Evidence for Amazonian northern mid-latitude regional glacial landsystems on Mars: Glacial flow models using GCM-driven climate results and comparisons to geological observations

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A fretted valley system on Mars located at the northern mid-latitude dichotomy boundary contains linedated valley fill (LVF) with extensive flow-like features interpreted to be glacial in origin. We have modeled this deposit using glacial flow models linked to atmospheric general circulation models (GCM) for conditions consistent with the deposition of snow and ice in amounts sufficient to explain the interpreted glaciation. In the first glacial flow model simulation, sources were modeled in the alcoves only and were found to be consistent with the alpine valley glaciation interpretation for various environments of flow in the system. These results supported the interpretation of the observed LVF deposits as resulting from initial ice accumulation in the alcoves, accompanied by debris cover that led to advancing alpine glacial landsystems to the extent observed today, with preservation of their flow texture and the underlying ice during downwasting in the waning stages of glaciation. In the second glacial flow model simulation, the regional accumulation patterns predicted by a GCM linked to simulation of a glacial period were used. This glacial flow model simulation produced a much wider region of thick ice accumulation, and significant glaciation on the plateaus and in the regional plains surrounding the dichotomy boundary. Deglaciation produced decreasing ice thicknesses, with flow centered on the fretted valleys. As plateaus lost ice, scarps and cliffs of the valley and dichotomy boundary walls were exposed, providing considerable potential for the production of a rock debris cover that could preserve the underlying ice and the surface flow patterns seen today. In this model, the lineated valley fill and lobate debris aprons were the product of final retreat and downwasting of a much larger, regional glacial landsystem, rather than representing the maximum extent of an alpine valley glacial landsystem. These results favor the interpretation that periods of mid-latitude glaciation were characterized by extensive plateau and plains ice cover, rather than being restricted to alcoves and adjacent valleys, and that the observed lineated valley fill and lobate debris aprons represent debris-covered residual remnants of a once more extensive glaciation.

1. Introduction

1.1. Background and evidence for climate change on Mars

Exploration of Mars over the decades has shown that there are significant volumes of water sequestered in various surface deposits at different latitudes, such as the polar layered terrains (e.g., Thomas et al., 1992; Keiffer and Zent, 1992; Phillips et al., 2008), in the shallow subsurface at high latitudes (e.g., Boynton et al., 2002; Feldman et al., 2002), in the deeper cryosphere (e.g., Clifford, 1993), and even in mid-latitude alpine-valley glacial landsystems (e.g., Head et al., 2010). Furthermore, analysis of the geological record of Mars shows a rich array of Amazonian and Hesperian-aged features and deposits that involved water, sometimes in huge amounts (e.g., Clifford, 1993; Carr, 1996; Masson et al., 2001). Noachian-aged surfaces display valley networks that suggest a “warmer and wetter” early Mars (e.g., Craddock and Howard, 2002; Fassett and Head, 2008). Clearly, the climate has changed during martian history from a time when liquid water flowed across the surface in valley networks and pluvial activity may have occurred (e.g., Craddock and Howard, 2002; Fassett and Head, 2008), to the very cold and hyperarid surface conditions observed today (e.g., Zurek, 1992; Head et al., 2003a; Carr and Head, 2010).

Coincident with this long-term climate change has been a transition in the global hydrological cycle from an early Mars when it appears that the hydrological cycle was vertically integrated (e.g.,
Carr, 1996; Clifford and Parker, 2001; Baker, 2001; Head et al., 2003b) to one in which the hydrological cycle is horizontally stratified, with a cryosphere separating the groundwater system from the surface system. The surface system then consists of three major water reservoirs: (1) surface deposits (currently the polar caps), (2) the atmosphere, and (3) the regolith, and most of the movement of water is associated with seasonal and longer-term climate changes related to spin-axis/orbital variations (e.g., Laskar et al., 2004).

In addition to these long-term global climate trends, there is abundant evidence for shorter-term climate fluctuations driving the surface water cycle related to variations in these spin-axis/orbital parameters, such as obliquity, eccentricity and precession (e.g., Laskar et al., 2004). The discovery and documentation of geologically recent gullies (e.g., Malin and Edgett, 2000) and further documentation of Amazonian tropical mountain glaciers (e.g., Williams, 1978; Head and Marchant, 2003; Shean et al., 2005, 2007; Milkovich et al., 2006; Kadish et al., 2008) are among the growing evidence suggesting that variations in orbital parameters may cause climate excursions resulting in the redistribution of water and a Mars very different from that typical of very recent history. Indeed, Laskar et al. (2004) predict that throughout much of its history Mars may have been characterized by an obliquity significantly higher than its current value.

One of these high obliquity states modeled by Forget et al. (2006) (for obliquity at 45° with a source of moisture at the present poles) produced regions of persistent ice accumulation on the northern part of Arabia Terra (Sharp, 1973), and Merrick (1978) noted that regions of ice accumulation were present throughout the history of Mars. Furthermore, geological observations (e.g., Malin and Edgett, 2000) predict that throughout much of its history Mars may have been characterized by an obliquity significantly higher than its current value.

Fretted terrain and fretted channels, well developed in the northern part of Arabia Terra (Sharp, 1973), were clearly identified in Viking data as the location of lobate debris aprons (LDA) and linear-laterally eroded valley fill (LVF) (Squyres, 1978, 1979). Carr and Schaber (1977) analyzed Viking Orbiter images, concluding that frost creep and gelifluction were the primary processes in LDA formation. Squyres (1978) documented the fact that debris aprons (LDA), characterized by sharply-defined flow fronts and convex-upward surfaces, extend from massifs and escarpments outward to distances of up to ~20 km. Squyres (1978, 1979) interpreted the aprons to have resulted from mass-wasted debris that flowed due to the presence of interstitial ice, envisioning that during periods of climate change, atmospheric water vapor would diffuse into talus piles formed at the base of steep topography, and interstitial ice would cause mobilization and flow of the talus to produce the lobes. Longitudinal ridges and grooves in surface debris seen on the floors of the fretted valleys (LVF) were interpreted to form by ice-assisted debris aprons flowing from valley wall talus slopes and converging in the middle of the valley floor (Squyres, 1978, 1979). A number of workers noted that some LVF in the Nilosyrtis region appeared consistent with down-valley flow; following earlier ideas by Kochel and Baker (1981). Kochel and Peake (1984) interpreted some debris aprons as formed by processes similar to those of terrestrial ice-cored or ice-cemented rock glaciers and further noted a transition from flow-parallel to flow-perpendicular ridge-and-furrow topography. Lucchitta (1984) interpreted the LDA and LVF as flow of debris with interstitial ice and showed evidence of local down-valley flow of LVF, likening some examples to flow patterns in glacial ice in Antarctica. Carr (1996, pp. 116–120) urged caution in the general interpretation of a range of landforms as being due to glaciation. He cited: (1) the fact that the glacial hypotheses requires major climate change late in Mars history (e.g., Baker et al., 1991) for which he found little supporting evidence and (2) that the low erosion rates in the Amazonian argue against any major climate excursions accompanied by precipitation, as envisioned by Baker (2001).

More recently, Pierce and Crown (2003) used new image and altimetry data and found evidence for a wide range of possible ice systems on Mars (e.g., rock glacier ice-assisted creep of talus, ice-rich landslides, debris-covered glaciers), with typical LDA resulting from flow of debris that was enriched in ground ice. Mangold (2003) showed that ice content in LDA may exceed pore space, and Li et al. (2005) compared LDA MOLA profiles to simple plastic and viscous power law models for ice–rock mixtures, concluding that LDAs were ice-rich rock mixtures with some perhaps >40% ice by volume. Chuang and Crown (2005), documenting the detailed character of 65 LDA, concluded that they consisted of mixtures of debris and ice, but that it was “difficult to constrain the internal distribution of ice or the method of debris apron initiation from the current datasets.” In summary, all workers agree that formation of LDA and LVF involve both talus and ice, but there is disagreement as to the amount of ice: One end-member calls on formation by ice-assisted creep of talus (often defined as producing a “rock glacier”) (e.g., Squyres, 1978, 1979), while another end-member calls on formation as debris-covered glaciers, (predominantly glacial ice with a cover of sublimation lag or tills) (e.g., Head et al., 2005, 2006a,b).

Head et al. (2010) developed criteria to distinguish between these end-members and investigated a wide range of occurrences to assess the origin of LDA and LVF. On the basis of the background studies described above, and using a range of terrestrial analogs applicable to the recent cold-desert environment of Mars (e.g., Marchant and Head, 2006, 2007), Head et al. (2010) developed the following criteria to assist in the identification of debris-covered glacial-related terrains on Mars (features, followed by the interpretation of each, listed in parentheses): (1) alcoves, theater-shaped indentations in valley and massif walls (local snow and ice accumulation zones and sources of rock debris cover); (2) parallel arcuate ridges facing outward from these alcoves and extending down slope as lobe-like features (flow-deformed ridges of debris); (3) shallow depressions between these ridges and the alcove walls (zones originally rich in snow and ice, which subsequently sublimated, leaving a depression); (4) progressive tightening and folding of parallel arcuate ridges where abutting adjacent lobes or topographic obstacles (constrained debris-covered glacial flow); (5) progressive opening and broadening of arcuate ridges where there are no topographic obstacles (unobstructed flow of debris-covered ice); (6) circular to elongate pits in lobes (differential sublimation of surface and near-surface ice); (7) larger tributary valleys containing LVF formed from convergence of flow from individual alcoves (merging of individual lobes into LVF); (8) individual LVF tributary valleys converging into larger LVF trunk valleys (local valley debris-covered glaciers merging into larger intermontaine glacial systems); (9) sequential deformation of broad lobes into tighter folds, chevron folds, and finally into linear-laterally eroded valley fill (progressive glacial flow and deformation); (10) complex folds in LVF where tributaries join trunk systems (differential flow velocities causing folding); (11) horseshoe-like flow lineations draped around massifs in valleys and that open in a down-slope direction (differential glacial flow around obstacles); (12) broadly undulating along-valley topography, including local valley floor highs where LVF flow is directed away from individual centers of accumulation; (13) integrated LVF flow systems extending for tens to hundreds of kilometers (intermontaine glacial systems); (14) rounded valley wall corners where flow converges downstream, and narrow arete-like plateau remnants.
between LVF valleys (both interpreted to be due to valley glacial streamlining). Taken together, the occurrence of a number of these types of features in an area is interpreted to represent the former presence of debris-covered glaciers and valley glacial systems in the Deuteronilus–Protonilus region (Head et al., 2010).

Detailed geological analysis of several different areas in the northern mid-latitudes (Fig. 1a) has shown strong evidence for the presence of valley glacial landsystems (Eyles, 1983; Evans, 2003) during the Amazonian and the sequestration and preservation of debris-covered glacial ice (Head et al., 2005a,b, 2010; Levy et al., 2007; Kress and Head, 2008; Morgan et al., 2009; Dickson et al., 2008, 2010; Hauber et al., 2008; Marchant and Head, 2008; Baker et al., 2010), interpretations supported by the detection of extensive buried ice by the SHARAD (SHAllow RA-Dar) instrument on board the Mars Reconnaissance Orbiter (e.g., Holt et al., 2008a,b; Plaut et al., 2009, 2010). Further atmospheric general circulation modeling by Madeleine et al. (2009) for an obliquity of 35° with the moisture source in the Tharsis region, rather than the poles, produced persistent ice accumulation along the mid-latitude region and the dichotomy boundary, also consistent with the presence of these geological features indicative of valley glaciation documented there.

In this contribution we use the results of the LMD/GCM (Madeleine et al., 2009) that reproduce conditions that favor ice accumulation in the mid-latitudes. From these predictions, we develop glacial flow models to assess the accumulation and flow patterns of ice to compare to the patterns documented by the geological observations, to test their validity, to further test the distinctions between the (1) ice-rich debris creep and (2) debris-covered glacier models for LDA and LVF, and if successful, to learn more about the nature of interpreted Amazonian valley and regional glaciation.

1.2. Geological observations

A unique fretted terrain, located along the northern mid-latitude dichotomy boundary (Fig. 1a), has long been held to provide clues to the processes responsible for the formation and degradation of the dichotomy boundary (Sharp, 1973). Parts of these fretted valleys (Fig. 1b and c) display a multitude of characteristics typical of integrated valley glacial landsystems on Earth (e.g., Eyles, 1983; Evans, 2003), including multiple theater-headed, alcove-like accumulation areas (Fig. 1d); sharp arete-like ridges typical of glacial erosion; converging patterns of downslope valley flow (Fig. 1e); valley lineation patterns typical of folding and shear (Fig. 1e); wrap-around features indicative of flow around obstacles; and broad piedmont-like lobes as the valley fill extends out into the northern lowlands (Head et al., 2006a, 2010). These characteristics suggest that the boundary area was subjected to large-scale glaciation during the Amazonian. Recognition and documentation of this important late-stage process provides information critical to reconstructing the original dichotomy boundary, and to the understanding of Amazonian climate history.

Fig. 1. (a) Northern mid-latitudes dichotomy boundary region showing the locations of key published study areas (1: Kress and Head, 2009; 2: Head et al., 2006b; 3: Head and Marchant, 2009; 4: Morgan et al., 2009; 5: Head et al., 2006a; 6: Baker et al., 2010; 7: Dickson et al., 2008; 8: Levy et al., 2007; 7/ Head et al., 2010). (b) Fretted channel and lineated valley fill system in the central Deuteronilus–Protonilus Mensae (DPM) described in Head et al. (2006a) (box 5 in (a)). Viking Orbiter mosaic and location map. Boxes and letters in (b) indicate locations of illustrative images in Head et al. (2006a). (c) Map showing the location and trend of small alcoves and lobate flow-like features (narrow arrows) and broad flow trends of LVF on the valley floor (wide arrows). Compiled using HRSC orbit 1395, THEMIS and MOC images. (d and e) Images illustrating the nature of alcoves and LVF features associated with the system shown in (b and c). (d) Four adjacent alcoves oriented convex-inward toward the plateau (bottom) with individual, convex outward lobes extending from each alcove out into the adjacent LVF. Note the convergence of these individual lobate extensions, their compression and deformation, and their convergence with the general trends of the LVF on the valley floor (left). Location is box A in (b), THEMIS V112080010. (e) Outflow of two ridged lobes from alcoves (bottom left and right); lobes join major LVF of area C (upper right) near convergence with B (c). Left lobe is swept westward, forming broad arcuate folds; right lobe is increasingly compressed until it resembles a tight isoclinal fold. Both lobes ultimately merge into the general LVF parallel to the valley walls. Location is box D in (b). MOC R1702578.
1.3. Interpretation as a valley glacial landsystem

Detailed analysis of a typical fretted valley network at ~34°E, 41°N in the central region between Deuteronis and Protonilus Mensae (Fig. 1a, box 5) was undertaken by Head et al. (2006a) (Fig. 1b and c). This set of fretted valleys extends for over 200 km in a N–S direction, from deep within the upland plateau to the edge of the continuous upland scarp at ~41°42′N. Using MOLA altimetry, THEMIS, HRSC, and MOC images, Head et al. (2006a) traced the LVF from the narrow parts of the deeper upland valleys progressively toward the northern lowlands. In the most inland regions of the fretted valleys analyzed (areas A and B in Fig. 1b) the valleys are narrow (~10–20 km in width), often have distinctively cuspate (convex outward) shoulder-to-shoulder alcoves along the walls (region A) or larger tributary valleys often forming hanging valleys along the wall margins (region B). In the A/B regions (Fig. 1c) the valley floor is about 1500 m above the northern distal end of the deposit.

In contrast to the sharp parallel valley walls and parallel along-valley lineated fill typical of some fretted valleys described elsewhere (e.g., Squyres, 1978; Carr, 1995, 2001), these alcoves are sources of LVF that can be seen to emerge from individual alcoves as lobes (Fig. 1c and d), descending onto the valley floor, bending and turning in a downslope direction (Fig. 1c and e), and merging with adjacent lobes, deforming in the process (Fig. 1c) ultimately to form the distinctive along-strike ridges of LVF. These types of alcoves, the arete-like borders between them, the arcuate-ridged lobes extending from them, the progressive compression and folding of the lobes as they converge onto the valley floor and lose their identity, are all hallmarks of glaciated valley landsystems on Earth (Eyles, 1983; Benn et al., 2003). In the context of such valley glacial environments, the alcoves in the DPM area were interpreted to represent microenvironments where ice and snow accumulate, al- cove walls shed debris onto the ice, and debris-covered glaciers emerge and move downslope into the main valleys, widening the alcoves and creating aretes over time (Head et al., 2006a). Mapped trends show the emergence of lobate material from these alcoves (Fig. 1c, narrow arrows) and the general trends with which they merge into the larger-scale LVF (wide arrows). The patterns in areas A, B, and C (Fig. 1c) clearly indicate the convergence typical of glaciated valley landsystems (Eyles, 1983; Benn et al., 2003) where abundant small debris-covered glaciers contribute to larger-scale valley glaciers. As the generally north-trending LVF emerges from areas A to C (Fig. 1c), the individual LVFs merge into two main trends, areas D and E. LVF C merges with B and continues ~95 km to the NNW where it bifurcates around a mesa to form LVF F and G, each continuing 50–75 km to the northern lowlands. Western portions of LVF B turn west into D, merging with LVF A to produce complex fold patterns. Portions of A turn NW and extend almost 100 km to the northern lowlands. Several basic trends are observed over this region extending from A–B–C to the northern lowlands (Head et al., 2006a). The elevation of the LVF floor decreases by ~1500–2000 m from the proximal areas (areas A–C) to the end of the LVF in the northern lowlands. LVF slopes tilt toward the north and are less than 1° throughout most of the LVF floor, but increase rapidly to >1–2° in the northern 25–50 km of the LVF adjacent to the lowlands. These areas are characterized by piedmont-like lobes extending into the northern lowlands between mesas. These features are reminiscent of the fronts of terrestrial glacial systems in terms of their lobate nature and relatively steep front, and in many cases are characterized by flow lines, pits and moraine-like ridges.

In summary, spacecraft data show compelling evidence for an integrated picture of LVF formation (Fig. 1b–e; Head et al., 2006a) with a significant role being played by regional valley glaciation (e.g., Eyles, 1983; Post and Lachapelle, 2000; Benn et al., 2003) in the modification of the valley systems. There is evidence for: (1) localized alcoves, which are the sources of dozens of narrow, lobate concentric-ridged flows interpreted to be remnants of debris-covered glaciers; (2) depressions in many current alcoves, suggesting loss of material from relict ice-rich accumulation zones; (3) narrow pointed plateau ridge remnants between alcoves, similar to glacially eroded aretes; (4) horseshoe-shaped ridges up-valley and upslope of topographic obstacles; (5) convergence and merging of LVF fabric in the down-valley direction, and deformation, distortion and folding of LVF in the vicinity of convergence, all consistent with glacial-like, down-valley flow (Fig. 1c); and (6) distinctive lobe-shaped termini where the LVF emerges into the northern lowlands, with associated pitting and concentric terminal ridges.

This assemblage of features can be traced and mapped in an integrated fashion (Fig. 1b and c) for over 200 km in length, and covers an area (~30,000 km²), ~10 times that of the Antarctic Dry Valleys (Marchant and Head, 2007), being more comparable in scale to the remnant North American ice caps on Ellesmere and Baffin Islands in the Canadian high Arctic. This assemblage of features suggests that the LVF in this region formed as part of a broad integrated valley glacial system extending for about 200 km in the S–N direction. The extensive development of similar lineated valley fill occurrences, often with very similar characteristics, in fretted valleys in the DPM region (Fig. 1a; see also Head et al., 2010), suggests that glaciation may have been instrumental in the modification of the dichotomy boundary over an area as large as several million km². The age of the LVF in different parts of the DPM region has been estimated as Amazonian (e.g., McGill, 2000; Mangold, 2003; Morgan et al., 2009; Baker et al., 2010), with relatively younger alteration by processes of mantle deposition and surface layer sublimation and modification (e.g., Mangold, 2003; Head et al., 2003a,b; Levy et al., 2009). On the basis of these types of observations, Head et al. (2006a) (and others; Fig. 1a) interpreted the process of alpine-style valley glaciation to have been extensive in this region.

To assess this glacial origin hypothesis for this valley system (Fig. 1b) and the region in general (Fig. 1a), Madeleine et al. (2009) explored the conditions under which GCMs reproduced sufficient ice deposition and seasonal preservation to cause glaciation. They found that the LMD/GCM for a dusty Mars atmosphere with obliquity set to 35° and a water source in the Tharsis region could generate ice accumulation in good agreement with these geological observations. According to Madeleine et al. (2009), this proposed climate is what one might expect to follow a higher-obliquity excursion (Forget et al., 2006; Levrad et al., 2004) of the sort thought to build ice sheets on the flanks of the Tharsis volcanoes (Head and Marchant, 2003). We now use the predictions of the LMD/GCM to formulate glacial flow models to compare to the patterns of LDA and LVF interpreted by Head et al. (2006a) to be due to glaciation, and to test the validity of these interpretations of valley glacial landsystems (Fig. 1a), and to test further the distinctions between the (1) ice-rich debris creep and (2) debris-covered glacier models for LDA and LVF. We then explore the implications of the locations of ice depocenters predicted by the LMD/GCM for more regional glacial landsystems.

2. Glacial flow modeling of the Deuteronis–Protonilus Mensae (DPM) region

2.1. The University of Maine Ice Sheet Model (UMISM)

UMISM is an adaptation for the martian environment (Fastook et al., 2004, 2005, 2006) of a terrestrial ice sheet model used for time-dependent reconstructions of Antarctic, Greenland, and
paleo-ice sheet evolution in response to changing climate on Earth (Fastook, 1993). UMISM uses a thermo-mechanically coupled Shallow-Ice Approximation (vertically-integrated momentum combined with continuity) where the dominant stress is internal shear and longitudinal stresses are neglected. Primary input to the model is the bed on which the ice sheet is to be reconstructed and the net annual surface mass balance (SMB), or accumulation rate. Secondary input includes the mean-annual surface temperature and the geothermal heat flux, used to calculate internal temperatures from which the mechanical properties of the ice are obtained. In addition, internal temperatures allow for the possibility that the base of the ice reaches the pressure melting point, at which point some sliding criteria can be invoked, a phenomena we have not observed in any modeled martian glaciers (Fastook et al., 2008a,b).

The fact that with the exception of the bed, these inputs are poorly constrained for Mars introduces uncertainty into the results that will be presented. However, choice of reasonable values for mean annual temperature, geothermal flux, and SMB still allow production of results comparable to the geologic observations. Given these uncertainties the choice of a Shallow-Ice Approximation model is also appropriate, since the considerable computational load of a higher-order model would not produce more accurate results.

2.2. Modeling: alcove-only accumulation areas

As a preliminary step in the modeling of the DPM interpreted glaciation (Head et al., 2006a), we first examine a simple case where accumulation only occurs in the alcoves along the valley walls. This simple treatment has terrestrial analogs in the Dry Valleys of Antarctica (Marchant and Head, 2004, 2005, 2006, 2007). In particular, we follow a model of ice deposition observed in the Dry Valleys of Antarctica on Mullins Glacier, a debris-covered glacier that originates in an alcove analogous to the martian alcoves (Kowalewski et al., 2011). Here deposition of ice occurs only in a very limited area at the base of a scarp that also serves as the source of the surface debris. Without this protective layer of rock debris, the glacier would sublimate rapidly as it flows down into the Dry Valleys, an area where normal sublimation removes any ice exposed at the surface. Marchant and Head (2004, 2005, 2006, 2007) and Marchant et al. (2010) have shown that buried ice in the Mullins Glacier may be several millions of years old, a situation that would be not be possible without the armoring effect of the debris cover (Kowalewski et al., 2006).

In reconstructions of Tharsis Montes ice sheets, we used either a parameterization of the SMB as a function of height (Fastook et al., 2004, 2005) or GCM results (Fastook et al., 2006) as the source of mass in the continuity equation. Here we chose the expedient of simply specifying a positive SMB of +1 mm/year in the alcoves, and a negative SMB 1/10th of that outside the alcoves (Fig. 2). These values are typical of debris-covered glaciers in the Dry Valleys, Antarctica (Kowalewski et al., 2011). The shape of the resulting profile is relatively insensitive to the magnitude of the SMB. SMB is raised to the 1/8th power in analytic solutions for uniform SMB elliptical profiles, implying a thickening of only 10% for a doubling of the SMB. Doubling SMB effectively doubles velocity because thickness is only increased by 10% and the increased flux must be accommodated by increased velocity. Changing SMB does not significantly affect the orientation of the flow field because flow direction is down the surface gradient and surface changes are small relative to changes in the SMB. These
higher or lower velocities will result in faster or slower advance of the glacier during the formation, and of course will affect retreat times during collapse. There are no constraints on either the formation times or the velocity magnitudes, so the simple expedient of +1 mm/a in the alcoves and /C0.1 mm/a elsewhere was chosen to yield a 10:1 ratio between accumulation area and ablation area.

Starting with no ice, the model is run for 2 Ma, which delivers a flow pattern (primarily the orientation of the velocity field) that can be compared to the Head et al. (2006a) glacial interpretation. Resulting ice thicknesses and ice flow velocities, with superimposed surface elevations, are shown in Figs. 3 and 4 for four time intervals during the growth of these ice deposits.

(1) At 300 Ka (Fig. 3, upper left), the ice is thin (less than 300 m; blue areas) and the flow from the sides is at low velocity (less than ~50 mm/a; Fig. 4, upper left). The flow has not yet merged along the center of the valleys (blue areas in Fig. 3 upper left). A configuration such as this would not yet produce the turning flow observed by Head et al. (2006a) (Fig. 1).

(2) By 500 Ka (Fig. 3, upper right), the beginning of a coherent down-valley flow is observed (green and blue areas in the lower right-hand and central part). The ice flowing from

\[ \text{Fig. 3. DPM lineated valley fill system color-coded thicknesses (m) at 300, 500, 1000, and 1500 Ka. Contour lines indicate surface elevation.} \]
(3) By 1000 Ka (Fig. 3, lower left), there is a well-established valley glacier system extending from the fretted valleys in the south to the mouths of the valleys along the dichotomy boundary, with thicknesses exceeding 400–500 m in the southern part of the valley system. Velocities (Fig. 4, lower left) are almost everywhere higher than 50 mm/a, and locally in excess of ~200 mm/year.

(4) As time progresses to 1500 Ka (Fig. 3, lower right), the valley glaciers become very well developed in the valleys and extend out of the valleys and onto the northern lowlands. Ice thicknesses are locally in excess of 600 m. Velocities (Fig. 4, lower right) are broadly in excess of 100 mm/a, and locally exceed 200 mm/a.

Comparison of time steps in Fig. 3 shows that ice accumulation and flow at either 1000 Ka (Fig. 3, lower left) or 1500 Ka (Fig. 3, lower right) would clearly be producing the kinds of landforms...
observed by Head et al. (2006a) and interpreted to be glacial in origin (compare to patterns in Fig. 1b–e).

We can further examine these correlations and test the glacial interpretation by comparing smaller areas of accumulation and flow orientations to the patterns observed by Head et al. (2006a). High resolution excerpts from the glacial flow model results, for both thickness and velocity at 1000 Ka and at 1500 Ka (Figs. 3 and 4), are compared to patterns in the THEMIS images of regions BC, F, and G from Fig. 1b (Head et al., 2006a) in Figs. 5–7. The velocity vector is always perpendicular to the surface elevation contours and aligns well with the interpreted flow feature orientations.

The region furthest from the dichotomy boundary and the valley mouths (area BC in Fig. 1b, THEMIS VIS V11208010) is characterized by eastward-opening adjacent alcoves with multiple convex-outward concentric-ridged lobes extending, converging, deforming, and merging with the valley floor LVF trends. Here the model at 1000 and 1500 Ka shows extremely low surface slopes (surface contours are only 50 m). Ice is 300–500 m thick, and velocities are generally less than ~50 mm/a. The flow is directed away from the alcoves with little indications of turning down-valley.

A region further down the valley toward the dichotomy boundary (area G in Fig. 1b) is away from the alcoves and in the central part of the valley system (Fig. 7). LVF bifurcates (bottom right) and flows around a massif forming a broad up-flow collar and a diffuse, down-flow “wake”. LVF in the narrow pass between massifs is compressed. The modeled flow clearly narrows as it pinches between the two bedrock highs and accelerates in these areas to values in excess of 200 mm/a.

In summary, the glacial flow and velocity orientation patterns in the ice flow model simulation clearly match many of the features observed in the lineated valley fill, interpreted by Head et al. (2006a) to be of glacial origin. Particularly significant were: (1) the turning flow observed in the lower valley emanating from the alcoves and merging with the general down-valley flow (Fig. 6); (2) the deflection of flow around obstacles (consistent with a relatively thin glacial ice deposit confined to the valley) (Fig. 7); thicker, more extensive ice sheets might follow the terrain less faithfully, as the obstacles would be overridden; and (3) the similarities in the thicknesses and extents of the modeled ice and the LVF deposits (Figs. 1b, c and 3).

We conclude that the glacial flow models described here successfully reproduce the lineated valley fill patterns and flow
scenarios interpreted by Head et al. (2006a) to be evidence for alpine-type valley glacial landsystems, and thus support this interpretation by providing a physically-reasonable scenario that produces results quantitatively similar to the geological observations. In this scenario, ice accumulates in alcoves and other protected areas, and debris cover derived from the adjacent exposed cliffs records the flow features of the underlying ice even after the ice beneath has partly sublimated away (Head et al., 2006a). A conclusion is that we would only expect to see such debris features where a source of surface debris was available, as is observed with the Mullins Glacier in the Dry Valleys of Antarctica (Marchant and Head, 2007; Marchant et al., 2010; Kowalewski et al., 2011; Shean and Marchant, 2010) and in the valley walls of the fretted terrain and along the dichotomy boundary. Otherwise, if not covered by debris, the ice at these latitudes would have sublimated away in the modern climate (e.g., Hauber et al., 2008) leaving no trace.

A corollary to this is that we would predict that any ice deposited at higher levels on the plateau above the valleys, with no higher scarps or nunataks from which debris could be deposited, would leave no record of ice flow. This raises the question: Could the currently observed LVF (Fig. 1b–e) represent not the greatest lateral extent of an alpine-type valley landsystem, but instead represent the waning stages and a remnant of a larger ice sheet preserved only when the scarps were exposed during retreat (Marchant and Head, 2006, 2007, 2008)? In the following section we examine the specific accumulation areas predicted by the LMD/GCM to assess whether the LVF might be the record of the waning stages of glaciation, after the regional ice surface in the valleys had dropped below the levels of the surrounding scarps, allowing debris to accumulate on the surface of the glaciers.

3. Glacial flow modeling of GCM-defined accumulation areas

3.1. The Mars GCM of the Laboratoire de Météorologie Dynamique

In order to specify the spatial distribution of the SMB of accumulated ice for a glacial flow model, one can arbitrarily choose values, and explore the consequences for specific values and predictions as we have successfully done above. An alternative approach is to use the results of a GCM that was chosen on the basis of its ability to reproduce the broad conditions that are observed geologically, as we did in the Tharsis region (e.g., Head and Marchant, 2003; Forget et al., 2006; Fastook et al., 2004, 2005, 2008a). The GCM used is the LMD/GCM (Laboratoire de Météorologie Dynamique, Forget et al., 1999); this GCM is able to reproduce the present-day water cycle with good accuracy (Montmessin et al., 2004). Extensive exploration of the parameter space found that necessary conditions for persistent ice deposition along the northern mid-latitude dichotomy boundary (Fig. 8a) included moderate obliquity (25–35°), high eccentricity (0.1) with perihelion at \( L_p = 270°\), high dust opacity (1.5–2.5), and a water source from sublimation of an ice sheet deposited on the flanks of the Tharsis volcanoes during a prior period of higher obliquity where tropical mountain glaciers are observed in the geological record (Head and Marchant, 2003; Shean et al., 2005, 2007; Milkovich et al., 2006; Kadish et al., 2008). The high dust content of the atmosphere was necessary to increase its water vapor holding capacity, thereby moving the saturation region to the northern mid-latitudes. Precipitation events are then controlled by topographic forcing of stationary planetary waves and transient weather systems, producing surface ice distribution and amounts that are consistent with the geological record. Both lower eccentricity and reversed
3.2. Specific global circulation model results for the northern mid-latitude boundary area

Work with the LMD/GCM (Forget et al., 1999; Montmessin et al., 2004; Levard et al., 2004; Hourdin et al., 1993) has provided a framework so that a map of potential accumulation rates for the dichotomy boundary region can be produced (Madeleine et al., 2009), which can then be used in an ice sheet model to describe a possible ice sheet with associated valley glaciation observed in the glacial geology (Head et al., 2006a; Fastook et al., 2008b). The distribution of positive SMB regions for the Madeleine et al. (2009) climate simulation showing best conditions for development of the mid-latitude glaciation are illustrated in Fig. 8a. The valley glaciation region described and tested above (Head et al., 2006a) and further discussed here lies in the area designated by the arrow in Fig. 8a (see also Fig. 1a, box 5). With peak values of ice accumulation reaching 16 mm/a, this pattern of SMB is clearly capable of producing a large ice sheet along the dichotomy boundary, but will its behavior agree with the geological observations (Fig. 1)?

3.3. Ice sheet modeling results

Clearly such a wide area of positive SMB will create a broad, extensive ice sheet (Marchant and Head, 2008), as opposed to only the localized valley glaciers observed in the geological record (Fig. 1a; Head et al., 2006a,b, 2010; Levy et al., 2007; Kress and Head, 2009; Morgan et al., 2009; Dickson et al., 2008, 2010; Hauber et al., 2008; Baker et al., 2010). We chose to focus attention again on the Deuterinolus–Protonolus Mensae (DPM) valley system described in Head et al. (2006a) (Fig. 1b–e). Running the ice sheet model at a resolution sufficient to resolve the valley, however, is prohibitive for such a broad area. We utilize instead the embedded-grid feature of UMISM. This feature allows us to run a broad-domain, low-resolution grid with a more limited-domain, higher-resolution grid embedded within it. This embedded grid obtains boundary condition information from a spatial and temporal interpolation of the low-resolution grid. This embedded feature allows us to nest grids, so that the jump in resolution is not so extreme as to produce spurious results.

The grids in the nest used in this model, with topography from MOLA, are also shown in Fig. 8. Starting with Fig. 8b, the broad grid outlined by the box in Fig. 8a (with 10,164 nodes and a resolution of 50 km), the nest progresses to Fig. 8c (with 16,625 nodes and 12 km resolution), onto Fig. 8d (with 7521 nodes and 6 km resolution), and finally to Fig. 8e, with the highest resolution (1.6 km, and 12,769 nodes). The three outer grids in the nest at 600 Ka are shown in Fig. 9, the time at which growth is stopped and retreat begins. The three vertical columns contain surface elevation, ice thickness, and ice velocity, respectively. The three rows correspond to the three outermost grids (Fig. 8b–d).

On the basis of this glacial accumulation and flow model, it is observed that the broad pattern of positive SMB in the northern plains (Fig. 8a) builds an ice sheet up to 4 km thick with a volume close to 9 million km³ in 600 Ka (Fig. 9, first row, left and middle

Fig. 7. A region further north near the dichotomy boundary (area G in Fig. 1b, THEMIS VIS V11208010 with model thickness and velocity at 1000 and 1500 Ka) far from the alcoves and in the central part of the valley where ice flow is down-valley. LVF bifurcates (bottom right) and flows around a massif forming a broad up-flow collar and a diffuse, down-flow "wake". LVF in the narrow pass between massifs is compressed. The lobate LVF extends into the northern lowlands at top left. Note the arcuate, piedmont-like configuration of the lobe and the high density of pits. Model results show the flow clearly narrowing between the two bedrock highs and increasing the velocity to values in excess of 200 mm/a.
This is considerably more than our own estimate of a Tharsis-region volume of 0.53 million km$^3$ (Fastook et al., 2008a,b), but the Tharsis-region estimate was very conservative and was constrained to be no greater than the clearly observable glacial deposits mapped in that area. Other estimates of the Tharsis volumes are considerably higher with much more extensive ice sheets. One reconstruction by Kite and Hindmarsh (2007) is 25–54 million km$^3$, a size hard to reconcile with the current volume of the North Polar Layered Deposits, which is 1.2–1.7 million km$^3$ (Zuber et al., 1998). If we restrict our ice sheet to the next grid in the nest (Figs. 8c and 9, second row) our volume is close to 2 million km$^3$, still larger than our estimate for Tharsis, but closer to the volume of the current North Polar Layered Deposits. We do in fact take depletion of the source into account, as that is the reason we turn off the accumulation portion of the SMB at 600 Ka when retreat and collapse of the ice sheet begins, leading to the patterns of ice thickness shown in the valley complex in Fig. 12. Clearly, the actual sources, volumes and transport pathways of water ice are not fully understood and subject to current investigation (e.g., Mischna et al., 2003; Levrard et al., 2004, 2007). Nonetheless, the broad modeled ice sheet covers much of the adjacent plateau and is particularly thick near the dichotomy boundary (Fig. 9, middle column); fretted valleys, particularly in areas interpreted to be the sites of local valley glacial

Fig. 8. (a) Mass balance in mm/yr from Madeleine et al. (2009) showing persistent ice deposition along the northern mid-latitude dichotomy boundary. GCM parameters included moderate obliquity (25–35°), high eccentricity (0.1) with perihelion at Lp = 270°, high dust opacity (1.5–2.5), and a water source from sublimation of an ice sheet deposited on the flanks of the Tharsis volcanoes; (b) the outermost of the nested grids with a resolution of 50 km, outlined by the box in (a); (c) with 12 km resolution, outlined in (b); (d) with 6 km resolution, outlined in (c); and (e) with the highest resolution of 1.6 km, outlined in (d).
landsystem deposits (e.g., Fig. 1a; Morgan et al., 2009; Head et al., 2006b; Baker et al., 2010; Dickson et al., 2008; Levy et al., 2007), show thick ice accumulations.

The specific DPM valley (Head et al., 2006a; Fig. 1b–e) lies on a GCM grid point where high accumulation rates are predicted (the resolution of the GCM is $5.625^\circ$ longitude by $3.75^\circ$ latitude). This results in a “peninsula” of ice that stretches into the highlands of the dichotomy boundary with thicknesses approaching 3 km. Note that in the logarithmic velocity scale, a value of 1 corresponds to a velocity of 10 mm/a. The DPM valley (Fig. 1b–e) lies in a saddle region where ice flow is clearly faster, ~100 mm/a, and is channelized in the trunk of the valley.

This channelization is more evident in the highest-resolution grid (Fig. 8e). In Fig. 10 the evolution during growth of the valley ice complex (top to bottom: surface, thickness, and velocity) is shown at 100 Ka intervals (left to right: 100–600 Ka). In the surface

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**Fig. 9.** Surface elevation, thickness, and velocity (columns) for the three outermost grids (rows) of Fig. 8b–d in the DPM lineated valley fill system.
figures (Fig. 10, top row), we see a gradual progressive drowning of the valley topography. Thickness (Fig. 10, middle row) begins with a uniform mantling, but quickly evolves into thicker ice in the valleys (4 km) but much thinner over the plateaus (less than 3 km) and aretes (~1 km). Recalling that ice flow directions are down the surface topographic gradient, perpendicular to surface elevation contours, we see the organization of a coherent flow pattern with maximum velocity (Fig. 10, lower row) in the valleys (~100 mm/a) but one, two, and even three orders of magnitude less over the more thinly covered aretes and plateaus between the valley trunks. Thus, to a first order, the general distribution of ice accumulation, as predicted by the LMD/GCM, readily reproduces regional plateau glaciation, and shows that glacial flow is focused into the fretted terrain valleys due to the effect of the pre-existing topography.

What is the fate of the regional ice sheet during its sublimation and retreat? As was mentioned, with the relatively high accumulation rates predicted by Madeleine et al. (2009), growth of a significant ice sheet is relatively rapid, and we grew the ice sheet under steady climate conditions for 600 Ka. At this point, we removed the positive accumulation component of the SMB, leaving only the negative sublimation component. This might be expected to occur,

Fig. 10. Growth to 600 Ka, showing surface (m), thickness (m), and velocity (log 10(mm/a)) at 100, 200, 300, 400, 500, and 600 Ka in the DPM lineated valley fill system.

Fig. 11. Ice evolution versus time for a typical point in Deuteronilus (41.25N,28.125E). By tracking increasing and decreasing thickness, the accumulation and ablation components of the mass balance can be separated.
begins at zero and climbs to close to 3 kg/m² (a period of positive accumulation) and negative (ablation) portions of the SMB correspond exactly to the values reported by Madeleine et al. (2009) in Fig. 8a, but we have a measure of potential negative mass balance (a point where the final endpoint of ice amount was negative) and we have the possibility to modify either the positive part (turn off the snowfall) or the negative part (increase or decrease sublimation) while keeping the rest of the climate true to the GCM results (the spatial distribution of mass balance and surface temperatures).

We now explore the retreat of the regional ice sheet. In this model we do not include any effects of debris cover that might be produced as the valley and dichotomy boundary scarps become exposed and could potentially provide a rock and debris cover to armor the ice and protect it from sublimation. Instead, the retreating ice is assumed to be pure ice. The configuration of the DPM valley at 600 Ka, the point at which retreat is allowed to begin by removing the positive portion of the mass balance, is shown in the right-most column of Fig. 10. The evolution of the ice sheet as this retreat progresses is shown in Fig. 12.

(1) By 800 Ka (Fig. 12, first column), 200 Ka after retreat begins, the surface is dropping and the underlying topography is becoming more evident as the first aretes emerge as nunataks. Velocities are considerably reduced to 1 mm/au in the valleys and 0.001 mm/a on the plateaus, although still clearly channelized in the valley proper.
In the second glacial flow model simulation, the regional accumulation patterns predicted by the LMD/GCM were used (Madeleine et al., 2009). This glacial flow model simulation produced a much wider region of thick ice accumulation, and significant glaciation on the plateaus and in the regional plains surrounding the dichotomy boundary (Figs. 1a and 8a). Deglaciation produced decreasing ice thicknesses, with flow centered on the fretted valleys. As plateaus lost ice, and scarps and cliffs of the valley and dichotomy boundary walls were exposed, there was considerable potential for the production of a rock debris-cover that could preserve the underlying ice and the surface flow patterns seen today. In this model, the lineated valley fill and lobate debris aprons were the product of final retreat and downwasting of a much more regional glacial landsystem (Marchant and Head, 2008), rather than the maximum extent of an alpine valley glacial landsystem. From these glacial flow models, information can be derived about the possible timing of the growth of these systems. In the alcove-only case, several millions of years are required for the flow system to achieve a pattern consistent with the geological record. Because the alcove-only case uses specified areal distribution of ice deposition we cannot constrain its occurrence to any specific time interval.

Coupling the ice sheet model with a GCM yields a pattern of ice sheet development, which during its collapse also yields a configuration consistent with the geological observations. Because of the high accumulation rates predicted by the LMD/GCM, the ice sheet grows much more rapidly than the alcove-only case. As the ice sheet collapses, presumably due to the exhaustion of the moisture source in the Tharsis region or spin-axis/orbital changes, a configuration emerges of a limited valley glacier system compatible with the geological observations. Importantly, as the surface lowers to the point where armoring debris would be available from the cliff scarps, the configuration resembles the alcove-only case. At this point, if debris is supplied from the adjacent newly exposed rock cliffs, we would expect the sublimation rates over the debris-covered valley-floor glaciers to be reduced by orders of magnitude (Kowalewski et al., 2006) potentially preserving ice even now in the valley trunks. This will also be the case with isolated massifs at the mouths of the valleys, leading to the possibility that the lobate debris aprons are ice-cored remnants of a much more extensive ice sheet that covered the area during climatically appropriate periods. We are currently exploring these options by modeling the accumulation of debris once cliffs are exposed, and the evolution and flow of such debris-covered glaciers (Fastook et al., 2010).

4. Conclusions

A fretted valley system located at the northern mid-latitude dichotomy boundary and containing lineated valley fill with extensive flow-like features interpreted to be glacial in origin (Head et al., 2006a) has been modeled using glacial flow models linked to atmospheric general circulation models. The glacial flow model results are consistent with the geological features observed and with the interpretation of these as alpine valley glacial landsystems.

In the first glacial flow model simulation, sources were modeled in the alcoves only and were found to be consistent with the alpine valley glaciation interpretation for various environments of flow in the system. These results indicated that the currently observed LVF deposits could be the result of initial ice accumulation in the alcoves, accompanied by debris cover that led to advancing alpine glacial landsystems to the extent observed today, with preservation of their flow texture and the underlying ice during downwasting in the waning stages of glaciation.

In summary, coupling the regional accumulation rates of the LMD/GCM (Madeleine et al., 2009) with a glacial flow model shows that the resulting regional plateau and plains glaciation that would have characterized the broad area during the period of glaciation is quite capable of producing results during its collapse that are consistent with the alpine valley glaciation interpretation of the LVF. Mapping of the patterns of retreat during the waning stages of the glaciation shows that the valley walls are eventually exposed and could readily serve as sources of debris to produce the lineated valley fill and related flow patterns seen in numerous places along the dichotomy boundary (Fig. 1a–e). Furthermore, the final configuration of the ice sheet upon its retreat is that of glacial ice surrounding massifs in the plains just north of the dichotomy boundary, a pattern very similar to the patterns for lobate debris aprons reported by Ostrach and Head (2008), Ostrach et al. (2008), Holt et al. (2008a,b), Plaut et al. (2008), Safaeinili et al. (2009) and Kress and Head (2009). These relationships predict that the lobate debris aprons (LDA) might slightly postdate the lineated valley fill (LVF) in their formation. These remnant patches occur along the base of steep topographic scarps because the ice is thickest where it flows from the high to lower terrain across a step in bed elevation. While not modeled in this experiment, the potential for debris armoring beneath scarps would increase the likelihood of LDA preservation.

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