Amazonian mid- to high-latitude glaciation on Mars: Supply-limited ice sources, ice accumulation patterns, and concentric crater fill glacial flow and ice sequestration

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ABSTRACT

Concentric crater fill (CCF) occurs in the interior of impact craters in mid- to high latitudes on Mars and is interpreted to have formed by glacial ice flow and debris covering. We use the characteristics and orientation of deposits comprising CCF, the thickness of pedestal deposits in mid- to high-latitude pedestal craters (Pd), the volumes of the current polar caps, and information about regional slopes and ice rheology to address questions about (1) the maximum thickness of regional ice deposits during the Late Amazonian, (2) the likelihood that these deposits flowed regionally, (3) the geological regions and features most likely to induce ice-flow, and (4) the locations and environments in which ice is likely to have been sequestered up to the present. We find that regional ice flow under Late Amazonian climate conditions requires ice thicknesses exceeding many hundreds of meters for slopes typical of the vast majority of the surface of Mars, a thickness for the mid-latitudes that is well in excess of the total volume available from polar ice reservoirs. This indicates that although conditions for mid- to high-latitude glaciation may have persisted for tens to hundreds of millions of years, the process is "supply limited," with a steady state reached when the polar ice cap water ice supply becomes exhausted.

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, and topographic slope data show that Amazonian impact craters have been clearly modified, undergoing crater interior slope reduction and floor shallowing. We show 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 spin–axis obliquity. We demonstrate that flow from a single episode of an inter-crater terrain layer comparable to Pedestal Crater deposit thicknesses (~50 m) cannot fill the craters in a time period compatible with the interpreted formation times of the Pedestal Crater mantled ice layers. We use a representative obliquity solution to drive an ice flow model and show 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. The cyclical pattern of waxing and waning mantling layers results 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 concentric crater fill. In this scenario, the formation of mantling sublimation till layers seals the accumulating ice and sequesters it from significant temperature variations at diurnal, annual and spin–axis/orbital cycle time scales, to produce ancient ice records preserved today below CCF crater floors.

Lack of meltwater features associated with concentric crater fill provides evidence that the Late Amazonian climate did not exceed the melting temperature in the mid- to high-latitudes for any significant period of time. Continued sequestration of ice with time in CCF and related deposits (lobate debris aprons and lineated valley fill) further reduces the already supply-limited polar ice sources, suggesting that there has been a declining reservoir of available ice with each ensuing glacial period. Together, these deposits represent a candidate library of climate chemistry and global change dating from the Late Amazonian, and a non-polar water resource for future exploration.

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1. Introduction

A wide range of evidence shows that the current distribution of ice on the surface of Mars (concentrated primarily at the polar caps) is anomalous, and that the Amazonian era was characterized by a variety of non-polar ice-related deposits (Head and Marchant, 2008; Carr and Head, 2010) ranging from the pole to the equator in distribution. These include (Fig. 1):

1. High-to mid-latitude mantles (Kreslavsky and Head, 1999, 2000; Mustard et al., 2001; Milliken et al., 2003; Head et al., 2003) and polygonal patterned ground (e.g., Mangold, 2005; Levy et al., 2010b).

2. Mid-latitude Lobate Debris Aprons (LDA) and Lineated Valley Fill (LVF) (Lucchitta, 1981, 1984; Mangold, 2003; Pierce and Crown, 2003; Li et al., 2005; Head et al., 2005, 2006a, 2006b, 2010; Levy et al., 2007; Ostrach et al., 2008; Morgan et al., 2009; Baker et al., 2010; Dickson et al., 2008, 2010), concentric crater fill (CCF) (Kreslavsky and Head, 2006; Levy et al., 2009, 2010a; Dickson et al., 2012; Beach and Head, 2012, 2013), phantom LDA (Hauber et al., 2008) and Pedestal Craters (Pd) (Kadish and Head, 2011a, 2011b; Kadish et al., 2009, 2010a).

3. Low-latitude Tropical Mountain Glaciers (TMG) (Head and Marchant, 2003; Head et al., 2005; Shean et al., 2005, 2007a; Kadish et al., 2008).

General circulation models (GCM) (e.g., Haberle et al., 2003; Forget et al., 2006; Madeleine et al., 2009) and glacial flow models (e.g., Fastook et al., 2008, 2011) illustrate the orbital parameter and atmospheric/surface conditions under which periods of non-polar glaciation are favored, and the resulting patterns of accumulation of snow and the flow of ice (Milliken et al., 2003; Forget et al., 2006; Madeleine et al., 2009; Fastook et al., 2008, 2011). Geological observations and impact crater size-frequency distribution data strongly suggest that during the Middle and Late Amazonian (Hartmann and Neukum, 2001), a significant part of the mid-to-high latitudes in both hemispheres was covered by regional snow and ice deposits, preserved today beneath Pedestal Craters (Kadish et al., 2009, 2010a; Kadish and Head, 2011b) and that local depressions (primarily impact craters) were the sites of significant ice accumulation, and preservation beneath a residual debris cover (concentric crater fill) (Kreslavsky and Head, 2006; Levy et al., 2009, 2010a; Beach and Head, 2012, 2013). Geological interpretations of buried ice in the lobate debris aprons and lineated valley fill (e.g., Head et al., 2006a, 2006b) have been confirmed by SHARAD radar data (Holt et al., 2008; Plaut et al., 2009). Indeed, the preservation of this buried ice strongly suggests that Amazonian climate temperatures in the mid-latitudes have not exceeded ice-melting temperatures for any significant period of time. Pedestal Crater heights show that a significant amount of snow and ice accumulated over broad regions in the mid-to-high latitudes during these periods. Regionally the mean height is ~50 m, but values up to 160 m are seen in Utopia Planitia (Kadish et al., 2010a). Accumulations in CCF are typically many hundreds of meters and can exceed several kilometers (Kreslavsky and Head, 2006; Levy et al., 2009, 2010a; Beach and Head, 2012, 2013), often substantially filling the crater (Dickson et al., 2008, 2010, 2012). We envision the concentric crater fill to be mostly ice, with a relatively thin debris cover from the crater walls. There are four pieces of evidence for this: (1) the distribution of “ring-mold craters” (Kress and Head, 2008) suggest that the thickness of the current debris cover in CCF is about 15–20 m, much thinner than the total thickness of the often several km thick CCF; (2) Where SHARAD data have resolved the Lobate Debris Aprons (LDA) (Holt et al., 2008; Plaut et al., 2009), the hundreds of meters of ice below the debris cover is relatively debris-free, and the debris cover is in the less than 15 m range. (3) Geological relationships in crater interiors where floors and lower walls are covered with concentric crater fill-like textured material (Head et al., 2008) strongly suggest that the ice is being formed on the crater walls, covered with debris as the upper rocky walls are exposed, and transported out onto the floor as debris-covered glaciers, rather than ice-cemented talus/soils (Berman et al., 2005), and (4) Models of LDA emplacement (Fastook et al., 2013) suggest that a debris thickness of this order is very realistic and plausible. Different phases or periods of glaciation separated in time may produce different thickness of overriding debris-covered ice, but each layer is interpreted to be largely ice rich and debris poor. Unfortunately, the small size of these craters and the scattering of the radar signal from the walls make it difficult to obtain measurements of the floor fill material to detect possible layering. We have undertaken a targeted search, but the data are now being collected and reduced and we plan to test this hypothesis in upcoming analyses. However, the preservation of these various deposits and generally
pristine landforms provides a basis to address important questions about Amazonian climate history.

Historically, diverse processes have been proposed to account for these types of landforms. For example, candidate hypotheses proposed to account for the CCF have included eolian processes (Zimbelman et al., 1989), debris from bedrock recession of scarps (Sharp, 1973), ice-assisted talus flow (Squyres, 1978), rock glaciers (Squyres, 1979; Squyres and Carr, 1986; Haeberli, 2006), internally deforming ice (Lucchitta, 1984), flow of an icy substrate down into a newly formed crater (Senft and Stewart, 2008), and debris-covered glaciers (Kreslavsky and Head, 2006; Garvin et al., 2006; Levy et al., 2009, 2010a; Beach and Head, 2012, 2013). Kreslavsky and Head (2006) and Levy et al. (2010a) discuss in detail these different hypotheses in the context of recent high-resolution image and altimetry data, and arrive at the conclusion that CCF is composed largely of debris-covered ice that at some point flowed and then sublimated, to produce the observed flow-like debris cover (Fig. 2). On the basis of a variety of evidence, numerous workers have concluded that the age of the CCF is likely to be Amazonian, reflecting processes that have taken place within the last several hundred million years (Kreslavsky and Head, 2006; Levy et al., 2010a; Beach and Head, 2012, 2013).

Several concepts have been proposed to account for the sources of ice interpreted to fill the crater interiors (Fig. 3). For example, Levy et al. (2010a) suggest that accumulation in crater wall alcoves could be feeding the ice forming the CCF (Fig. 3a). They also point out, however, that the surface of the CCF, which appears to have been lowered by as much as 300 m, could have overfilled some of the craters. They thus raise the question of sources of ice from outside the crater (Fig. 3b). For example, they discuss the possibility, strongly suggested by the presence of regional ice deposits revealed by pedestal craters, that the CCF material could have been part of, and derived from, a larger inter-crater ice complex that covered the entire landscape. In this context, rather than local crater accumulation and crater interior fill (Fig. 3a), the CCF might be the remnant of a larger regional ice sheet complex (Fig. 3b). In this scenario, the regional ice sheet reaches a thickness sufficient to flow into the crater and create the concentric crater fill deposits; following return to non-glacial conditions at lower obliquity, the regional ice deposits sublimate and return to the

![Fig. 2. Morphologic and topographic characteristics of an example of classic concentric crater fill, located at 351.87°, 44.15° (from Levy et al., 2010a). (a) Crater with classic, relatively concentric crater fill, showing nested concentric ridges and a moat between the crater wall and the edge of the CCF deposit. The white line indicates the location of the topographic profile A–A′–A″, shown as an HRSC DTM profile in (b). CTX image P14_006570_2241. (b) Current HRSC DTM topographic profile of crater shown in (a), compared to a fresh, unfilled crater profile (inset; Garvin et al. 2002). Wavy line in the inset is the current CCF profile, suggesting that the thickness of CCF material may be as much as 350 m.](image-url)
2. Analysis of key questions

2.1. What ice thickness is required to initiate ice flow on a flat, intercrater area under Late Amazonian environmental conditions?

We approach this question by assessing the thickness, slope and the temperature–dependent flow behavior of ice under Amazonian climate conditions. We assume here that any ice deposition from obliquity-controlled climate variation is relatively uniform over a broad area, with no localized concentrations of accumulation such as might occur in wind- or sun-sheltered alcoves. We dealt with a similar situation in Fastook et al. (2011) in modeling the flow in a valley system, and the resulting flow features were not consistent with alcove-originated flow, but were
more likely the remnants of a collapsing regional ice sheet that covered, but left no imprint, on the plateaus between the valleys. In this analysis we also examined the regional variations in accumulation rates on the surrounding broad plateau predicted by the GCM and found that such lateral variations were regional in nature and did not predict intensive local snow and ice buildup sufficient to induce ice flow due to local thickness perturbations.

Ice flow is obtained by integrating strain rates through the vertical ice column, with strain rates related to stresses through the Flow Law with a temperature-dependent rate factor, A (Paterson, 1994).

\[ \dot{e} = A \tau^3 \]  

(1)

With the shallow-ice approximation, we can neglect all but the vertical gradient of the horizontal velocity that results from a driving stress obtained from conservation of momentum. Assuming that stress in the vertical varies linearly from zero at the ice surface \( (z=H) \) to the maximum driving stress at the bed \( (z=0) \) we have the following equation:

\[ \frac{1}{2} \frac{du_x}{dz} = A[\rho g(H-z)\nabla h]^4 \]  

(2)

We distinguish between \( H \), the ice thickness, and \( h \), the ice surface elevation, to accommodate variations in the bed elevation. We can integrate the strain rates from the bed where the velocity is zero to some height \( z \) within the ice column, thus obtaining the dependence of the horizontal velocity on depth within the ice column.

\[ u_x(z) = \frac{1}{2} A[\rho g|\nabla h|^4][H^4 - (h-z)^4] \]  

(3)

Integrating this one more time from the bed to the surface and dividing by the thickness yields the column-averaged ice velocity necessary for inclusion in the Continuity equation for ice flow, as follows.

\[ \bar{U} = \frac{2}{5} A[\rho g|\nabla h|^4]H^4 \]  

(4)

The most important factors are that the flow velocity depends on surface slope to the third power and the ice thickness to the fourth power, with a strong Arrhenius temperature-dependence through the flow constant \( A \). Dust content and grain size also influence the flow constant allowing for ice that at high stresses is 2–3 times softer than pure ice at the same temperature (Goldsby et al., 2013). In Goldsby et al. (2013), two separate laboratories reported results on the effects of varying amounts of dust. Different results from the two labs are currently being investigated in order to increase convergence and more fully understand the relationships between dust content, grain size, temperature, and rheology. As these important high-stress, high strain-rate experimental data become better understood, they can be applied more confidently to the low-stress regimes of ice sheets. In the meantime, given the uncertainty in dust content and grain size of the ancient ice on Mars, and the current level of uncertainty in the experimental ice deformation data, we utilize the pure ice case (Paterson, 1994), which in the worst case will yield a flow velocity that is low by a factor of no more than 2 or 3 when compared with the new results reported in Goldsby et al. (2013).

Flow velocity (mm/yr) as a function of ice thickness and surface slope for mean-annual temperatures of 215, 225, and 235 K (Haberle et al., 2003) are shown in Fig. 4. It is worth noting that annual and diurnal temperature variations have little impact on the flow of ice, since even the annual amplitude damps out at a depth of approximately 10 m and is insignificant below this depth where the bulk of the flow deformation takes place.

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**Fig. 4.** Flow velocity (mm/yr) as a function of ice thickness (m) and surface slope (%) for a temperature of 215 K (left), 225 K (middle), and 235 K (right) from Eq. [4] with an Arrheniushly-activated flow constant, \( A \), illustrating the strong dependence of velocity on surface slope, ice thickness, and temperature. The horizontal axis is ice thickness, and the vertical axis is surface slope, so lowest velocities are in the lower left, corresponding to cold, thin ice, and highest velocities are on the right for warm, thick ice. The heavy solid line in each figure indicates a velocity of 1 mm/yr.
What is the role of surface slope on Mars? The distribution of surface slopes provides a context in which to interpret Fig. 4 and the regions where ice might be expected to flow under these conditions. Kreslavsky and Head (1999, 2000) analyzed global surface roughness and slopes on Mars and their baseline dependence, showing the cumulative slope-frequency distribution for
Fig. 5. (continued)
baselines at 0.4, 3.2 and 25 km (Fig. 5a). Clearly, the vast majority of slopes at these baseline lengths are less than 1–2°. At the other end of the spectrum, the steepest point-to-point slopes on Mars (Fig. 5b) occur in association with three major types of features; (1) polar cap scarps can range up to 35–55°, testimony to the resistance to flow of ice under current Mars polar conditions; (2) tectonic features, typically linear scarps and related erosional features, such as the walls of Valles Marineris and the dichotomy boundary, can exceed 35°; and (3) impact craters also contain slopes in excess of 35°. The global distribution of these slopes (Fig. 5c and d) illustrates these associations. The vast majority of the surface is characterized by slopes of less than 1°–2°, even in the much more heavily cratered southern uplands (Fig. 5c), and the steepest slopes are concentrated in the equatorial regions (Fig. 5d) in contrast to the mid- to high-latitudes, where regional ice-related deposits dominate (Fig. 1). Detailed analysis of slopes in three specific areas (Fig. 5e–g) shows that steeper slopes in excess of a few degrees are very highly concentrated in crater interiors, specifically on the upper parts of interior crater walls.

Returning to ice-flow velocities (Fig. 4), at 215 K, a mean-annual temperature typical of the current climate on Mars at mid-latitudes (Haberle et al., 2003), low surface slopes (~1°) typical of the flat inter-crater terrain (Kreslavsky and Head, 1999, 2000) (Fig. 5) require a considerable thickness of ice (800–1000 m) to initiate flow. Only for considerably greater surface slopes (> 5°) is there any significant flow for ice thicknesses of 200 m, and such slopes are not common on typical geological feature baselines (~1 km; Fig. 5a). Even at slopes as high as 20° (typical of interior crater walls, Fig. 5b–f; Kreslavsky and Head, 2006) a 50 m ice thickness only yields ice flow rates of ~0.2 mm/yr (Fig. 4). Warming the mean annual temperature to 225 K softens the ice and increases all velocities by a factor of 4, reducing the threshold for 200 m to a slope of 3.5°. A further increase to 235 K reduces the necessary slope to 2.2°; nonetheless, a 100 m thick accumulation of ice still requires a slope greater than 5° (Fig. 4) to generate flow, and even under these circumstances, the flow velocity is only 1 mm/yr, considerably slower than any known glacial ice flow on Earth.

These very low flow velocities are clear testimony to the significant temperature dependence of flow rates; in contrast, typical terrestrial alpine glacier flow velocities are 90–150 m/yr with larger valley glaciers up to 180 m/yr. Some terrestrial ice surges approach and exceed 6–12 km/yr. For wet-based glaciers, ice flow rates of a few centimeters per day are common, and velocities of 3 m/day are the exception. The flow rates in extremely cold terrestrial polar environments are considerably slower; ice flow velocities over the Antarctic ice sheet (Fig. 6) are typically less than 10 m/yr (Rignot et al., 2011), with the Vostok station flow velocities less than 2 m/yr (Richter et al., 2013). Terrestrial Antarctic cold-based debris-covered valley glaciers are even more dramatically slower, and begin to approach martian flow rates; for example, the Antarctic Dry Valley Mullins debris-covered glacier flow rate is ~40 mm/yr (Rignot et al., 2002).

For the environment of the Amazonian period of Mars, the non-linear dependence of the flow velocity on surface slope and ice thickness is clearly evident (Fig. 4), with lowest velocities in the lower left for small slopes and thicknesses, and the highest velocities in the upper right for higher topographic slopes and higher ice thicknesses. Also clearly evident is the dependence of velocity on temperature, with higher velocities in the warmest right-hand figure. Given the accumulation rates and mean-annual temperatures (215 K) anticipated during the Late Amazonian at these latitudes (Haberle et al., 2003), an ice sheet would require a thickness of 750 m to initiate flow.
(Fig. 4) on these low regional slopes (Fig. 5). Even with softer ice at warmer temperatures, a thickness of 500 m is required at 225 K, and 370 m is required at 235 K (Fig. 4).

Returning to the original question as to what thickness is required to initiate ice flow on a flat, inter-craterr area (Fig. 5) under Late Amazonian environmental conditions (Fig. 4), we conclude that for the vast majority of regional slopes on Mars, no significant flow would occur unless ice thicknesses exceeded many hundreds of meters (~750 m).

2.2. Could the current Pd mean thickness value (~50 m) be the remnant thickness of equilibrium flow?

Could the regional ice layer have reached the requisite ~750 m ice thickness to initiate flow in the mid- to high-latitude plains, intercrater areas and within craters (equilibrium flow), and then undergone sublimation to yield the remnant deposits seen today? One estimate of the average thickness of candidate Late Amazonian ice sheets is the height of scarps at the margins of pedestal craters (Pd). Analysis of the distribution of Pd shows that the mean value of Pd scarp height is ~50 m, interpreted to represent the regional thickness of ice deposits that characterized the mid- to high-latitudes during Pd formation (Kadish et al., 2009, 2010a; Kadish and Head, 2011b). During interglacial periods, the ice between pedestal craters sublimated, leaving the height of the scarp as a proxy for ice thickness. Could this average ~50 m thickness of the Pd layer instead be the last phase of a much thicker persistent ice sheet that reached a configuration that supported flow (Fig. 3b)? Alternatively, could the average ~50 m thickness of the Pd layer represent a transient, relatively thin ice layer that covered the surface and locally deformed and flowed into crater depressions only where slopes are large enough (Fig. 3a)?

Evidence exists for the latter case (thin ice layer) in the form of Pedestal Craters, Perched Craters, and Excess Ejecta Craters described in detail in Kadish et al. (2009, 2010a), Kadish and Head (2011b). These three types of craters have been interpreted to be related to impacts into an ice layer that is typically less than ~60 m thick (Fig. 7). Each crater type reflects differing degrees of impact penetration into the substrate, followed later by complete sublimation of any un-armored regions of the ice deposit that were exterior to the crater deposits. Evidence for the repeated deposition and removal of such relatively thin ice layers is suggested by the ~80 m height difference observed between two pedestal craters (Pds) ~20 km apart. In addition, 30 Pd craters have superimposed Pds. For this to occur the first Pd would have to impact into an ice layer and that entire ice layer outside the armored zone would have to be removed. A second ice layer would have to reform, and the superimposed Pd would impact into this new ice layer, arming it as the first Pd armored the earlier layer. Subsequently all of this new ice layer outside its newly armored zone would be removed, leaving the second Pd superimposed upon the first Pd. Clearly this requires that there be multiple episodes of ice-rich layer cover. An additional observation that suggests that these ice layers did not experience significant flow is the lack of any observed pedestal crater elongation or deformation (Kadish and Head, 2009, 2011a, 2011b; Kadish et al., 2010a).

Additional evidence supporting the presence of regional ice layers comes from general circulation model (GCM) results (Mischna et al., 2003; Levrad et al., 2004; Mischna and Richardson, 2005; Forget et al., 2006; Levrad et al., 2007; Madeleine et al., 2009). These GCMs predict ice accumulation rates as high as 10 mm/yr during periods of high obliquity, precisely in the mid-high latitudes where these craters occur. An important question is why the regional transient layer thickness apparently never greatly exceeds the ~50 m regional suggested by the current Pd thicknesses (Fig. 7), since GCM results suggest that a layer of this thickness could form in as little as 20 kyr, a time period much shorter than the duration of single obliquity excursions (~125 kyr).

Estimates of the total volume of the global Pd-defined layer during its regional latitude-dependent presence have been made (Kadish et al., 2009, 2010a; Kadish and Head, 2011b; Fastook and Head, 2013). These estimates are close to the known current volumes of the polar caps (Fastook and Head, 2013). Polar ice deposits are thought to be the main source of moisture for the higher-obliquity mid-high latitude snow precipitation. Estimates for the combined current north and south polar cap volumes range from a minimum of ~3 million km³ to a maximum of ~5 million km³ (see discussion in Fastook and Head, 2013). If an average of these two values for the entire volume of the current polar caps (~4 million km³) is spread evenly across the area known to be characterized by pedestal craters and other mid- to high-latitude ice related deposits (from 30° to the poles; Fig. 1), the mean ice deposits thickness would be ~55 m (Fig. 8). If the latitude-dependence is more limited, the mean thickness increases, but is still well below thicknesses that would begin to induce glacial flow on the slopes characteristic of the plains and intercrater areas (Fig. 5). For example, if the total current polar ice supply is redistributed from 40° to the pole, the mean thickness is ~77 m, and from 50° to the pole, ~118 m (Fig. 8). We therefore suggest that the transient ice layer is “supply-limited” in that when these caps are exhausted, the source is removed, and deposition ceases even if the obliquity is still high.

In summary, given the accumulation rates and mean-annual temperatures (215 K) anticipated during the Late Amazonian at these latitudes, an ice sheet would require a thickness of ~750 m
to initiate flow on these low regional slopes. Even with softer ice at warmer temperatures, a thickness of 500 m is required at 225 K, and 375 m is required at 235 K. Evidence from a variety of Pedestal and Excess Ejecta Craters suggests that the regional ice thicknesses were never this high (Fig. 7), and thus that regional glacial flow was unlikely to have taken place. In addition, superimposed Pedestal Craters require multiple episodes of layer emplacement, presumably during periods of high obliquity, with subsequent complete removal when lower obliquity no longer favored positive mass balance. Constraints on the total volume of ice currently available in the polar cap reservoirs, coupled with the broad areal distribution of Pedestal Craters, suggest that regional mid-latitude layers were supply-limited, with thicknesses of the order of 50 m (Fig. 7), a thickness insufficient to have supported regional flow (Fig. 4).

2.3. What slopes are required to initiate ice flow under Late Amazonian conditions and where do such slopes occur geologically?

We previously saw (Fig. 4) that only for surface slopes > 5° is there any significant flow predicted for ice thicknesses of 200 m, and such slopes are not common on typical geological feature baselines (Fig. 5). Furthermore, even at slopes as high as 20° (typical of interior crater walls, Fig. 5; Kreslavsky and Head, 2006) a 50 m ice thickness only yields ice flow rates of ~0.2 mm/yr (Fig. 4). Warming the mean annual temperature to 225 K increases all velocities by a factor of 4, reducing the threshold for 200 m to a slope of 3.5° and a further increase to 235 K reduces the necessary slope to 2.2°, but 100 m thick ice still requires a slope greater than 5° to generate flow, and even under these circumstances, the flow velocity is only 1 mm/yr. We therefore adopt a slope of > ~5° to locate the most likely places for ice flow under Late Amazonian climate conditions (Fig. 5).

Under these circumstances, where would we expect the most significant ice flow, even at these low flow rates? From an analysis of individual impact crater topography, Garvin et al. (2002) showed that crater-wall slopes correlate strongly with crater size, with slopes ranging from 10 to 30°. Analysis of regional and global slope maps (Kreslavsky and Head, 1999, 2000; Fig. 5c-g) shows that impact crater rims and interior walls are the most abundant landform with steep slopes. Furthermore, Kreslavsky and Head (1999) mapped the distribution of the steepest slopes, found a north–south topographic slope asymmetry (2003), and analyzed the processes modifying impact crater slopes and assessed the implications for Amazonian climate history (2006). They confirmed that impact craters, together with tectonic scarps, were the landforms that displayed the steepest slopes (Kreslavsky and Head, 1999; their Fig. 3). They also found that equatorial-latitude impact craters maintained steep interior wall slopes (Fig. 5d), but that impact craters at mid- to high-latitudes were typically characterized by much lower slopes (Kreslavsky and Head, 2006), and that at mid-latitudes (40°–50° latitude) there was a distinctive north–south topographic asymmetry interpreted to be related to degradation of pole-facing slopes to < ~20° (Kreslavsky and Head, 2003). In summary, on the basis of global altimetry data and regional topographic slopes, impact crater interior walls are the most likely places for the occurrence of slopes consistently in excess of the ~5° slope value most likely to cause flow under Late Amazonian climate conditions. The data plotted in Fig. 4 indicate that these slopes would easily induce glacial ice flow, even for layers significantly less than 200 m in thickness.

Given that the most abundant geological features with slopes sufficient to drive appreciable flow are crater interior walls and rims (Fig. 5), we examine two possible scenarios (Fig. 9): (1) that a single episode of a widespread ice-rich layer blanketed the landscape and flowed locally down the steep slopes of the crater walls into the centers of the craters, and (2) that repeated episodes of blanketing and subsequent sublimation occurred, with each layer emplacement contributing to the crater fill.

As we have described previously, evidence exists for a widespread transient ice-rich layer typically ~50 m in thickness in the mid- to high-latitude area where GCM simulations are known to deposit ice at high obliquity. How does this thin layer become the several-hundred-meters–to-kilometers-thick deposit observed as CCF?

In the single-episode case (Fig. 9a), we assume that the ~50 m thick regional ice sheet is evenly draped over the surface, and that the underlying crater topography is that of a fresh impact crater (Fig. 10a). This configuration is likely to be somewhat more complicated due to the effects of wind-blown snow and the local effects of crater topography on snow accumulation and retention (e.g., Madeleine et al., 2012), factors that we will return to in the next section. Important to the understanding of the subsequent behavior of this ~50 m thick ice layer are the basic topographic elements of the crater itself (Fig. 10a), consisting of the raised rim crest (typically < 5° slope), the rim crest, the crater interior walls (typically > 20° slope) and the undulating, relatively flat crater floor. The cross-sectional diagram (Fig. 10a) thus shows that slopes somewhat less than 5° exist on the outer rim, facing away from the crater, and that slopes much greater than 5° exist on the interior wall, facing into the crater interior. Ice deposited on the crater rim would tend to flow outward, away from the crater interior, except at the crater rim crest, where it would flow inward. In order for ice to flow into the crater from the outside, ice would have to very significantly exceed the ~50 m thickness interpreted from the pedestal craters (by factors of ten to twenty) (Fig. 4), also exceeding the apparently supply-limited polar source regions of the ice (Fig. 8). A critical observation concerning outward versus inward-flowing ice is that the crater rim crest is typically much greater than 50 m for the range of crater diameters observed in the CCF population (Fig. 10b) (Garvin et al., 2002, 2003). This means...
that unless there is an anomalously low depression in the rim crest continuity, any flow from a snow/ice mantling deposit will always be outward from the region of the upraised rim. Since there is unlikely to be any significant exposed bedrock to act as sources of debris for these icy rim deposits, a debris cover that might preserve the ice deposits subsequent to the glacial period, when the regional snow and ice deposit sublimated away, is not likely to form.

Clearly, the distribution of slopes (Fig. 10a) heavily favors the flow of ice from the rim crest and inner wall into the interior of the crater. For the single layer uniform blanket case (Fig. 9a), ice at the top of the crater wall thins as ice flows down the crater walls (Fig. 9a and b), leading to a local inward-sloping surface that might draw some ice inward from the immediate rim crest area for a very short period of time, but is unlikely to draw in significant amounts due to the relatively thin mantle and the outward-sloping rim topography. Flow of ice from the steep inner wall will be accompanied by mass wasting of rocks and soil from the exposed crater wall (Fig. 9b), beginning a process of debris cover (Fastook et al., 2013). Despite the availability of steep slopes to induce inward ice flow, and the very slow predicted flow rates, the supply-limited nature of this scenario (Figs. 9 and 10) means that the crater can only be partially filled (Fig. 9c) by a single layer of ~50 m thickness. Model results in the next section will help to quantify this time scale.

In the multiple-episode example (Fig. 9d), we consider the same type of transient layer that uniformly blankets the terrain and, as with the single-episode case, flows down the steep inner walls into the crater interior, thickening the deposit there (Fig. 9d and e). When the climate changes and the blanketing layer sublimates away, the thicker deposit in the crater interior is less likely to be completely removed (Fig. 9e). One reason is that the layer has been thickened by flow down the steep slopes of the crater walls. More importantly, however, the deposit in the crater interior can be armored by debris accumulating on its surface derived from the crater walls, significantly inhibiting sublimation.

**Fig. 9.** Two models for the duration of formation of interior concentric crater fill. In (a), a uniformly thick 50 m snow and ice layer is deposited over the crater landform during a single obliquity excursion. The presence of steep inner wall slopes means that flow initiates there (b) even with the relatively thin ice cover, and glacial flow extends down the crater walls toward the crater floor; ice on the rim is too cold to flow on the shallow slopes. At the end of this glacial phase, sublimation removes the ice layer, except where is has been thickened by flow and covered by wall-slope debris (c). Mass balance calculations show that the single 125 thousand year long 50 m thick depositional event alone is insufficient to form a significant deposit of CCF inside craters. Deposition of a new layer (d), however, will cause flow of the new ice on the steep upper slopes and rejuvenate downslope movement of the entire debris-covered glacier toward the crater center (e). Multiple glacial periods of progressive deposition of ice and its preferential movement downslope cause the debris-covered glacier to move toward the crater center, converge, and fill the crater floor (f).

**Fig. 10.** Guidelines for locations of potential ice flow in typical impact craters. (a) Topographic profile (black line) for the crater from Fig. 2. Exterior rim slopes (blue line) face outward and are less than ~5°, while interior wall slopes face inward and are in excess of 10°, up to ~20° (also see Fig. 5). For exterior ice to flow into the crater, it must be thicker than ~200 m to be able to flow over the rim raised into the interior. Ice flow is thus favored on the inner crater wall slopes. (b) Crater rim heights relative to the surrounding exterior terrain. The rim heights of impact craters more than 2–3 km in diameter are in excess of the ~50 m snow/ice mantling layer mean thickness measured from the heights of pedestal crater scarps (e.g., Kadish et al., 2009, 2010a). This suggests that only under very exceptional circumstances (e.g., a rim/rim crest breach) could a regional ice deposit flow into impact crater interiors. Data from Craddock et al. (1997) and Garvin et al. (2003). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
This mechanism of debris accumulation on the ice surface has been used to describe a formation mechanism for lobate debris aprons (Fastook et al., 2013) and is supported by terrestrial evidence from debris-covered glaciers in the Dry Valleys, Antarctica (Marchant and Head, 2007; Marchant et al., 2002, 2007; Kowalewski et al., 2011). On the basis of this scenario, mantling layer formation and removal happens many times (Fig. 9f), with the layer repeatedly forming, flowing, and then sublimating away in the crater exterior and the non-mantled part of the crater interior. In this obliquity-driven movement of water ice to and from the mid-high latitudes, a small amount of ice is deposited in the crater interior depressions in each cycle, having accumulated there by flow down the steep slopes of the crater walls, and then being preserved by increasing debris cover (Fig. 9f). Model results in the next section quantify the time scale of this process.

2.4. What was the nature of snow and ice cover and glaciation during periods of maximum ice accumulation in the Late Amazonian?

On the basis of our analysis thus far, what can we say about the transient inter-crater glacial ice cover and the distribution of glaciation in areas of steeper slopes, such as crater interiors? We have ruled out the presence of a many hundreds of meters thick persistent regional ice sheet based on the excessive volume of ice that would be required relative to its assumed source in the current polar caps. We have also seen that the observed pedestal crater thickness (~50 m) would not result in glacial flow in the intercrater areas, but could produce flow on the steep slopes associated with crater interior walls (Figs. 9 and 10). Flow should be limited to these crater rim crests and interior walls; the outward-sloping nature of the raised crater rim suggests that any ice deposits there would flow outward, if they flowed at all. Thus, the most likely region for preservation of glacial deposits during mid- to high-latitude glacial periods are crater interiors, due to their steep slopes (Fig. 5), and to the availability of debris on the crater walls to enhance preservation of the ice in interglacial periods. Using this background scenario, we ask the following questions: What would these deposits look like if they were produced by a single, transient episode of relatively thin ice-rich layer deposition that deformed as it flowed into the crater interior from the wall? Is this sufficient to produce the observed features, or were there multiple episodes of ice-layer deposition that formed and disappeared repeatedly? How often would such events be likely to form and with what frequency?

Detailed dating of pedestal craters (Kadish et al., 2010b; Kadish, 2011) provides an estimate of their formation time as well as how frequently and for how long the ice layer represented by the pedestal remnant had to be in place for the observed distribution of PdS to be produced. The best fit formation time for the observed distribution of PdS is ~90 Myr, but this may be a cumulative time, with the actual formation possibly spread out over a much longer time due to the coming and going of the ice-rich layer. Crater size-frequency distribution data on the Pd surface from HiRISE/CTX image data (Kadish et al., 2010b; Kadish, 2011) yields individual ages from 1 Myr to 3.6 Gyr, with a median of 140 Myr, and with 70% less than 250 Myr old. Of note is the fact that during the period 25 Myr to 175 Myr (a 150 Myr time period) there is at least one Pd emplaced every 15 Myr. The pattern that emerges is that the pedestal craters were likely to have been emplaced into a fluctuating ice layer with a period that is no longer than approximately 15 Myr. Assuming (1) this periodicity, (2) the 90 Myr necessary to form the pedestal craters, and (3) that mass balance variation is approximately sinusoidal (half accumulation, half ablation), we arrive at ~180 Myr (12 cycles).

Given these guidelines, and our goal of understanding the difference between the single and multiple episode cases (Fig. 9), we turn to a 1D flowband model based on the University of Maine Ice Sheet Model (UMISM) (Fastook, 1993; Fastook et al., 2004, 2008). UMISM is a shallow-ice model that we have coupled with an advection-based model of surface debris transport. Whenever the ice surface drops below the crater rim crest, we assume that debris is deposited locally on the ice surface in the crater interior, and that the debris is then transported forward with the movement of the downward and inward-flowing ice. UMISM is fully time-dependent and can track the evolution of the ice surface as it responds to changing climate. To characterize these two cases we perform experiments to see how long it takes for the crater to fill to a level that matches an observed CCF crater (Levy et al., 2010a) shown in Fig. 2. The white line in Fig. 2a indicates the track A–A’ shown as an HRSC DTM profile in Fig. 2b. The inset in Fig. 2b shows an estimated unfilled crater shape and the potential depth of CCF (Garvin et al., 2002).

For the single-episode case, we model a rimless crater to optimize filling from the outside and to simulate older craters with degraded or missing rim crests. As we discussed previously, the ice layer must overtop the rim crest in order for ice to flow in from the surrounding intercrater terrain to supply ice to fill the crater. If the only source is the ice on the rim crest, it is impossible to fill the crater to the observed level. Alternatively, one may think of the ice thickness as being the amount by which the layer must exceed the rim crest height. We begin with a blanket of layer with two values: first, a uniform thickness of 200 m, and secondly, with the typical average Pd crater pedestal heights of ~50 m. In both cases, we hold the thickness to be fixed at the boundary of our model domain. Holding the boundary thickness fixed is equivalent to providing an infinite reservoir of ice that can flow into the model domain to replenish that which has flowed into the crater. As ice is removed from the top of the crater wall as it flows down into the crater, the lowered surface provides a sloping surface from the fixed domain edge that can draw ice from outside the model domain. A comparison between the model results (black) and the observed CCF crater profile (red) is shown in Fig. 11a. At a temperature of 215 K this process takes ~11 Gyr for the model crater to fill to the level in the CCF crater, approximately 800 m thick. Warmer temperatures would accelerate this filling process; if the temperature were raised to 225 K, rates would increase by a factor of 4 with a 250 Myr filling time. Warming to 235 K yields another factor of 4, with a filling time of 65 Myr. On the other hand, in the second case, the use of a lower layer thickness more in keeping with that indicated by the Pedestal Craters (~50 m) increases the filling time to the point where the crater was only 6% full after 3 Gyr (where 50% is the target fill percentage required to match the CCF crater from Fig. 2). In summary, given the excessively long time required for the crater to fill to an appropriate level, we interpret this candidate scenario as being unrealistic for CCF formation. More importantly, if we relax the constraint of having boundary thickness fixed (equivalent to providing an infinite ice reservoir from the crater rim), it is clear that although ice flow from upper wall sources will still be very slow by terrestrial standards, the process is completely supply-limited and the crater will never fill to the observed level in a single phase. Fig. 11a shows the level of crater floor fill after 1.1 Gyr at 215 K; in the final state, the initial ~200 m thick uniform blanketing layer has formed an ~800 m thick layer on the crater floor that is almost exclusively ice that flowed in from the persistent 200 m on the intercrater terrain. Since the ice layer is persistent and continuous even as it flows down the crater walls, less debris is deposited on its surface.

For the multiple-episode case (Fig. 9d–f) we subject this ice sheet/debris model to a climate driven using an obliquity scenario shown in Fig. 12 (Laskar et al., 2004), with repeated cycles of ice-layer formation during the Amazonian when the pedestal craters...
formed (Kadish et al., 2010b; Kadish, 2011). Obliquity predictions are only robust for the last 20 Ma, prior to which the solutions are chaotic (i.e., extremely sensitive to initial conditions, but not random). The chosen scenario is one in which the mean obliquity is relatively high from 40 Myr until 5 Myr, at which point it drops to its current value. Fig. 12b shows an enlargement of the transition period from 7.5 to 2.5 Myr. We chose an obliquity threshold of $35^\circ$, above which we consider the ice layer to have a positive mass balance of 1 mm/yr, typical of GCM results for high obliquity (Forget et al., 2006; Madeleine et al., 2009), and below which we ablate the ice, with the rate depending on the amount of debris armoring the surface (e.g., Fastook et al., 2013). Prior to 5 Myr, the mean obliquity is above this threshold, although there are regular excursions to lower obliquity during which the unarmored portion of the layer would partially or completely sublimate away. After 5 Myr, the mean obliquity drops below the $35^\circ$ threshold, although there continue to be briefer and less frequent excursions above the threshold, during which an ice layer is deposited. Note that after 3 Myr there are no further obliquity excursions that exceed the threshold, and that the ice layer is in a continual state of sublimation.

We limit the deposited layer to a specified thickness of 50 m by turning off the accumulation when the layer volume has reached its "supply-limited" value. Again, we use a rimless topography, since in the case of episodic layer formation, the source of the ice that flows into the crater will be that deposited on the steep crater walls as the layer is draped over the terrain during each of many episodes. Even at a temperature of 215 K, ice is able to flow down the steep slopes of the crater walls into the crater interior, resulting in thicker ice there and thinner ice on the slopes and inter-crater terrain. During low obliquity periods of negative mass balance (Fig. 9c), ice on the steeper unarmored slopes (and rim and inter-crater terrain) may be entirely removed, but the thicker ice in the crater interior, now below the height of the crater walls and subject to being covered by armoring debris from those walls, may not all be removed and the crater can fill with ice and transported debris (Kress and Head, 2008). Indeed, as much as 3% of the crater volume can be deposited in each obliquity episode that exceeds the threshold, rapidly filling the crater to its rim. What is deposited in the crater must, however, survive the period of continuous sublimation that occurs after 3 Myr (Fig. 12) when there are no obliquity episodes that exceed the threshold. The start-and-stop nature of the forward motion of the ice dictates that the transported debris layer will not be uniform in thickness.

Fig. 11. Glacial ice surface (blue-dashed lines) filling a crater interior (assumed profile as a black line) for different formation assumptions, compared with the surface of the CCF-containing crater (surface as a red line, see Fig. 2 and Levy et al. 2010a). (a) Comparison between simulated crater filling from a persistent 200 m layer after 1.1 Gyr. (b) Simulated crater driven by the 301003BIN_A_P001_N obliquity solution (Laskar et al. 2004) with an additional green line showing the surface with its accumulated debris cover. The excess time required for formation with a persistent layer argues against this (a) scenario. The obliquity-driven scenario (b) fills the crater in just 50 Myr, and in addition, the start-stop nature and accumulated debris from the crater walls generates a non-uniform surface similar to that observed in CCF-containing craters. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 12. Laskar et al. (2004) obliquity scenario 301003_BIN_A_P001_N. (a) Obliquity for the last 50 Ma predicted in this scenario. (b) Obliquity for the period 7.5 Ma to 2.5 Ma predicted in this scenario, indicated with grey vertical band in (a). During this period the mean obliquity shifts from high to low. The horizontal green line shows the $35^\circ$ obliquity threshold above which an ice layer is deposited. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
The final state of this 50 Myr simulation, driven by the obliquity signal of Fig. 12, is shown in Fig. 11b. The ~800 m of fill is attained in less than one-twentieth the time it takes for an infinite persistent layer to flow into the crater to a comparable depth (Fig. 11a). For comparison, in an analysis of the generation of debris cover on lobate debris aprons, Fastook et al. (2013) estimated a time of ~400–500 kyr between the point that a receding regional ice sheet first exposed bedrock outcrop at the dichotomy boundary, and complete debris cover of the lobate debris apron. The cyclic, start-and-stop nature of this process is very likely to be directly correlated with formation of the concentric ridges observed in the CCF (Fig. 2a), a correlation also suggested for very slow moving debris-covered glaciers in the Antarctic Dry Valleys (e.g., Marchant et al., 2010; Shean et al., 2007b; Shean and Marchant, 2010).

Since sufficient ice is deposited at each episode even at 215 K, and since the episodes are so frequent and short in duration, warmer temperatures have little effect on the formation time. The warmer ice simply flows down the crater walls into the interior that much faster. Since there is no inflow from outside the crater rim crest, the only contribution to the interior is that portion of the layer deposited on the steep crater wall. A reasonable fit to the CCF crater surface can be obtained by starting the simulation as late at 10 Myr, but knowing that the obliquity solution is robust at least back to 20 Myr, and given our understanding of the formation times of the Pd (Kadish et al., 2010b; Kadish, 2011), we have taken a conservative course, and demonstrate that this mechanism can be started as far back as 50 Myr and still obtain a good fit to the present configuration.

3. Summary of analysis of key questions

In summary, snow and ice cover and glaciation during periods of maximum ice accumulation in the Late Amazonian appear to be characterized by episodic regional (mid- to high-latitude) deposits of snow and ice, of the order of 50 m thickness. The small regional thickesses, and the temperature-dependent rheology of ice and its extremely slow flow under typical Late Amazonian conditions (Fig. 4), mean that slopes steeper than typical intercrater regions (Fig. 5) are required to induce flow. The global distribution of slopes shows that the steep slopes conducive to glacial ice flow are found primarily within crater interiors (the steep crater interior walls below the rim crest) (Fig. 5c–g). The typical rim crest height and the outward facing slopes of the uplifted rim and crater ejecta topography (Fig. 10) mean that any flow induced on these slopes is very likely to be away from the crater, not into the crater. Thus, glacial flow in the Late Amazonian is likely to be concentrated in microenvironments in crater interiors, as snow and ice deposits preferentially flow down the steep interior wall slopes.

This particular crater interior microenvironment is also conducive to the preferential preservation of the deposition and flow of glacial ice because of the close proximity of arming material in the steepest parts of the crater walls. Mass wasting of fractured and fragmented uplifted target material and crater rim ejecta provides a ready source of debris from some of the steepest slopes on Mars. Deposition of talus onto the accumulating and flowing ice provides a rock and soil debris cover that will become concentrated into a sublimation till cover through sublimation of ice; thickening of the debris-rich sublimation till will progressively retard or halt ice-loss in periods of net regional sublimation, preserving the ice below. This process is likely to vary along the flow direction, as observed in terrestrial debris-covered glaciers (Marchant et al., 2010), with newly deposited ice having the little to no sublimation till and earlier deposited ice being covered with a progressively thickening sublimation till.

This formation scenario predicts that: (1) ancient glacial ice will be preserved preferentially at the base of crater interior walls as it flows from the steep walls to the floor; (2) this ice will be characterized by a debris cover shed from the steep interior walls, with more pure ice toward the proximal accumulation zone and debris concentrated in the distal parts in progressively thicker sublimation tills; (3) these debris-covered glaciers should reflect the waxing and waning of glacial conditions through surface textures and concentric ridges representing greater and lesser debris cover, sublimation and flow rates, (4) in major interglacial periods, these debris-covered glaciers are likely to be one of the most prominent records of these glacial periods due to their propensity to flow on steep slopes, and the ready supply of debris that arms the surface against ice-loss during these interglacials, (5) during interglacial periods, thinly covered ice in the proximal zones may sublimate, creating spatuale depressions (Head et al., 2008) and when glacial conditions return, ice will again accumulate and additional flow of the debris-covered glacier is likely to occur, (6) due to the long-term constancy of the major topographic features of Mars, concentric crater fill deposits are likely to be the preferred sites of both formation and preservation of the record of glacial epochs on Mars.

We have outlined a simple scenario for the formation and evolution of crater interior glacial deposits (concentric crater fill) derived from deposition of an initial ~50 m thick ice layer distributed uniformly across the crater interior, rim and intercrater regions (Fig. 9). Although this model has provided insight into the range of likely glacial flow processes, their time-scales and their manifestation as deposits, the actual scenario is likely to be more complicated and regionally varied. For example, the topographic configuration of an impact crater structure (raised rim and deep, steep walled interior) is not likely to produce deposition of a passive draped layer of even thickness. First, during glacial epochs, global insolation patterns and surface temperatures (Haberle et al., 2003; Kreslavsky et al., 2008, their Fig. 6), and thus snow and ice emplacement patterns, will be locally influenced and modulated by the presence of this depression. Furthermore, regional weather trends, as well as surface wind velocity and direction, will influence total snow accumulation and location within craters. For example, wind-blow snow will certainly result in significant local deviations in snow thicknesses, favoring deposition in the topographic traps of crater interiors. Craters themselves can also create distinctive climate microenvironments (Madeleine et al., 2012) in which atmospheric temperature profiles can vary and locally enhance or inhibit snow precipitation. Finally, preservation of snow and ice on yearly and longer time scales is clearly influenced by crater topographic landforms and their relation to insolation geometry (Kreslavsky and Head, 2003; Kreslavsky et al., 2008; their Fig. 6), as well as crater size, slopes (Kreslavsky and Head, 2006) and depth-diameter relationships (Beach and Head, 2012, 2013). Indeed, Dickson et al. (2012) have shown that concentric crater fill deposits in the 45°–60° latitude bands show concentric flow, while in the 25°–45° latitude bands CCF flow directions are oriented asymmetrically, with poleward flow heavily favored, indicating a preference for ice preservation on colder, pole-facing slopes.

4. Conclusions

We combined a variety of geological evidence (CCF characteristics, Pd thicknesses and distribution, current polar cap volumes, regional slopes) and ice flow-modeling approaches to address four fundamental questions concerning the nature of Amazonian mid- to high-latitude glaciation on Mars in order to assess ice sources, ice accumulation patterns, and glacial flow and ice sequestration in impact crater concentric crater fill. These questions included (1) What ice thickness is required to initiate ice flow on a flat, intercrater area under Late Amazonian environmental conditions?
(2) Could the current Pd mean thickness value (~50 m) be the remnant thickness of equilibrium flow? (3) What slopes are required to initiate ice flow under Late Amazonian conditions and where do such slopes occur geologically? (4) What was the nature of snow and ice cover and glaciation during periods of maximum ice accumulation in the Late Amazonian?

Regional ice flow under Late Amazonian climate conditions (Fig. 4) requires ice thicknesses exceeding many hundreds of meters in the mid- to high-latitudes for slopes typical of the vast majority of the surface of Mars (Fig. 5). This thickness, distributed across the mid-latitudes (Fig. 8), is well in excess of the volumes of available polar ice reservoirs. This indicates that although conditions for mid- to high-latitude glaciation may have long persisted, the process is “supply limited”, with a steady state reached (Fig. 8) when the polar ice cap water ice supply becomes exhausted.

Global slope analysis (Fig. 5) shows that impact craters are by far the most abundant landform with associated slopes (interior walls) sufficiently high to induce flow under Late Amazonian climate conditions. Furthermore, topographic slope data show that Amazonian impact craters have been clearly modified, undergoing crater interior slope reduction and shallowing of crater floors. 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 spin–axis obliquity. Flow from a single episode of an inter-crater terrain layer compatible with Pedestal Crater deposit thicknesses (~50 m) is insufficient to fill craters in a time period compatible with the interpreted formation times of mantled ice layers beneath pedestal craters. A representative obliquity solution (Fig. 12) driving an ice flow model shows that a cyclical pattern of multiply recurring layers can both (1) fill the craters with a significant volume of ice, and (2) transport debris from the crater walls out into the central regions of the craters (Fig. 11b). The model predicts that the cyclical pattern of waxing and waning results in a rippled pattern of surface debris extending out into the crater interiors, comparable in appearance to concentric crater fill surface textures. In this model, the accumulating ice forms a mantling sublimation till layer that seals and sequesters it from significant temperature variations at diurnal, annual and spin–axis/orbital cycle time scales. We have driven the simulation with a representative obliquity solution (Fig. 11b) where the layers are assumed to form when obliquity is above a 35° threshold, providing insight into which of the many predicted solutions (e.g., Laskar et al., 2004) is the most likely to have occurred.

Observations suggest that this process has produced ancient ice records preserved today below CCF crater floors. The lack of any associated meltwater features is evidence that the Late Amazonian climate did not exceed the melting temperature in the mid- to high-latitudes for any significant period of time. This process implies that continued sequestration of ice with time in CCF and related deposits (lobe debris aprons and lineated valley fill) further reduces the already supply-limited polar ice sources. This suggests that there has been a declining reservoir of available ice with each ensuing Late Amazonian glacial period. Concentric crater fill deposits represent a candidate library of climate chemistry and global change dating from the Late Amazonian, and a non-polar water resource for future exploration.

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