Transient post-glacial processes on Mars: Geomorphologic evidence for a paraglacial period

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A B S T R A C T

On Earth a transitional phase between glacial and interglacial periods is referred to as the paraglacial period. This period immediately postdates glacial retreat and is characterized by ice removal, glacial unloading, and the exposure of steep slopes and large sediment stores. These responses led to the development of a suite of morphologic units (e.g., talus cones, gullies, sackungen, and polygons) which, when observed together, are indicative of the paraglacial period. A similar period of transitional climate and deglaciation is identified on Mars in the Late Amazonian, characterized by the association of features in a glaciated 10.6 km diameter mid-latitude crater. This crater contains concentric crater fill (CCF) formed by debris-covered glaciers, as well as a suite of stratigraphically younger geomorphic units (e.g., spatulate depressions, washboard terrain, gullies, and polygonal terrain) that are all indicative of the local environmental response to deglaciation. These features are interpreted to represent a geologically recent martian paraglacial period within this crater. The morphology and relative stratigraphic relationships among these paraglacial features are described in order to assess the processes operating during deglaciation and to document the recent history of glaciation on Mars: spatulate depressions formed by the differential sublimation of pure glacial ice near the base of the crater wall; subsequently, due to the loss of basal support and steepened slopes, remnant ice on the crater wall began to flow downhill, and formed transverse crevasses that created washboard terrain. Continuous thermal cycling of sediment-mantled ice on crater walls created fractures that formed polygonal terrain. During this time and after, gullies formed by the transport of sediment downslope from crater rim alcoves. Analyses of modeled obliquity variations suggest that the paraglacial period could have operated within the last ~5 Myr and may still be ongoing, suggesting that the current martian paraglacial period is much longer in duration than typical paraglacial periods on Earth. Understanding the nature and sequence of paraglacial activity can help to identify variations in climate in recent Mars history.

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1. Introduction

The climate on Mars in the Late Amazonian is characterized by arid, freezing conditions, broadly analogous to the current conditions in the McMurdo Dry Valleys in Antarctica (Marchant and Head, 2007). During this period on Mars, orbital parameter variations (principally obliquity) have led to repeated latitudinal changes in insolation geometry (Laskar et al., 2004), which in turn have led to variable climate regimes in specific regions; these variations resulted in the repeated migration of snow and ice from the polar caps to the mid-latitudes (Head et al., 2003, 2005; Madeleine et al., 2009), and even to the equatorial regions (Head and Marchant, 2003; Forget, 2006).

Due to the regional climate variation that has characterized Amazonian Mars, comparisons have been made between terrestrial and martian climates with regard to the waxing and waning of ice ages and glacial/interglacial periods (Head et al., 2003). But what geomorphic processes signal the change from glacial to non-glacial conditions? On Earth, the paraglacial period describes the time period in which an environment responds to deglaciation through the rapid transportation of sediment, and during which time a suite of diagnostic geologic features is generated (Ballantyne, 2002a; Mercier, 2008). Mars may regularly experience modification during paraglacial periods, similar to terrestrial environments.

In this work, we present a detailed geomorphologic analysis of an unnamed glaciated 10.6 km diameter mid-latitude crater as evidence of paraglacial modification in the Late Amazonian period on Mars. First, the basic phenomena and processes in terrestrial paraglacial locations are reviewed, as well as the nature of martian...
glaciation, specifically focusing on crater glaciation. The geomorphologic analysis of the crater is then presented as a case study, and the various geologic units are analyzed and assessed as potential paraglacial features. Finally, the paraglacial modification history of the crater is discussed, as well as a comparison of the martian and terrestrial paraglacial periods.

Understanding the nature, characteristic features, and sequence of a paraglacial period on Mars will help us to understand the nature and duration of climate transitions in recent martian history. Specifically, analyses of paraglacial features will help to determine when and where ice loss has occurred, how the local environment responded to this ice loss, where trapped or buried ice may still be present, and how the nature of the paraglacial period can be used to assess and understand climate transitions elsewhere on Mars.

2. The paraglacial period

In terrestrial glaciated settings, the paraglacial period has been defined as “a transitional landscape which is in the process of recovering from the disturbance of glaciation” (Slaymaker, 2009). Initial deglaciation causes an instability or metastability of materials in the environment due to large-scale ice loss, leaving it susceptible to rapid modification (Ballantyne, 2002a). Specifically, this ice loss exposes large sediment stores which become available to erosion and modification by multiple processes. These sediment stores are modified until the supply has been exhausted and the glacially-induced instabilities are accommodated, and the paraglacial period ends when the system returns to a non-glacial or equilibrium state (Fig. 1) (Mercier, 2008). Paraglacial environments differ from periglacial environments which refer to locations proximal to glaciers, ice caps, or ice sheets (and proglacial environments, which refer to locations where processes are dominated by frost action and permafrost); (Ballantyne and Harris, 1994; Benn and Evans, 1998; Slaymaker, 2011).

The paraglacial period is characterized by the combination of geologic processes acting on recently deglaciated terrains (e.g., Figs. 1 and 2). Paraglacial settings include a suite of landsystems, each dominated by one or more erosive and/or depositional processes. Ballantyne (2002a) described several paraglacial landsystems for terrestrial environments: (1) rock slopes, (2) sediment-mantled slopes, (3) glacial forelands, (4) fluvial sediment transport, (5) lacustrine sedimentation, and (6) coastal modification. In this work, we compare the terrestrial paraglacial model to the Mars environment, so we will focus on landsystems that are characteristic of the Late Amazonian (Carr and Head, 2010).

2.1. Paraglacial landsystems: rock slopes

On Earth, the paraglacial rock slope landsystem is characterized by unstable steep slopes that are modified by mass wasting and failure. Recently deglaciated environments are predisposed to modification for two reasons: (1) glacial erosion can act to steepen slopes, which increases the vertical overburden shear stress within the rock; and (2) debuttressing, or removal of glacial ice that supported adjacent rock slopes, causes lateral stress release upon deglaciation (Bovis, 1982; Augustinus, 1995; Ballantyne, 2002a). These vertical and lateral factors can act to create large-scale catastrophic rockslope failures (RSF); smaller-scale, frequent rockfall events can accumulate to form talus debris and talus cones; and slow, large-scale rock deformation can occur via rock mass creep (Fig. 3) (Wyrwoll, 1977; Caine, 1982; Ballantyne, 2002a). Rock mass creep can be manifested as uphill-facing scarps (antiscarp), also known as sackung (plural sackungen) meaning “sagging” (Fig. 3(C)) (Zischinsky, 1966). Sackungen are characterized by complexes of linear extensional features found in bedrock and generally 1–5 m high that lie normal to the topography on steep slopes; proposed triggers range from seismic shaking to sheet jointing, often with preconditioning from lithology or glacial erosion (Zischinsky, 1966; Bovis, 1982; Ballantyne, 2002a; Gutiérrez-Santolalla et al., 2005; Mège and Bourgeois, 2011; McColl, 2012; Makowska et al., 2016). While many disparate formation mechanisms exist for these features, common factors to most formation models are the prerequisites of glacial oversteepening and lateral debuttressing of slopes due to glacial ice removal.

In summary, the paraglacial rock slope landsystem can be characterized by geologic features indicative of mass–wasting and slope failure such as talus cones and sackungen. Features in this landsystem are caused by the failure of destabilized slopes, either vertically through oversteepening, or laterally through debuttressing.
2.2. Paraglacial landsystems: sediment-mantled slopes

Deglaciation exposes sediment-mantled slopes characterized by unconsolidated, unvegetated deposits that are susceptible to rapid modification via debris flow or sediment-gravity flow and the rapid downslope movement of sediment and water (Ballantyne, 2002a). This flow is characterized by mass-movement and failure of sediment, as well as the creation of gullies, or channelized flow paths marked by parallel levees which can cross-cut one another over multiple generations (e.g. Figs. 1(A) and (B) and 2) (Ballantyne and Benn, 1994; Ballantyne, 2002a). Gullies on Earth have been documented to form and develop both through dry sediment mass-wasting off steep slopes and through water-assisted flow, in fluvial-assisted movement, water is mainly sourced from a combination of rainfall and runoff and/or melting of snow accumulated in gully alcoves (Ballantyne and Benn, 1994; Ballantyne, 2002a; Curry et al., 2006; Dickson et al., 2015b). Initially steep slopes, >30° with thick sediment cover (>5 m) favor extensive gully formation and incision (Curry, 2000; Curry et al., 2006). This extensive gully reworking has been shown to decrease overall slope angles by an average of 5° in terrestrial paraglacial settings, with a convergence of modified slope profiles of ~30°, and an overall reduction in slope concavity (Ballantyne and Benn, 1994; Curry, 2000; Ballantyne, 2002a).

In addition to gullies, modification and failure of lateral moraines are common in the sediment-mantled slope landsystem (Fig. 2). These moraines, some of which may be ice-cored, are common in sediment-mantled slope landsystems (Fig. 2). Moraines are frequently modified by slumping, sliding, block falls, etc., triggered by both dry and wet activity (Ballantyne, 2002a; Ballantyne and Benn, 1996; Blair, 1994; Fitzsimons, 1990; Hugenholtz et al., 2008; Johnson, 1971; Mattson and Gardner, 1991; White, 1981). Observations of failure events in these types of ice-cored moraine systems at the Boundary Glacier in Alberta showed that 83% of debris movement was credited to rainfall-triggered failure, while 14% of debris sliding was due to melting of buried ice (Mattson and Gardner, 1991; Ballantyne, 2002a).

In summary, the sediment-mantled slope landsystem can be characterized by features created by sediment transport, most commonly gullies. Features in this landsystem are created by the downslope transportation of exposed unconsolidated sediment, often assisted by the presence of water, although dry sediment transport also occurs.

2.3. Paraglacial landsystems: glacial forelands

The glacial foreland landsystem (Ballantyne, 2002a) includes small-scale processes that affect exposed, unvegetated deposits on valley floors. These include mass movement, freeze-thaw or frost action, and fluvial and eolian processes. Solifluction, very
broadly defined as the small-scale processes of slow mass movement (~1 m/year) (Matsuoka, 2001), can include the slow flow of materials with or without the presence of ice, while gelification refers to internal deformation along buried ice lenses or pore ice (Marchant and Head, 2007). These processes are characterized by high soil moisture content due to the warmer conditions accompanying deglaciation which promotes ice lens growth; sediments are then susceptible to modification from thermal cycling and meltwater saturation. Thermal cycling is also responsible for the formation of polygonal fractures and ice-wedge polygons which are seen in many periglacial environments, including the arid, polar deserts of the Antarctic Dry Valleys (Lachenbruch, 1962; Marchant et al., 2002; French, 2007; Marchant and Head, 2007). Additional glacial foreland processes include eolian processes such as deflation (winnowing of fine sediment from surficial deposits, resulting in a blocky lag deposit) that remove and transport unconsolidated sediments, particularly in the early stages of the paraglacial period prior to stabilization by vegetation (Ballantyne, 2002a).

In summary, the glacial foreland landsystem is characterized by geologic features created through small-scale transportation of sediment or ice, creating features such as polygons or solifluction lobes. Features in this landsystem are formed through mass wasting, thermal cycling, and eolian processes.

### 2.4. Paraglacial chronology

The paraglacial period initiates and operates in a diagnostic manner (Ballantyne, 2002a; Ballantyne and Benn, 1996) (Fig. 1): deglaciation exposes unconsolidated sediment on steep slopes which are then immediately modified (via RSF, sapping, and gully formation) (Fig. 1(A)). As paraglacial modification continues, rock slopes shed debris and form talus, sediment-mantled slopes erode and form gullies, and glacial forelands are modified by thermal cycling (Fig. 1(B)). Eventually, sediment stores are exhausted and the rate of sediment transport declines, valley wall slopes have decreased and stabilized, and gullies have become dormant and vegetated (Fig. 1(C)). Paraglacial modification effectively halts, and sediment transport rates reach non-glacial or equilibrium conditions. At this point the paraglacial period is over.

Each landsystem within the paraglacial period functions on a distinct time scale that commences immediately following the initiation of deglaciation (Fig. 4). For example, the period of rock slope failure can last 1–2 orders of magnitude longer than that of glacial foreland modification (Ballantyne and Matthews, 1982; Cruden and Hu, 1993; Ballantyne, 2002b; McColl, 2012). Therefore, to determine when the paraglacial period has ended, the sediment transport rate and paraglacial feature development rate of the entire region must be assessed.

In summary, the terrestrial paraglacial period is described and identified by a suite of landsystems that collectively describe the manner in which an environment responds to deglaciation. Each landsystem acts independently, over a distinct timescale, but collectively paraglacial modification acts via sediment transport to return an environment to nonglacial “equilibrium” conditions.

### 3. Martian glaciation

Glaciation has occurred at virtually all latitudes on Mars throughout the Amazonian period: polar caps contain significant reservoirs of water ice in the form of polar layered deposits (PLD) (e.g., Picardi et al., 2005; Plaut et al., 2007); extensive evidence suggests that Martian mid- to high-latitudes experienced periods of glaciation, including the widespread population of geomorphic features (Squyres, 1979) including concentric crater fill (CCF) (Squyres and Carr, 1986; Dickson et al., 2010; Levy et al., 2010), lobate debris aprons (LDA) (Pierce and Crown, 2003; Mangold, 2003; Holt et al., 2008; Plaut et al., 2009), linedate valley fill (LVF) (Head et al., 2006, 2010), viscous flow features (Milliken et al., 2003), and latitude-dependent mantle (LDM) (Kreslavsky and Head, 2000, 2002; Mustard et al., 2001; Head et al., 2003). Equatorial regions also show evidence of glaciation in the fan-shaped deposits (FSD) around the Tharsis Montes (Head and Marchant, 2003; Shean et al., 2005). Persistent low temperatures currently and throughout the Late Amazonian imply that glaciation was cold-based (e.g., Head and Marchant, 2003). Mars general circulation models (GCMS) (e.g., Forget, 2006; Madeleine et al., 2009, Madeleine et al., 2014) show that the latitudinal redistribution of snow and ice is driven primarily by changes in obliquity, and glacial flow models (Fastook et al., 2011; Fastook et al., 2014; Fastook and Head, 2014) show how GCMBased accumulation rates map out into the observed array of glacial landforms.

#### 3.1. Martian glaciation: concentric crater fill

As mentioned above, many morphologic features attributed to the accumulation and flow of ice are concentrated in the mid-latitudes, formed during periods of higher mean obliquity (~35°). This work focuses on glaciated craters, or craters that contain CCF, to search for paraglacial features. As such, recent studies regarding the nature of CCF are briefly discussed below.

CCF is generally identified as a unit containing multiple rings or concentric lineations within the crater interior (Levy et al., 2010; Squyres, 1979). Previous authors have studied CCF and have concluded that these deposits represent relict debris-covered glaciers (Levy et al., 2010; Fastook and Head, 2014) based on the documentation of glacier-like flow features (Head et al., 2006, 2010), the presence of ring-mold craters (indicative of impacts into ice) (Kress and Head, 2008), and the discovery of nearly pure ice within LDA and LVF deposits (Holt et al., 2008; Plaut et al., 2009) using SHARAD data. Furthermore, filled crater depth-diameter ratios show that in many locations hundreds of meters of ice are likely to still be present under desiccated CCF surficial debris (Levy et al., 2009b). Analyses of breached craters and distal glacial deposits suggest CCF-related ice was at one point several hundred meters higher than its current level, and this ice has sublimated in the recent past (Kadish et al., 2010; Levy et al., 2010).

Fastook and Head (2014) assessed the specific glacial environment in which CCF formed (using characteristics of the CCF and the polar caps) and also used ice rheology to assess the maximum thickness of regional ice deposits in the CCF regions during the Late Amazonian. This study found that glaciation was cold-based and impact craters are by far the most abundant landform with associated slopes (interior wall and exterior rim) sufficiently high.
to induce glacial ice flow under Late Amazonian climate conditions; they then documented with topographic slope data that impact craters have been clearly modified, undergoing crater interior slope reduction and floor shallowing. Fastook and Head (2014) concluded that these trends are the predictable response of ice deposition and preferential accumulation and retention in mid- to high-latitude crater interiors during episodes of enhanced obliquity. Finally, they used a representative obliquity solution to drive an ice flow model and showed that a cyclical pattern of multiply recurring layers can both fill the craters with a significant volume of ice, as well as transport debris from the crater walls out into the central regions of the craters. They predicted that the cyclical pattern of waxing and waning of mantling layers would result in a rippled pattern of surface debris extending out into the crater interiors that would manifest itself as an observable concentric pattern, comparable in appearance to CCF. In their scenario, the formation of mantling sublimation till layers seals the accumulating ice and sequesters it from significant temperature variations at diurnal, annual, and orbital parameter cycle time scales.

Fastook et al. (2014) found that as a regional ice sheet collapses, the surface is predicted to drop below cliff and massif bedrock margins, exposing bedrock and regolith, and initiating debris deposition on the surface of a cold-based glacier. Reduced sublimation due to debris-cover armoring of the proto-LDA surface produces a surface slope and consequent ice flow that carries the armoring debris away from the rock outcrops. As collapse and ice retreat continue, the debris train eventually reaches the substrate surface at the front of the glacier, leaving the entire feature armored by debris cover. These treatments thus provide context for the role that CCF plays in the formation, collapse, and decay of mid-latitude glaciation, and in the analysis of the paraglacial period on Mars.

4. Geomorphic analysis of a recently glaciated crater interior

Over 600 glaciated craters are present in the mid-latitudes of Mars (Dickson et al., 2012). To perform a detailed analysis of the presence and nature of the martian paraglacial period, an unnamed 10.6 km-diameter crater in eastern Newton basin (155.3°W, 40.13°S, Fig. 5) was selected that had previously been identified as containing evidence of glacial deposits as well as stratigraphically younger features including gullies and spatulate depressions (Head et al., 2008). Analyses were made using high-resolution images from Context Camera (CTX) (Malin et al., 2007) and High-Resolution Imaging Science Experiment (HiRISE) (McEwen et al., 2007), and altimetry data from Mars Orbiter Laser Altimeter (MOLA) (Smith et al., 2001).Crater walls are typically characterized by talus slopes and slump terraces, but this crater also displays an asymmetry (Dickson et al., 2012) due to well-developed gullies around the northern crater rim, as well as more typical large wall slumps on the southern crater wall. The floor topography slopes poleward by ~3-5° across the crater floor (Fig. 5(B)).The sloping floor is due to the large-scale lobate fill, extending from the base of the northern portion of the crater wall to the southern portion, where the termination of the flow lobes extend below the southern crater wall (Fig. 5(A)) (Head et al., 2008). The crater fill is characterized by concentric arcuate ridges which initiate at the base of the northern crater wall, and become deformed and linear towards the southern crater wall (Fig. 6). The stratigraphic relationship of this concentric crater fill (CCF) with the host crater indicates that it is younger than the crater, yet older than several other geologic units present on the rim, walls, and floor of the crater that are su-
perposed on or incise into the CCF. These stratigraphically younger geologic units (Head et al., 2008) include (Fig. 6) spatulate depressions, washboard terrain, gullies, polygonal terrain, broad pits, and pitted terrain. Each of these late-stage units are described and assessed in the context of the paraglacial framework.

4.1. Spatulate depressions

4.1.1. Observations

At least fifteen arcuate spatulate depressions are located at the base of the crater wall along the northern portion of the crater (Figs. 6 and 7). The largest of these depressions are on the northern, pole-facing side of the crater (Head et al., 2008). Depressions are rimmed by a curved outer ridge extending up to 1500 m from the base of the crater wall, and often contain a sharp, steep-sloped inner ridge (Fig. 7(B)). The floors of large depressions are flat with a fine-scale hummocky texture with visible blocks smaller than ~2 m in diameter. The spatulate depressions are on average ~265 m wide on the short axis and ~470 m in length on the long axis, the largest of which is 450 m wide and 600 m long (bounded by the inner ridge) (see Fig. 7(C)). HiRISE Digital Elevation Model (DEM) data show that the spatulate depressions are ~25–50 m deep, corresponding to an area and volume of the largest spatulate depression of ~0.3 km² and ~0.01 km³, respectively. The spatulate depressions have a well-preserved morphology in the sharp inner ridges and flat, hummocky floors; in addition, some spatulate depressions also show evidence of more recent modification in the form of washboard terrain-like fractures (e.g., Fig. 7(F)) and gully sediment fans which have accumulated inside and adjacent to the spatulate depressions. Spatulate depressions are most often spatially associated with crater rim alcoves higher on the crater walls (Head et al., 2008) – within this crater, alcoves present in the crater rim can be observed upslope of each of the largest spatulate depressions (Fig. 6). In addition, it has been proposed that the shape of the arcuate ridges surrounding the spatulate depressions can mimic the geometry of the alcoves (Berman et al., 2005; Head et al., 2008).
Based on the observations of the spatulate depressions and the associated features in the crater, the spatulate depressions are stratigraphically younger than the majority of the CCF, as they have formed within the crater filling unit; additionally, the spatulate depressions are stratigraphically older than the gullies and the washboard terrain, as these two units have both formed on or have incised into the spatulate depressions.

In summary, the spatulate depressions are large, flat-floored depressions at the base of the crater wall. This unit is spatially associated with stratigraphically older CCF and younger washboard terrain and gullies. Spatulate depressions are also associated with crater rim alcoves higher on the crater wall.

4.1.2. Interpretations

The similarity in shape and orientation of the spatulate depressions to the ridges in the CCF (Figs. 5 and 6) suggest that they are related to the flow of glacial ice onto the crater floor; however, the depressions are interpreted to have formed via the sublimation of exposed snow and ice at the base of the crater wall, forming a depression (Head et al., 2008).

To assess in further detail the mechanism of spatulate depression formation within the crater, it is useful to review a recent analysis of debris-covered alpine glaciers in the Antarctic Dry Valleys (Mackay et al., 2014). Mullins and Friedman glaciers (Fig. 8) can be viewed as analogs for martian debris-covered glaciers due to similar climate settings (Marchant and Head, 2007) as well as morphologic similarities such as patterns of ridges that become increasingly deformed downslope. The proposed formation and evolution of these cold-based debris-covered glaciers is therefore relevant to the discussion of martian CCF and spatulate depressions, and will be briefly reviewed: Mackay et al. (2014) used ground-penetrating radar and surface morphologic analyses to study the structure of Mullins and Friedman glaciers. These glaciers contain mostly pure glacial ice (~1% englacial debris) and are overlain by a supraglacial debris layer (~8–75 cm thick). Both glaciers are interspersed with englacial bands of debris that intersect the surface and form arcuate ridges. The interpreted formation mechanism for these englacial debris bands and arcuate surface discontinuities is as follows: during periods of reduced net ice accumulation, differential sublimation would cause ice to ablate near valley headwalls, forming “spoon-shaped hollows” that would become mantled by a rocky lag; at the point that net ice accumulation renews, the lag would become buried by snow and ice which would accumulate and flow into the hollows – where the sublimation lag layer intersects the ice surface, a raised arcuate ridge would remain exposed, and a band of buried englacial debris would persist under the ice (see Fig. 17 in Mackay et al., 2014). This process would repeat over multiple climate cycles to form the patterns of ridges seen on the Mullins and Friedman glaciers (Fig. 8). This model implies that the englacial structure and surface morphology present in these glaciers provides a record of climate variation over the past 10^4–10^6 years (Mackay and Marchant, 2017).

The morphologic characteristics of the spatulate depressions and their relationship to the CCF in the martian crater suggest that a similar process could be operating as that documented at Mullins and Friedman glaciers; the patterns of concentric ridges present in CCF deform as they extend downslope and encounter topographic obstacles. These can be seen in both martian glacial features (e.g., Head et al., 2010) as well as in the Antarctic glaciers (compare Figs. 5–6 and 8). In addition, the location, morphology, and stratigraphic age of the spatulate depressions is consistent with a formation mechanism similar to that proposed by Mackay et al. (2014), where differential sublimation near the headwall leads to the formation of a depression. Therefore, the proposed model for spatulate depression formation is as follows (Fig. 9): (I) snow and ice accumulate on the crater rim and wall during a period of net accumulation, leading to the flow of ice downslope (Fastook and Head, 2014); (II) the climate shifts to a period of net ablation –
minimal morphologic evidence exists indicating the presence of meltwater, so ablation is expected to have occurred predominantly via sublimation. Exposed ice on the crater rim and walls would sublimate, exposing crater wall material that could be shed off steep slopes and could act as a source of debris cover for the ice. The exposed ice surface would lower, forming a broad depression near the base of the crater wall and concentrating a rocky lag on top of the ice (armorimg it from further sublimation). (III) In the next period of net accumulation, snow and ice would again accumulate in the crater and flow downslope, filling in the depression and leaving a raised ridge. In this model of cyclic ice accumulation, sublimation and depression formation, and subsequent accumulation, sequences of debris-covered, ice-cored ridges will accumulate within the crater, in a similar mechanism as has been proposed in the Antarctic debris-covered glaciers (Mackay et al., 2014).

In this model of spatulate depression formation, debris would be sourced only from material shed off the exposed crater rim and walls. As a result, this model predicts that spatulate depressions would only be preserved in settings where the debris supply (i.e., headwall erosion rate) is low; a steady supply of debris would lead to more rapid arming of exposed ice, and therefore less differential sublimation in the area of the spatulate depression. The fact that there are several large, well-developed spatulate depressions in this crater suggests that the debris supply rate must have been low. This interpretation is supported by the overall circular appearance of the crater shape (Fig. 5) – rapid erosion of the northern crater rim would have created a highly asymmetric crater shape not observed in this crater.

The formation process for the martian spatulate depressions fits within the paraglacial framework, as these features are evidence of the initial stages in deglaciation required to trigger the paraglacial period. The transition from a net accumulation to net ablation climate regime on Mars would initiate a loss of ice from the walls and floors of craters, creating spatulate depressions as a result of this ice loss. The ice loss introduces structural instabilities due to debuttressing of ice on the crater wall, and also exposes unconsolidated sediments and sublimation till on the crater wall that was previously mantled by snow and ice. Therefore, the formation of the spatulate depressions can be directly linked to the formation of other paraglacial features identified in the crater, discussed below.

4.2. Washboard terrain

4.2.1. Observations

As noted by previous studies (Head et al., 2008), the eastern portion of the crater wall immediately above the spatulate depressions contains a laterally continuous zone of deformation characterized by sets of parallel scarps referred to in this work as “washboard terrain” due to its similarity in appearance to a washboard (Figs. 6 and 10). The fractures that make up washboard terrain are normal to the crater wall slope and are uphill-facing in orientation. The zone of deformed crater wall extends from the approximate head of the gully sediment fans, downslope to the base of the crater wall (Fig. 6 and 10(B)). This deformed region extends vertically over an average distance of ~265 m, although this height varies with location in the crater from ~65–420 m. The most extensive deposits of washboard terrain are located on the crater wall and extend to the base of the wall, to the edge of the spatulate depressions; however, parallel fractures that are morphologically consistent with washboard terrain also extend onto the crater floor, between or into spatulate depressions (Fig. 7(E) and (F)). The portion of the crater wall containing washboard terrain is distinct in texture from the crater wall immediately upslope (Fig. 10(C)) and the transition is indicated by a laterally continuous, downhill-facing scarp that extends for ~4400 m around the crater wall (Fig. 10(A) and (C)).

The parallel uphill-facing scarps typical of washboard terrain extend downslope from this laterally continuous downhill-facing scarp to the crater floor, with an average spacing of ~12 m between each scarp (see histogram in Fig. 10(D)). Individual scarps extend laterally for ~100 m, although many scarps have been mantled by younger gully sediment fans (Figs. 10(B) and 11) and therefore could be more laterally extensive (Fig. 11). The scarps do not increase in size or spacing down- or across-slope.

The washboard terrain is generally stratigraphically older than the gully sediment fans, as the youngest generations of gully sediment fans superpose the washboard terrain. However, portions of the gully fans have been visibly modified and cross-cut by fractures in the washboard terrain (Fig. 11(C)). In addition, chains of small pits are observed in the gully fans (Fig. 11, black arrows). These pits are ~3 m in diameter, separated from each other by ~4 m, and are aligned in parallel sets that are of the same orientation and lateral scale as the washboard terrain fractures. The laterally extensive, downhill-facing scarp is stratigraphically consistent with the rest of the washboard terrain, as it is also cross-cut by young gully channels and fans. As noted above, fractures consistent with washboard terrain are also observed within several spatulate depressions, as well as on the ridges of the depressions. We interpret this relationship to indicate that the washboard terrain is stratigraphically younger than the spatulate depressions, and formed as a post-emplacement feature after the spatulate depressions.

In summary, the washboard terrain is composed of parallel sets of downhill-facing scarps that are present on large swaths of the lower portion of the crater wall. These fractures are stratigraphically younger than the spatulate depressions, and are stratigraphically older than the youngest generation of gully sediment fans.

4.2.2. Interpretations

Characteristics of the washboard terrain can be used to create a formation model, including (1) morphology, parallel or sub-parallel well-organized fractures normal to the slope, which we interpret to have formed in an extensional stress regime that led to downslope motion of material, rather than closely spaced horst and graben morphology (Fig. 10(D)); (2) distribution on the lower portion of the crater wall and on portions of the crater floor inside and between spatulate depressions, suggesting a continuous formation process operating over this entire region; and (3) stratigraphic relationships that indicate the washboard terrain formed after the period of spatulate depression formation, and before (but interspersed with) the period of gully formation. The formation of the washboard terrain fractures therefore had to occur in a relatively coherent manner over a large spatial area over a relatively long period of time.

The downslope motion of material necessary to create the washboard terrain scarps could have been triggered by debuttressing (removal of basal support of wall materials) via the formation of spatulate depressions; the removal of ~20–50 m of ice or a volume of ~0.01 km³ of ice in the formation of the largest spatulate depression would act to steepen crater wall slopes from ~19–23°. In addition, during this period of spatulate depression formation, the overall ice surface within the crater would have been lowering due to ice loss on the CCF surface through sublimation and eventual arming by debris. Crater wall steepening would occur as a combination of these two processes and would induce stresses within the crater environment. These new stresses could cause remnant ice on the crater wall to renew flow downslope; as the ice flowed down the crater wall, the principal stress would have been in the downslope direction. The flow of ice could form fractures parallel to the slope, resulting in the formation of transverse crevasse-like features (Fig. 12). In terrestrial glaciers,
transverse crevasses form due to extending flow of ice in glaciers where the principal tensile stress is parallel to glacier flow, leading to crevasses opening up at right angles to the glacier center line (Benn and Evans, 1998). We interpret the fractures seen in the washboard terrain to be indicative of remnant crevasses formed due to this reinvigorated flow of ice downslope caused by debuttressing.

The widespread distribution of washboard terrain across the crater wall, not only in regions immediately adjacent to spatulate depressions (Fig. 6) suggests that the combined debuttressing effects of the formation of spatulate depressions and the net lowering of the CCF surface serve to induce washboard terrain formation.

The distribution of the washboard terrain is clearly identifiable; within the crater, fractures are densely concentrated on the lower portion of the crater wall, and a distinct textural contrast is apparent in regions containing washboard terrain and those that do not (Fig. 10(C)). In addition, the uppermost region of washboard terrain is outlined throughout the crater by a distinct arcuate, downhill-facing scarp (Fig. 10(A) and (C)), above which no fractures are ap-
parent (Fig. 11). If we adopt the proposed formation model for the washboard terrain forming via glacial crevassing, the portions of the crater containing washboard terrain indicate the presence and distribution of a recently mobile, deformable layer of subsurface ice only in these portions of the crater. In addition, the characteristics of the downhill-facing scarp, namely the location, morphology, and apparent boundary between washboard terrain and fracture-free crater wall, all indicate that this scarp may represent a bergschrund, a deep transverse crevasse that is believed to separate relatively immobile ice at the glacier head from mobile ice below (Benn and Evans, 1998). Bergschrunds are conspicuous, concave features that may be laterally continuous or short and staggered and are often crescent in shape (Mair and Kuhn, 1994). Field analyses of terrestrial bergschrunds suggest that the geometry of bedrock is a major control on the nature, location, and width of a bergschrund and subsidiary crevasses (Osborn, 1983). The arcuate scarp representing a bergschrund within the crater would explain why no washboard terrain fractures are observed upslope, while such a high concentration of fractures is present downslope of the scarp, as this downslope region corresponds to the region of deforming ice. Other potential identifications of bergschrunds have been proposed on Mars in mid-latitude glacial deposits (e.g., Head et al., 2006; Hubbard et al., 2014).

Stratigraphically, the washboard terrain is generally younger than the spatulate depressions, although some washboard terrain-like fractures are present within spatulate depressions (Fig. 7E and F). In addition, washboard terrain is on average older than gully formation, but the older generation of gullies are interspersed...
4.3. Polygonal terrain and broad pits

4.3.1. Observations

Polynomials and patterned ground are present in a wide range of settings across the martian surface (not restricted to crater interiors) predominantly at mid- to high latitudes, and generally signify the presence of ground ice (Mangold, 2005; Levy et al., 2009a). In the crater analyzed in this work, polygonal terrain is visible on large portions of the northern crater wall (Figs. 6 and 13). Crater wall polygons were measured on an area of ∼250 m by 150 m (37,500 m²). These polygons range from 1.2 m to 20.3 m in diameter (Fig. 13) and are on average ∼9 m in diameter, with fractures ∼1 m in width (see histogram in Fig. 13(C)). The polygons in this area in the crater in general have a convex-up shape (also described as peak-top polygons by Levy et al., 2009a).

In a portion of the crater wall adjacent to a broad pit a small portion of polygonal fractures are aligned in the downslope direction, resulting in a linear pattern of polygon fractures (Fig. 13(B)). The polygons in this aligned region do not appear to be elongate in the downslope direction, and the fractures appear to intersect at close to 90° angles.

The largest expanse of polygonal terrain in the crater extends 2.6 km across the entire northern portion of the crater wall, and ∼500 m down slope (Fig. 6). This western part of this portion of the crater wall has also been modified by gullies and a broad pit forming within the polygonal terrain. Most gully channel interiors are free of polygons, although there are portions of several gullies in the crater that are polygonalized. Polygons are also found inside other broad pits (Fig. 13(B)). The polygons located inside the broad pits are similar in morphology and size to crater wall polygons (polygons within the broad pit in Fig. 13(B) are on average ∼9 m in diameter, as seen in the histogram in Fig. 13(C)) and do not appear to be preferentially oriented downslope.

Broad pits are present in three locations in the northern section of the crater wall (Fig. 6). Located in the crater wall, these pits are rimless oblong depressions that are approximately 120–250 m long and 50–90 m wide, and are oriented with their long axes downslope (Figs. 6 and 13(B)). Broad pits are only found on portions of crater walls that are polygonalized, and the interiors of the broad pits also contain polygons (Fig. 13(B)). One of the broad pits is located in a polygonalized gully channel (e.g. Fig. 14(B), yellow arrow), while the other two are located in flatter, gully-free expanses of the crater wall (Fig. 6). The broad pits are located immediately upslope of several of the largest spatulate depressions in the crater (Fig. 6).

Stratigraphically, the broad pit present in a gully channel (Fig. 14(C)) has been cross-cut by younger gully channels, indicating that the period of broad pit and polygon formation in this region occurred before the youngest generation of gully formation; however, a gully to the west of this broad pit that does not appear to have experienced sediment transport through its channel recently (Fig. 14(B)) contains polygons in the gully alcove and channel. This stratigraphic relationship suggests that both polygon formation and gully erosion occurred within the same general period, although the youngest gully activity has occurred more recently than polygon formation.

In summary, the polygonal terrain and broad pits are present on the northern portion of the crater wall. The polygons are small-scale features bound by fractures and are widely distributed across the crater wall, while broad pits contain polygons and are only present in three locations in the crater. Stratigraphic relationships suggest that polygon and broad pit formation occurred after the onset of gully formation, but ceased before the most recent gully modification occurred.
4.3.2. Interpretations

Terrestrial polygonal patterned ground can form through several mechanisms which are dependent largely on the climate and substrate (e.g., Black, 1976). One such type, sublimation polygons, forms when there is excess ice (ice exceeding the pore space) in the near-surface (Marchant et al., 2002; Marchant and Head, 2007). Due to periodic thermal cycling (expansion and contraction driven by insolation variations that change on diurnal, seasonal, and orbital time scales) of this ice-rich permafrost or buried ice, fractures form in the ice and become infilled by sediment. These polygons are present in widespread periglacial environments (Lachenbruch, 1962; French, 2007) as well as the hyperarid polar desert of the Antarctic Dry Valleys (Marchant et al., 2002; Marchant and Head, 2007). Similarly, polygons have been identified in many locations on Mars (Mellon, 1997; Mangold, 2005; Levy et al., 2009a; Gallagher et al., 2011) and many of these have been identified as sublimation polygons (Levy et al., 2009a). The polygons in this crater appear to be convex-up, similar in morphology to sublimation polygons rather than to typical sand-wedge or ice-wedge polygons (Marchant et al., 2002; Marchant and Head, 2007). We interpret the polygons described in this crater to be indicative of the presence of near-surface, buried ground ice. The portion of aligned polygons (Fig. 13(B)) may indicate post-emplacement deformation via downslope movement or deformation of the crater wall materials subsequent to polygon formation, although the evidence of this alignment is spatially limited.

Due to the spatial association between polygonalized crater wall and broad pits (as well as the presence of polygons inside broad pits), the pits are interpreted to form via a similar mechanism, namely thermal cycling and sublimation of a subsurface ice layer. Differential sublimation of a thicker subsurface ice layer may have created the depression; as the broad pits are located in polygonalized terrain, the differential sublimation may initiate at the location of a polygon fracture where ice was exposed (Marchant et al., 2002; Marchant and Head, 2007). The broad pits are located immediately upslope of the largest spatulate depressions. As the spatulate depressions are interpreted to represent regions of past large-scale ice accumulation and flow, the locations of broad pits are also expected to have been characterized by large quantities of ice in the geologically recent past, which may explain why there was a thicker layer of ice in this region that underwent differential sublimation relative to the surrounding area.
Based on the interpretation that the polygons within the crater are sublimation-type polygons and formed via the presence of subsurface ice, the presence and distribution of crater wall polygons and broad pits can be used to identify near-surface ice reservoirs that indicate areas where ice was abundant during the last glacial period. Using sublimation polygons as indicators of subsurface ice predicts that a large portion of the northern region of the crater contains a subsurface ice layer, which also corresponds to the locations of the largest spatulate depressions (Fig. 6) – this observation is evidence that the northern portion of the crater was the major accumulation area for snow and ice in the crater.

Broad pits are morphologically similar to the scalloped terrain (Lefort et al., 2009). Scalloped terrain is formed of broad, rimless, flat-floored depressions present in thick ice-rich mantles in the
southern hemisphere of Mars, concentrated in the southern walls of Hellas Basin and northern Malea Planum (Zanetti et al., 2010). Scalloped depressions are believed to form via solar-insolation driven sublimation, due to the sublimation of interstitial ice from increasingly large cracks in an ice-rich mantle (e.g., Zanetti et al., 2010). Similarities in regional setting, morphology, and association with sublimation-type polygons and ice-rich substrates, suggest a similar formation mechanism for broad pits as for scalloped depressions.

Based on the interpreted formation mechanism of the broad pits and the polygons via thermal cycling and buried ice sublimation, and the stratigraphic observations between the broad pits and the polygons relative to other features in the crater, the polygonal terrain and broad pits fit into the paraglacial framework: stratigraphically, the polygons form after the initial period of washboard terrain formation, and after the older generation of gully formation, and continues until just before the cessation of gully activity, as morphologically young gully activity is observed to cross-cut broad pits and polygons.

4.4. Gullies

4.4.1. Observations

Martian gullies are well-documented features (Malin and Edgett, 2000; Treiman, 2003; Balme et al., 2006; Dickson et al., 2007, 2015a; Diniega et al., 2010; Dundas et al., 2010, 2012, 2015; Harrison et al., 2015; Pilorget and Forget, 2016; Auld and Dixon, 2016). Originally defined as consisting of three morphologic features: alcoves, channels, and fans (Malin and Edgett, 2000), gullies are believed to have formed at various points in the Amazonian (Dickson et al., 2015a). Traditional gullies have been observed to be concentrated mainly in the mid- to high-latitudes of Mars (Auld and Dixon, 2016; Harrison et al., 2015), although other morphologic classes of gullies have also been defined, such as gullies with no channels (only alcoves and fans), gullies with no alcoves (only channels and fans), and linear gullies (no alcoves or fans) (Auld and Dixon, 2016). More gullies have been identified in the southern hemisphere than in the northern, with the highest concentration of gully density occurring in the southern hemisphere in the Terra Cimmeria/Terra Sirenum region (Auld and Dixon, 2016; Harrison et al., 2015).

Within the crater analyzed in this work, at least 15 distinct, well-developed gullies were mapped (Figs. 6 and 14) and are distributed across the northern and eastern crater wall, in addition to several less well-developed gullies distributed throughout the crater. These gullies are composed of the traditional tripartite structure including an alcove near the crater rim, (Fig. 14(D)) a channel or system of channels (Fig. 14(E)), and a depositional sediment fan (Fig. 14(F)). Many channels intersect and cross-cut one another, vary in depth and width, and can be traced from the same source alcove.

The gullies vary in stratigraphic age: the stratigraphically youngest gullies show well-defined channel boundaries often with large, distinct sediment fans, while stratigraphically older gullies show poorly defined channels (compare black and red arrows in Fig. 14(B)) and small, diffuse, or mantled sediment fans. A broad pit is visible being cross-cut by the channel of a stratigraphically young gully (Fig. 14(C)), and stratigraphically older gullies have alcoves and channels that have become polygonalized (Fig. 14(B)). Polygons in the channels and alcoves display the convex-up morphology indicative of sublimation polygons (Marchant et al., 2002), similar to the “gullygons” described by Levy et al. (2009a) and morphologically similar to polygons seen elsewhere in the crater. In addition, gullies are generally stratigraphically younger than washboard terrain, although older generations of gully sediment fans have been cross-cut by washboard terrain fractures (Fig. 11).

In summary, the gullies present in this crater conform to the traditional gully morphology containing an alcove, channel, and fan. Gullies are concentrated on the northern and eastern portion of the crater rim and are generally stratigraphically young, although evidence for multiple stages of gully activity is indicated. Relationships to other features in the crater suggest that gully activity occurred over a period of time that overlapped with multiple other processes in the crater, including polygon and washboard terrain formation.

4.4.2. Interpretations

Terrestrial gullies in the Antarctic Dry Valleys as well as in more temperate regions are understood to form by a combination of dry sediment mass-wasting from steep slopes and water-assisted flow, the water sourced mainly from a combination of rainfall (in temperate climates) and melting snowpack in depressions at gully heads (in cold climates) (Ballantyne and Benn, 1994; Ballantyne, 2002a; Curry et al., 2006; Decauline and Søemundsson, 2007; Marchant and Head, 2007; Dickson et al., 2015b).

Much debate has surrounded the mechanism by which gully formation occurs on Mars; the initial interpretation from Malin and Edgett (2000) proposed a liquid water origin via the release of water from shallow aquifers. However, more recent models of gully formation include a variety of mechanisms including CO2-related processes (Musselwhite et al., 2001; Dundas et al., 2010; Pilorget and Forget, 2016), dry mass wasting (Treiman, 2003), the melting of ice from earlier periods (Dickson et al., 2015a), or a combination of factors (Vincentdon, 2015).

The gullies in this crater have a clear spatial and stratigraphic link with other features including polygons, washboard terrain, and spalutate depressions (Fig. 6). These features are all interpreted to indicate either the past presence of large quantities of ice, or the current extent of buried ice in the crater (Dickson et al., 2017). Given the abundance of ice indicated by these other features within the crater, as well as the close spatial association to the regions where gullies are present, the favored model for gully formation here is one in which water-assisted flow played a role in forming the gullies in this crater. It is important to note that this does not preclude the role of dry and CO2-assisted flow in gully formation – rather, it is expected that these processes would have aided in creating the gully morphology seen currently.

As mentioned above, the alcove located upslope of the spalutate depressions are interpreted to represent the areas of major snow and ice accumulation in the crater (Head et al., 2008). Therefore, even after the period of ice accumulation has ended, windblown snow could have become trapped in these gully alcoves (Dickson et al., 2017; Head and Marchant, 2014). Melting of this snow combined with meltwater generated from ice trapped in the substrate could have carried sediment downslope in a similar mechanism as operates on Earth (Dickson et al., 2017, 2015b; Head et al., 2007; Marchant and Head, 2007). The small-scale melt-assisted flow, along with dry gravity-assisted debris flow and CO2-mobilized flow (Pilorget and Forget, 2016), could act to channelize erosion over time and create gullies with well-developed alcoves, channels, and sediment fans as are currently seen in the crater. There is no evidence within the crater of ponds or streams distal to the sediment fans – so while meltwater may have been the driving factor behind gully formation, there was insufficient water available to pond inside the crater or at the base of the fans.

The stratigraphic relationships of the gullies to other proposed paraglacial features show that the gullies in this crater fit within the paraglacial framework: gullies are stratigraphically younger than washboard terrain, but appear to be active during the latter period of washboard terrain formation, and they also span the period of polygon formation. The sharp, fresh morphology and lack of other superposing units of the stratigraphically youngest gully
systems (Figs. 11 and 14(B)) indicate that gully modification could have occurred quite recently, and gullies are the stratigraphically youngest paraglacial features within the crater. Therefore, the period of gully formation occurred coevally with the formation of other paraglacial features. In addition, the interpreted formation mechanism of gullies, one in which waning snow accumulation provided a source of fluid to mobilize sediment downslope after the major period of ice accumulation has halted, also supports its paraglacial origin.

4.5. Pitted terrain

4.5.1. Observations

A discontinuous unit of irregular pitted terrain is present on the crater rim (Figs. 6 and 15(B)). This unit is composed of patches of small pits ~10 m in diameter. In certain locations, pits merge to create a more irregular hummocky texture. This pitted terrain is located in a smooth deposit surrounding the crater, and the pits are forming in this smooth terrain. Several pits have formed near the crater rim crest on a gully alcove (Fig. 15(B)), suggesting the pitted terrain is stratigraphically younger than the most recent period of alcove incision.

4.5.2. Interpretations

The morphology, stratigraphic youth, location, and appearance of the pitted terrain is typical of dissected latitude-dependent mantle (LDM) (Head et al., 2003; Kreslavsky and Head, 2002; Milliken et al., 2003; Mustard et al., 2001). LDM is widely distributed across the mid- to high latitudes of Mars, covering at least 23% of the surface (Kreslavsky and Head, 2002), and is characterized by its smooth appearance that forms a mantle of uniform thickness and drapes across topography in locations where it is not disrupted. When it is disrupted or dissected in places, it forms pitted or knobby surface textures, particularly in the 30°–50° latitude region (Mustard et al., 2001). This deposit has been found to be meters thick, composed of an ice-rich dust mantle interpreted to have formed as an airfall deposit (Conway and Balme, 2014; Mustard et al., 2001). The deposition of this LDM is believed to have occurred during higher obliquity excursions between ~2100 and 400 kyr ago when Mars’ obliquity exceeded 30°, depositing snow and dust, and has subsequently been desiccated and undergone ice loss over the last ~300 kyr due to the current low obliquity ~25° (Head et al., 2003). There is also evidence of multiple layers within this deposit, indicating that several generations of the LDM have been deposited and subsequently removed due to orbital variations (Head et al., 2003; Kreslavsky and Head, 2002; Milliken et al., 2003; Schon et al., 2012). The presence of this unit is further evidence of the cyclical history of ice deposition and removal in the recent geologic history of Mars. The role of the LDM to the larger paraglacial period will be addressed in the following section.

5. Discussion

Taken together, the suite of geomorphic units discussed above describes a large-scale environmental adjustment to deglaciation. It is the association of these units that describe the martian paraglacial period; as on Earth, individual paraglacial features (i.e. gullies, polygons) can be found in non-glacial settings. Therefore it is the specific association of multiple features such as those documented above, which formed in response to deglaciation, that describes the paraglacial period.

5.1. The paraglacial transition

The stratigraphic relationships between the paraglacial features described above allow us to create a relative chronology of the activity within the crater (Fig. 16). Spatulate depressions are the stratigraphically oldest unit relative to the CCF. The washboard terrain is the next oldest feature identified in the crater, in general stratigraphically older than the gullies, and small-scale deposits within the spatulate depressions; however, evidence exists that the washboard terrain formed during older periods of gully formation and therefore acted over an extended period of time. Polygons and broad pits also appear to be stratigraphically older than gullies, as a broad pit is seen to be cross-cut by a gully channel, but an older gully system also contains polygons, suggesting that these processes overlapped and occurred in the same general period. Despite the stratigraphic overlap between other units suggesting a fairly long period of activity, the gullies showed the most recent modification within the crater, and are believed to be the stratigraphically youngest unit of the various features identified in the crater.

This stratigraphic analysis provides the framework around which the paraglacial period can be conceptualized, allowing us to
better understand the specific steps by which the crater became deglaciated and underwent paraglacial modification (Fig. 17). Using the terrestrial paraglacial period as an analogue, the martian paraglacial period would have initiated at the point where ablation outpaced accumulation, and deglaciation initiated. Ablation is believed to have occurred through sublimation (rather than melting), as there is extremely limited evidence of the presence of liquid water within the crater. Once sufficient ice was removed such that the crater rims became exposed, sediment mass-wasting off slopes would provide a debris cover for ice that slowed sublimation. Following the terrestrial analog described by Mackay et al. (2014) sublimation would have removed glacial ice from the base of the crater wall and crater floor that was not armored by a protective debris cover, leading to broad, flat-floored paraglacial depressions (Figs. 9 and 17(B)). The continued vapor-diffusive ice loss in the crater would have led to the broad-scale lowering of the crater floor and thickening of sublimation till on the crater floor and on portions of the crater wall. The lowering of the CCF surface and formation of paraglacial depressions would have created steeper slopes that triggered local ice flow; extensional stresses within the flowing ice caused transverse crevasses to form, creating washboard terrain (Fig. 10). Enhanced mass wasting in certain locations on the crater rim would create alcoves that allowed for channelized sediment transport (Fig. 14). Trapped snow and subsurface ice in these alcoves that underwent localized albedo-induced melting would lead to enhanced erosion and the creation of the first generations of gullies (Fig. 17(B)), which in turn would further erode and enlarge the alcove; in later periods of ice accumulation, these nascent crater rim alcoves would provide localized cold traps that preferentially accumulated ice and would eventually represent key ice accumulation areas within the crater (Head et al., 2008). Over time, periodic thermal cycling of the ice-cored wall sediment would form polygonal terrain and broad pits (Fig. 13) as washboard terrain eventually ceased to form, and gullies continued to evolve (Fig. 17(C)). As the sediment transport rates decreased, the rate of feature development would also have decreased.

In the future, if mean obliquity increases to \( \sim 35\degree \), ice will once again accumulate and flow onto the crater floor (Fig. 17(D)) (Fastook and Head, 2014). Gully alcoves around the crater rim will again act as initial accumulation areas for snow and ice, and even-

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**Table: Martian Paraglacial Duration**

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<thead>
<tr>
<th>Landsystem</th>
<th>Martian Paraglacial Duration</th>
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<tr>
<td>Spatulate Depressions</td>
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<td>Gullies</td>
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<td>Polygonal Terrain and Broad Pits</td>
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**Fig. 16.** Chronology of paraglacial processes on Mars. The relative ages and durations were created based on stratigraphic relationships observed between the features in the crater. Dashed lines indicate uncertainty in duration or in evidence of small-scale activity in a given landform.

**Fig. 17.** The martian paraglacial period. (A) Ice accumulates and flows into the crater. (B) When the climate changes, exposed ice begins to sublimate and paraglacial modification initiates. The removal of ice near the crater floor forms a paraglacial depression. Steeper slopes on the crater wall causes ice to begin flowing downslope, and transverse crevasses in the ice form washboard terrain. As the crater rim becomes further exposed through the loss of ice, more material will move downslope. As an alcove forms in the crater rim, material will become channelized as it flows downslope, eventually forming a gully. (C) Paraglacial modification continues; periodic thermal cycling creates polygonal terrain and broad pits, and gullies continue to evolve. (D) In the future, ice will again accumulate and flow into the crater, mantling the previously formed paraglacial features.
obliquity variations (such as in the past ~400 kyr, Fig. 18) the LDM was desiccated, forming the pitted terrain. The LDM desiccation can be thought of as an independent, localized paraglacial period separate from the larger paraglacial period addressed in this work, as the pitted terrain is deposited and removed on distinct timescales compared to the other features discussed in the crater.

On Earth, the paraglacial period is a transient phase that is relatively short-lived, although different processes operate on distinct time frames (Ballantyne, 2002a, 2002b) (Fig. 4). Terrestrial paraglacial morphologies can be rapidly eroded by subsequent pluvial and fluvial processes. In addition, melting and removal of ice is relatively easy in most terrestrial environments. For this reason, terrestrial paraglacial settings typically do not maintain large quantities of relict, subsurface ice, with the exception of some ice-cored moraines (Johnson, 1971) and stagnant, buried glacier ice (Marchant et al., 2002; Sugden et al., 1995). On Mars, however, the pressure and temperature conditions in the Late Amazonian are not conducive to ice melting and loss over short time frames, and there is only limited, transient evidence for the presence of liquid water. As such, large quantities of subsurface ice remain sequestered between net accumulation periods, a characteristic that is distinct from most terrestrial paraglacial settings. Several paraglacial features addressed in this work, including polygonal, craters, and Martian terrain, are interpreted to indicate the presence and distribution of remnant, buried ice present within the crater. Their widespread distribution throughout the northern and eastern portion of the crater suggest that this crater contains abundant near-surface ice on the crater walls, in addition to the ice that is expected to be present within the CCF.

Unlike most terrestrial paraglacial settings, martian paraglacial features are preserved and appear to have persisted for hundreds of thousands to millions of years, suggesting that the martian paraglacial period is longer in duration than in most terrestrial settings. This longer duration is due largely to the extremely cold and dry martian climate, the lower erosion rates and potentially larger sediment supplies exposed by ice loss than on Earth, as well as subsurface ice reservoirs present in crater interiors. Additionally, the lack of both vegetation and intense pluvial processes (e.g. rainfall) could act to preserve otherwise transient paraglacial landforms on Mars.

Without dedicated monitoring of modification rates within a paraglacial landscape, it is difficult to determine if a location is still undergoing paraglacial modification, and therefore whether it is still in the paraglacial period. However, the stratigraphic youth of certain paraglacial features within the crater analyzed in this work suggest that within this glaciated crater, the paraglacial period may still be ongoing and in its waning stages. Craters containing CCF in other locations that have experienced paraglacial activity under different conditions may deviate from this trend; a larger analysis of the global inventory of paraglacial reworking on Mars is therefore necessary and underway (Jawin and Head, 2016).

5.3. Outstanding problems

Several outstanding questions remain with regard to deglaciation, the martian paraglacial period, and the associated paraglacial features. To what extent were the crater floors and walls filled with glacial ice in peak-glacial periods? How would the paraglacial features be modified if a new period of ice accumulation commenced? What is the variation in distribution of the paraglacial features across the martian mid-latitudes, as well as the rest of the planet? Are the paraglacial features always spatially and temporally associated as is seen in the crater analyzed in this study (Fig. 6)? Assuming that a paraglacial period postdates every high-obliquity period, is it possible to identify relict paraglacial features from previous climate excursions? Is the paraglacial period evident in other glacial features, such as lobate debris aprons and lineated valley fill (LDA and LVF, respectively)? Does evidence for a paraglacial period exist in glacial features outside the mid-latitudes, such as those at the poles or near the equator? Further analyses of glacial features present in the martian mid-latitudes will help to answer many of these questions.

6. Conclusions

The terrestrial paraglacial period is characterized primarily by elevated sediment transport rates immediately postdating glacial retreat. This period describes the broad environmental response to deglaciation, and ends when modification rates return to non-glacial conditions. This paraglacial period fits well as a conceptual model when applied to martian mid-latitude glaciated craters, and explains the local environmental response to global obliquity-driven climate variations. The climate transition is investigated in a southern hemisphere mid-latitude glaciated crater by observing the presence and association of spatulate depressions, washboard terrain, gullies, polygonal terrain, and broad pits, all of which are consistent with formation as a response to ice loss within the crater. The formation of spatulate depressions, washboard terrain, polygons, and gullies appear to be temporally linked by their spatial proximity and interrelated stratigraphic relationship. Cycles of deposition and subsequent desiccation of the LDM also occurred throughout the recent past, forming pitted terrain.

Despite the rapid, transient nature of the paraglacial period on most places on Earth, the martian paraglacial period may have lasted ~5 million years due to the lack of broad-scale fluvial activity and the general long-term topographic stability on Mars in the Late Amazonian.

Further analyses are required to better understand the timing, duration, and variability of the martian paraglacial period. A global analysis of paraglacial features is warranted, as well as quantitative estimates of the amounts of sediment transported through the various paraglacial processes outlined here. These data will allow a deeper understanding of the climate variability of the Late Amazonian as well as erosion and modification rates throughout this period.

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Supplementary materials

Supplementary material associated with this article can be found, in the online version, at doi:10.1016/j.icarus.2018.01.026.

References

