
<td>2,908,684.659</td>
<td></td>
<td></td>
<td></td>
<td>Inverted, circular in shape, slightly pitted with a slight drop off at the edges.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>-74,978.426</td>
<td>P04_002626_2304</td>
<td>On flat pitted ground/smooth</td>
<td>On top of/underneath</td>
<td>Depressed, discontinuous, not</td>
<td>~73</td>
<td></td>
</tr>
<tr>
<td></td>
<td>2,885,376.429</td>
<td></td>
<td></td>
<td></td>
<td>Inverted, relatively elongate and slightly concave outwards. Has a</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number</td>
<td>Coordinates</td>
<td>Image ID</td>
<td>Description</td>
<td>Measurement</td>
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<td>------------------------------------------------------------------------------</td>
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</tr>
<tr>
<td>19</td>
<td>-63,333.098 2,881,111.337</td>
<td>P16_007452_2306</td>
<td>On flat pitted ground/smoother mantle, On top of/underneath (?) mantle unit, Depressed, slightly sinuous, smooth interior. Comes from the end off a VFF.</td>
<td>~72</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>21</td>
<td>-113,286.856 3,068,728.282</td>
<td>P03_002270_2303</td>
<td>On flat pitted ground/smoother mantle, On flat pitted ground/smoother mantle, Depressed and discontinuous. Slightly sinuous.</td>
<td>~86</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22</td>
<td>-98,146.806 2,805,651.808</td>
<td>P03_002270_2303</td>
<td>On flat pitted ground/textured mantle, On top of/underneath (?) mantle unit, Depressed but shallow. Slightly sinuous, split up by depressions with</td>
<td>~141</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Slight drop off at the edges of the deposit. Delta surface is slightly pitted. Inverted, relatively long and is fractured. Inverted, textured, slightly lobate in form. Possible drop-off at edge. Formed at edge of depression. Inverted. Relatively smooth on top due to superposing mantle unit. Long and narrow lozenge shaped deposit. Inverted, rough textured, dissected, lobate in form. Drop-off at edge. Formed when
| 23 | 43,372.982  
2,982,196.936 | P04_002494_2310 | On flat pitted ground | Not particularly close to any mantling units. | Short, depressed, very slightly sinuous channel with a smooth interior. Potentially a short inverted section at the end. | Not obvious | Inverted, arrow shaped deposit. The top of the unit is ruggedly textured with some potential polygonal texture. The deposit is potentially in a depression. |
Appendix C

This appendix presents the complete channel dataset.
<table>
<thead>
<tr>
<th>FID</th>
<th>Z_Min (m)</th>
<th>Z_Max (m)</th>
<th>Z_Mean (m)</th>
<th>SLength (m)</th>
<th>Min_Slope (°)</th>
<th>Max_Slope (°)</th>
<th>Avg_Slope (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-5938.955648</td>
<td>-5040.362025</td>
<td>-5528.883671</td>
<td>9274.535534</td>
<td>1.845168069</td>
<td>18.0924369</td>
<td>9.750442625</td>
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<td>2</td>
<td>-6252.143192</td>
<td>-4799.128585</td>
<td>-5560.863178</td>
<td>18808.04435</td>
<td>1.952919218</td>
<td>19.4303621</td>
<td>9.31509293</td>
</tr>
<tr>
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<td>-6508.9545</td>
<td>-6439.726337</td>
<td>-6490.205662</td>
<td>2646.190672</td>
<td>0.030372682</td>
<td>8.34930335</td>
<td>2.741800535</td>
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<tr>
<td>4</td>
<td>-5705.730229</td>
<td>-4997.751367</td>
<td>-5198.773764</td>
<td>13235.86486</td>
<td>3.226589113</td>
<td>14.97201511</td>
<td>8.407803009</td>
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<td>5</td>
<td>-5551.995594</td>
<td>-4450.063438</td>
<td>-4905.750563</td>
<td>12149.38471</td>
<td>0.020159203</td>
<td>18.12166432</td>
<td>9.096336495</td>
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<tr>
<td>6</td>
<td>-4530.866886</td>
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<td>2579.67943</td>
<td>4.708389438</td>
<td>18.13968572</td>
<td>7.536836153</td>
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<tr>
<td>7</td>
<td>-6216.417077</td>
<td>-5347.840626</td>
<td>-5810.105882</td>
<td>6804.035605</td>
<td>6.863511736</td>
<td>20.31063001</td>
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</tr>
<tr>
<td>8</td>
<td>-6253.462951</td>
<td>-4990.87899</td>
<td>-5551.267458</td>
<td>14010.48596</td>
<td>2.270339329</td>
<td>22.27977783</td>
<td>12.9779404</td>
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<tr>
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<td>-6759.56749</td>
<td>-6563.019648</td>
<td>-6550.216131</td>
<td>8018.676862</td>
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<td>5.342235784</td>
</tr>
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<td>10</td>
<td>-6619.572498</td>
<td>-5961.327455</td>
<td>-6310.260974</td>
<td>11591.38829</td>
<td>0.789277721</td>
<td>12.55132706</td>
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<td>-5196.051235</td>
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<td>7550.409412</td>
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</tr>
<tr>
<td>12</td>
<td>-5240.421001</td>
<td>-4653.205355</td>
<td>-4954.654694</td>
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Appendix D

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