Volcano–ice interactions in the Arsia Mons tropical mountain glacier deposits

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Fan-shaped deposits (FSD) superposed on the sides of the Tharsis Montes volcanic edifices are widely interpreted to have been formed by cold-based glaciation during the Late Amazonian, a period when the Tharsis Montes were volcanically active. We survey the ~166,000 km² Arsia Mons FSD using new, high-resolution image and topography data and describe numerous landforms indicative of volcano–ice interactions. These include (1) steep-sided mounds, morphologically similar to terrestrial tindar that form by subglacial eruptions under low confining pressure; (2) steep-sided, leveed flow-like landforms with depressed centers, interpreted to be subglacial lava flows with chilled margins; (3) digitate flows that we interpret as having resulted from lava flow interaction with glacial ice at the upslope margin of the glacier; (4) a plateau with the steep sides and smooth capping flow of a basaltic tuya, a class of feature formed when subglacial eruptions persist long enough to melt through the overlying ice; and (5) low, areally extensive mounds that we interpret as effusions of pillow lava, formed by subglacial eruptions under high confining pressure. Together, these eruptions involved hundreds of cubic kilometers of subglacially erupted lava; thermodynamic relationships indicate that this amount of lava would have produced a similar volume of subglacial liquid meltwater, some of which carved fluvial features in the FSD. Landforms in the FSD also suggest that glaciovolcanic heat transfer induced local wet-based flow in some parts of the glacier. Glaciovolcanic environments are important microbial habitats on Earth, and the evidence for widespread liquid water in the Amazonian-aged Arsia Mons FSD makes it one of the most recent potentially habitable environments on Mars. Such environments could have provided refugia for any life that developed on Mars and survived on its surface until the Amazonian.

1. Introduction

Volcanism is known to be a fundamental process in shaping the surface of Mars throughout its history, including the early volcanic construction of Tharsis, the widespread late Noachian and Early Hesperian ridged plains, and Amazonian volcanism of the Tharsis and Elysium region (e.g., Tanaka et al., 1991; Head, 2007; Werner, 2009; Carr and Head, 2010a). Historically, water ice on Mars has been thought to reside primarily in the regolith (e.g., Mellon and Jakosky, 1993; Mellon et al., 2004), upper crust cryosphere (e.g., Clifford, 1993), and thick polar ice deposits (e.g., Thomas et al., 1992; for extensive reviews see Carr, 1996; Byrne, 2009; Carr and Head, 2010b). In the mid-1990s, increasing numbers of workers began to appreciate the fact that past spin axis–orbital parameter variations could cause polar ice to be mobilized, transported and deposited equatorward (as pointed out by Jakosky and Carr, 1985), raising the possibility that volcanic eruptions might interact with deposits of ice, not just in the subsurface (Chapman and Tanaka, 2002), but also in glacial (Head and Wilson, 2002) and frozen lake environments (Chapman, 1994; Chapman and Tanaka, 2001). The development of analytical solutions for the possible histories of spin axis/orbital parameters on Mars (Laskar et al., 2004) paved the way for an increased understanding of locations where non-polar ice might reside under different past conditions (e.g., Head et al., 2003; Head and Marchant, 2003; Shean et al., 2005; Baker et al., 2010) and the climatic and atmospheric settings in which they might have been deposited (e.g., Forget et al., 2006; Madeleine et al., 2009, 2013). Concurrently, interest in astrobiology has led to a search for specific environments that might have been conducive to initiation and sustenance of life (The MEPAG Special Regions–Science Analysis Group, 2006).
including microbial habitats related to volcano–ice interactions (Cousins and Crawford, 2011).

New high-resolution image, spectral, geochemical and altimetry data collected by recent spacecraft missions have provided the ability to analyze the interaction of volcanic intrusions and eruptions with widespread paleo-ice deposits in unprecedented detail. In this contribution, we analyze the Arsia Mons fan-shaped deposit (FSD), a ~166,000 km² deposit on the northwest flank of the Arsia Mons shield volcano interpreted to have formed in the Late Amazonian as a tropical mountain glacier (Head and Marchant, 2003; Forget et al., 2006; Shean et al., 2005, 2007; Fastook et al., 2008). We review the properties of the FSD and the nature and history of Tharsis volcanism, outline the basic principles of volcano–ice interactions, and examine specific examples of volcano–ice interactions in the Arsia Mons deposit. We find abundant examples of glaciovolcanic landforms in the Arsia Mons deposits.

2. Volcanism, glaciation and volcano–ice interactions in the Tharsis region

2.1. Volcanism in the Tharsis region

Volcanic activity has been key in forming the Tharsis rise (e.g. Phillips et al., 2001) as well as the individual major volcanic edifices that characterize its surface: Olympus Mons, Alba Patera, and the Tharsis Montes, Ascaraeus, Pavonis and Arsia Mons. For Arsia Mons in particular, associated lava flows date from Middle Hesperian to Late Amazonian in age, with major episodes of volcanism ending at ~100–200 Ma, ~500 Ma, ~800 Ma and 2 Ga and major shield construction ending ca. 3.54 Ga (Werner, 2009). Based on crater counts and stratigraphic relationships, Crumpler and Aubele (1978) suggested that volcanism at Arsia Mons followed a sequence in which (1) thick lava plains accumulated throughout the Tharsis region, followed by (2) construction of the main shield by short multiple flow units similar to those that build terrestrial shields; (3) the onset of intense eruptive activity forming “parasitic shields” along a southwest to northeast trending linear rift or fissure; (4) formation of concentric graben, concentric fractures, and the caldera, possibly contemporaneously with (3); and (5) continuing eruptions on the northeast and southwest flanks. Bleacher et al. (2007) noted evolution towards shorter flow lengths and from tube-fed flows to channel-fed flows over the history of each of the Tharsis Montes; terrestrial basaltic tube-fed flows are associated with lower-viscosity lavas, lower effusion rates and longer-term, more stable eruptions than their channel-fed counterparts.

Increasing silica content decreases the capacity of lava to melt ice and increases the aspect ratio of glaciovolcanic landforms (Hickson, 2000; Smellie, 2007; McGuire, 2009). The composition of Tharsis lavas has been difficult to discern, however, because the thick mantle of dust covering the region (Ruff and Christensen, 2002) prevents definitive spectroscopic measurements of mineralogy. Linear deconvolution of TES spectra from relatively dust-free flows near Arsia, predating rift apron formation, indicated that the flows have a basaltic composition similar to martian shergottite meteorites (Lang et al., 2009). A basaltic composition can also be inferred for some Tharsis flows based on the distance they traveled from their source vents (Scott and Zimbelman, 1995; Hiesinger et al., 2007).

2.2. Glaciation in the Tharsis region

The FSDs on the west and northwest flanks of the Tharsis Montes consist primarily of three units (Zimbelman and Edgett, 1992; Scott and Zimbelman, 1995): a ridged facies, a knobby facies, and a smooth facies that is uniformly stratigraphically higher than the other two. Previously proposed hypotheses have postulated a wide variety of candidate origins for the FSDs, including mass-wasting processes, pyroclastic flows or other volcanic processes, tectonic deformation, or glaciation (see review in Shean et al., 2005). Williams (1978) was the first to propose that the deposits might be glacial. Lucchitta (1981) noted the similarities between ridges in the Olympus and Arsia FSDs and recessional glacial ridges in Alaska and Iceland, and suggested that the even superimposition of some martian ridges upon high topographic obstacles implied the removal of a large quantity of material, most easily explained by the sublimation of ice.

Head and Marchant (2003) further refined this explanation of the ridged facies by drawing an analogy to the ridge-forming drop moraines left by cold-based glaciers in the Mars-like Antarctic Dry Valleys. Based on its superposition relationships, its association with the ridged facies, and its lack of fluvial features, they argued that the knobby facies represents sublimation till; based on its morphology and spatial associations, they argued that the smooth facies consists of remnant debris-covered glaciers. Shean et al. (2005), using data from the Thermal Emission Imaging System (THEMIS), the Mars Orbital Laser Altimeter (MOLA), and the Mars Orbiter Camera (MOC) in addition to Viking imagery, mapped glacial and glaciovolcanic features in the Pavonis Mons FSD, including tindar, outflow channels, and lava flows with steep, ice-chilled margins. Shean et al. (2005) presented counterarguments to each of the problems with the glacial hypothesis for Tharsis FSD formation raised by Edgett (1989); Kadish et al. (2008) mapped a similar array of features in the Ascaraeus Mons FSD, as well as a prominent plateau resembling terrestrial subglacially erupted tuyas. Shean et al. (2007) identified a younger (~65 Ma), smaller set of cold-based glacial deposits, including remnant debris-covered glaciers, superimposed on the Arsia Mons fan-shaped deposit.

Climate reconstructions and results from ice-flow modeling have lent additional support to the geologic arguments that the Tharsis FSDs are glacial in origin. Simplified ice stability models have predicted for decades (Jakosky and Carr, 1985; Jakosky et al., 1995) that water vapor would be transported from the poles of Mars to lower latitudes during periods of high spin-axis obliquity, though these models were unable to predict the resulting ice distribution in detail. More recently, three-dimensional general circulation models (GCMs) that include water cycle physics have consistently simulated the process (Haberle et al., 2000; Mischena et al., 2003; Levrard et al., 2004). The GCM experiments of Forget et al. (2006) were the first with sufficient resolution to allow direct comparison with geological features. In their simulations at 40° and higher obliquity, sublimation of polar ice increased the atmospheric water content relative to the current martian climate. As this moist air ascended the Tharsis Montes and Olympus Mons in the prevailing westerly to northwesterly winds, precipitation occurred on the windward flanks of the volcanoes. These results further supported the interpretation of the Tharsis FSDs as tropical mountain glaciers (Head and Marchant, 2003).

In the Forget et al. (2006) GCM simulations, ice precipitated at the upslope region of the deposit, where the accumulation zone of the hypothesized glacier would be expected. Fastook et al. (2008) used climate fields from these simulations as input in model runs using the University of Maine Ice Sheet Model (UMISM). By parameterizing mass balance in terms of local temperature and restricting the orientation of the long axis of glacial flow, the Arsia Mons ice sheet reproduced in model runs matched the footprint of the mapped Arsia Mons FSD with an accuracy comparable to that of terrestrial models in simulating modern ice sheets (Fastook et al., 2008). In summary, the current consensus
favors the glacial hypothesis for the origin of the Tharsis FSDs (Williams, 1978; Lucchitta, 1981; Head and Marchant, 2003). We thus proceed to investigate the interaction of volcanism and glaciation in the Tharsis region, with particular emphasis on the Arsia Mons fan-shaped deposit.

2.3. Principles of volcano–ice interactions

The interaction of ascending magma and glacial ice involves significant heat transfer and melting, and thus the process can have a profound influence on the nature and fate of the magma, the properties of the glacial ice, the resulting volumes and fate of meltwater produced, and the resulting landforms, both glacial and volcanic. Several terrestrial environments (e.g., Iceland, the Antarctic Peninsula, and western Canada) have proven to be natural laboratories for the study of these interactions and resulting landforms and for the development of the basic physical principles of heat transfer and melting (e.g., Smellie and Skilling, 1994; Gudmundsson, 1997; Hickson, 2000). We use the series of principles and physical processes related to magma–ice interactions in terrestrial glaciers developed by Höskuldsson and Sparks (1997), Wilson and Head (2002, 2007a, 2009), Gudmundsson (2003), Head and Wilson (2007), Tuffen (2007), and Wilson et al. (2013), and applied to Mars by Head and Wilson (2002, 2007), to assess the relationships in the Arsia Mons fan-shaped deposit.

2.4. Volcano–ice interactions and microbial habitats

Interaction between glaciers and lava flows on Earth creates environments that can function as microbial habitats, including subglacial lakes, subglacial basaltic edifices, permafrost hydrothermal systems, and glacial springs (Cousins and Crawford, 2011). Volcano–ice interactions can also induce locally wet-based conditions in cold-based glaciers (e.g., Fahnestock et al., 2001; Fox Maule et al., 2005; Head and Wilson, 2007; Hambrey et al., 2008), which introduce new habitable environments including wet subglacial sediments. Wet-based glacial ice also contains more water-filled fractures above the bed (Smellie, 2006), which can serve both as habitats and as pathways for any potential life to better distribute itself throughout a glacier (Fountain et al., 2005). In this work, we document evidence for many episodes of volcano–ice interaction in the Arsia Mons FSD; Scanlon et al. (in preparation, 2014) document the effects of this heat transfer on the style of glaciation as recorded in the deposit, and the potential for habitable environments in the Late Amazonian Arsia Mons FSD glacier.

2.5. Data and methods

A photographic basemap covering Arsia Mons and the Arsia fan-shaped deposit was constructed from Mars Reconnaissance Orbiter (MRO) Context Camera (CTX) images (Malin et al., 2007) with ~5 m per pixel resolution using ArcMap 10.0. This basemap was augmented with images from the High Resolution Stereo Camera (HRSC) with 10–30 m per pixel resolution (Neukum and Jaumann, 2004). Topographic data from MOLA (Zuber et al., 1992; Smith et al., 1999) at 128 pixel per degree (~463 m per pixel) resolution and, where available, HRSC-derived Digital Elevation Maps (DEMs) with ~100 m per pixel resolution (Dumke et al., 2008), were superimposed on the CTX basemap. These data were used with the Spatial Analyst toolkit in ArcMap 10.0 to create slope maps, contour maps, elevation profiles, and other derived products. The Arsia FSD and its surroundings (Fig. 1) were examined and mapped in detail.

3. Candidate glacial and glaciovolcanic features of the Arsia Mons fan-shaped deposit

Zimbelman and Edgett (1992), Scott and Zimbelman (1995), and Head and Marchant (2003) describe the geological setting and the main units of the Arsia Mons FSD. Shean et al. (2007) present a study of recent glaciation on Arsia Mons. Here we catalog glacial and glaciovolcanic features of the Arsia Mons FSD (Fig. 1), including those previously noted by Wilson and Head (2007b, c) but not described in detail (Fig. 2), and present a comprehensive examination of candidate glaciovolcanic features.

3.1. Northwest plateau

Toward the northwest edge of the deposit (Fig. 1b, box 3.1) lies a plateau ~17 km long by ~15 km wide and ~140 m high (Fig. 2a and b). The plateau is steep-sided compared with nearby subaerial lava flows, having typical slopes between 6° and 12° (Fig. 2c). An elongate, southwest–northeast oriented mound, ~10 km long and ~3 km wide, stands at the center of the plateau. The mound slopes steeply (angles up to 15°) to the northwest and slightly more gently to the southeast, and stands ~150 m (up to 200 m) above the rest of the plateau (Fig. 3). At its northern edge, the plateau conforms closely to and becomes indistinguishable from the rim of a heavily degraded crater ~3.5 km in diameter. The circularity and rim structure of the crater distinguish it from volcanic pit craters found elsewhere in the deposit, and we interpret it as an impact crater predating the formation of the plateau. The plateau lies along the trend of a line of subaerial spatter cones beginning ~45 km to the north (Fig. 4) and Wilson, 2007), suggesting a possible genetic relationship along the same dike-emplacment strike.

Superposed on the plateau are two sinuous ridges (Fig. 5a and b), originating near the base of the mound. The southern ridge is 4–6 m tall and extends at least 25 km downslope; the northern ridge is up to 20 m tall but becomes difficult to trace ~5 km beyond the edge of the plateau. Both ridges have sharply defined flat crests atop the plateau (Fig. 5c) and take on a more diffuse appearance downslope. Downslope (i.e., to the northwest) of the plateau, unusually tall and thick moraines are bowed outwards relative to the surrounding drop moraines (Fig. 4a and b). Knobs across a region downslope of the plateau appear aligned or elongated in the downslope direction (Fig. 4).

More than a dozen channels emerge from beneath the prominent moraine downslope of the plateau (Fig. 4a and b); ~6 km further downslope the channels coalesce into a braided channel which is traceable for ~35 km (Fig. 6a and b). The channels emerge from beneath a layer of fine sediment at the base of the moraine. Where HRSC DEM coverage is available, we determined that the main channel is ~35 m deep and has a V-shaped profile (Fig. 7a); the tributaries are shallower (Fig. 7b).

Calling on the series of principles and physical processes developed by Wilson and Head (2002, 2007a, 2009), Head and Wilson (2007), Tuffen (2007), and Wilson et al. (2013), and applied to Mars by Head and Wilson (2002, 2007; Fig. 8), we interpret this collection of features to be related to the interaction of one or more dike emplacement events with the Arsia tropical mountain glacier, and the aftermath of the contact of hot intruding magma with the overlying glacial ice.

When the confining pressure of ice above a subglacial eruption is sufficient to suppress explosive degassing, the lava emerges as pillows (e.g., Wilson and Head, 2002; Tuffen, 2007). The pillows form low mounds, ridges, or sheets that slope gently when emplaced into water, but can become much steeper if they chill against ice walls (Smellie, 2009). Meltwater produced by the eruption is confined
Fig. 1. (a) Geomorphological map of the Arsia Mons fan-shaped deposit (FSD), after Zimbelman and Edgett (1992) and Scott and Zimbelman (1995). Red lines denote large volcanic graben, white lines denote contacts between units, and black lines denote the outlines of glaciovolcanic landforms and distorted moraines. Dashed lines denote inferred contacts. Blue and white dots in the smooth facies denote pits and knobs, respectively. THEMIS 100 m/pixel daytime image mosaic. (b) Numbered white boxes indicate the regions discussed in Sections 3.1–3.7 of this paper. All map figures in this paper are oriented north upwards. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
to a cavity or cavities in the glacier, above the flow (Fig. 8a). If the pressure in the melt cavity is reduced sufficiently by the lava flow approaching the edge of the glacier, meltwater draining from under the glacier, or the growing edifice thinning the ice above it by melting, the eruption can become explosive (Wilson and Head, 2002; Tuffen, 2007; Wilson et al., 2013). Explosive eruptions melt ice much more efficiently than pillow lava effusion because the lava fragmentation creates a much larger surface area per unit volume of lava (Head and Wilson, 2002; Tuffen, 2007), and the melt cavity becomes much larger immediately above the vent, growing outward as the eruption continues (Fig. 8b).

On the basis of morphology, morphometry, and landform associations (Figs. 5–7), and physical volcanology elucidated by terrestrial subglacial eruptions (Fig. 8), we interpret the history of the Northwest plateau complex as follows (Fig. 9):

1. Lava erupted under thick ice cover and spread outward and downslope as pillows, embaying a nearby, pre-glacial, impact crater (Fig. 9a). This flat, steep deposit formed the majority of the Northwest plateau. While it is larger than any known terrestrial examples, the main body of the Northwest plateau lies on the area-to-volume trend line of terrestrial pillow mounds and
sheet-like sequences (Smellie, 2009, Fig. 6). The water melted by the pillow lava effusion remained confined in a lens above the flow (Fig. 9a).

2. Elevated geothermal heat in the vicinity of the eruption, or lifting of the glacier by the growing melt body (Smellie, 2000), may have caused the glacier to decouple from the ground (Fig. 9b). This layer of water would have been in contact with the melt cavity, at glaciostatic pressure. Due to the decoupling, the glacier transitioned from regionally cold-based to locally wet-based and was then able to slide along its bed, sculpting knobs downslope of the plateau into elongated forms (Fig. 9a and b). This caused a lobe to extend farther downslope than the surrounding (cold-based) glacier, as evidenced by the downslope bulge of the glacial drop moraines in the vicinity of the plateau: while the moraines are a recessional feature that would not have been deposited until after the eruption ceased, the local deformation of the ice in this region would have changed the local outline of the glacier, and hence the shape of the drop moraines left behind as it receded.

3. Once the layer of liquid water reached the edge of the glacier, the pressure difference between the subglacial environment and the atmospheric pressure outside the glacier caused the water to flow out catastrophically in a jökulhlaup (Fig. 9c); a transition from subglacial to subaerial flow is indicated by (a) the braided channel morphology outside the moraine we interpret as marking the syneruptive glacial margin and (b) the many tributary heads emerging from this moraine. This outflow carved the channels downslope (Figs. 5–7).

4. The drainage of the melt cavity suddenly reduced the water pressure (and hence the pressure above the vent) from glaciostatic to atmospheric; the eruption was no longer constrained to a pillow effusion. Since volatiles were then free to exsolve and water could still enter the vent, the eruption entered an explosive phase (Fig. 9d), melting a taller cavity into the ice above the vent (Wilson and Head, 2002; Wilson et al., 2013). The tephra from this eruption, confined in the new cavity, formed the mound over the center of the plateau.

5. Melt continued to escape through the jökulhlaup channels until the collapse or deformation of the ice under its own weight closed the thin cavity over the pillow sheet. Due to the creation of the new melt cavity, and the downward motion of the ice surface above it caused by density differences between ice and water (Smellie, 2000), the ice above the vent thinned considerably and the eruption could proceed as explosive until the magma supply was exhausted.

6. As the edifice cooled, some of the remaining meltwater eroded channels into the ice at the base of the glacier (Fig. 9e). More and coarser hyaloclastite tephra grains fell out of suspension at the heads of the channels. The sediment in the melt channels formed the present-day eskers, which are tall and sharp-crested close to the vent, but low and broad-crested downslope (e.g. Shreve, 1985). This transition from sharp- to broad-crested eskers away from the plateau therefore supports the hypothesis that meltwater in the channels was generated by heat from the volcanic edifice.

3.2. Elongate plateau

Centered ~45 km north of the Northwest plateau (Fig. 1b, box 3.2) lies an elongate plateau ~39 km long, with a maximum width of ~18 km (Fig. 10a and b). The plateau trends approximately
north–south and has a height of up to ~140 m. Irregular lobes extend downslope from its center and a gently sloping lobe extends downslope from its northern terminus (1–6°, compared with 12° or steeper slopes on other edges of the plateau; Figs. 10c and 11b). At its center lies an elongate, west–southwest to east–northeast trending mound which stands up to ~140 m higher than the rest of the plateau (Figs. 10a, b and 11a, c). Several overlapping, highly sinuous ridges extend ~5–20 km north and west from the central mound of the plateau. As with the Northwest plateau, many drop moraines near the plateau...
bow outwards relative to superposed and nearby drop moraines (Fig. 10a and b). The moraine that we interpret as representing the glacier boundary at the time of plateau formation, due to its enlarged and knobby appearance and stratigraphic relationships with the other local landforms, is also bent outwards.

A gently sinuous channel, sharply incised at the edge of the plateau and surrounded by smooth debris farther downslope, begins near the northern terminus of the plateau (Fig. 12). It is superposed by the large moraine that we interpret as having been deposited by the syneruptive glacial margin, and crosscuts several drop moraines farther downslope (Fig. 12). The contact between the flows emerging from the line of spatter cones west of the plateau and the debris associated with the channel is not visible at CTX resolution, but the edges of the flows are sharply defined whereas the boundaries of the debris are diffuse, and the surface texture of the flows is knobby compared with that of the channeled debris.

The long axis of the plateau is collinear with a curving graben to the south and a series of pits to the north (Fig. 4a and b). The pits cross-cut a ridged lava flow; this flow, in turn, is bifurcated by the outermost moraine of the FSD (Fig. 4b) and appears to embay the spatter cones to the north of the Northwest plateau. Since the elongate plateau, graben, and pits would have been emplaced in the same eruptive phase if they erupted from the same dike, and the Northwest plateau would have been emplaced contemporaneously with the spatter cones for the same reason, this indicates that the Northwest plateau was emplaced after the outermost drop
moraines (i.e., those between the bold-outlined “prominent moraines” and the very outermost moraine in Fig. 4b) but before the bifurcated ridged lava flow, whereas the elongate plateau was emplaced after the bifurcated ridged flow.

This plateau appears to have formed by a mechanism broadly similar to that of the Northwest plateau, with a few important differences. We interpret the history of the plateau and its surroundings as follows.

1. Before the emplacement of the elongate plateau, flows associated with the spatter cones to the west of the elongate plateau and north of the Northwest plateau interacted with remnant ice in the glacial moraines of the ridged facies. Steam explosions caused by this interaction imparted a chaotic texture to the flows. There is the possibility that peperite flows could be formed in this environment (e.g., Skilling et al., 2002), but orbital resolution is insufficient to document this. The flows sharply incised the terminal moraine of the FSD, and continued past the edge of the deposit.

2. Lava effused from a dike-related fissure below thick glacial ice cover, forming a steep-sided, elongate plateau enclosed by a thin melt cavity.

3. Since this vent is much closer to the syneruptive glacier margin than the vent at the Northwest plateau is, the lava flow near the edge of the glacier would still have been partially molten when the melt cavity reached the glacier edge, contacted the atmosphere, and rapidly depressurized. Partial fragmentation of the flow resulted (Wilson and Head, 2002; Wilson et al., 2013). Simultaneously, the melt cavity drained catastrophically. The meltwater entrained the chilled lava fragments and deposited sediment along a channel downslope of the plateau.

4. After the meltwater drained, the source fissure was no longer under glaciostatic pressure and, as at the Northwest plateau, began to erupt locally as a fire fountain, building a tephra mound and ridge, and eroding a taller melt cavity into the surrounding ice due to more efficient heat transfer in explosive eruptions.

5. After the eruption ceased, melt from the cavity over the cooling tephra mound and ridge drained through sinuous englacial tunnels, carrying some tephra and leaving eskers when the glacier ablated (Fig. 10a and b).

The lack of a well-defined contact between the channel deposits and the chaotic volcanic flows allows several alternate possibilities for the history of the region downslope of the elongate plateau (Fig. 12). First, what we have mapped as channel sedimentary deposits has been mapped by others as part of the volcanic flow associated with the cones (Garry et al., 2013). This, however, would require the flow to have breached several moraines while continuing in a narrow, upslope-directed flow, rather than spreading between moraines. The ability of such a flow to penetrate beneath the glacier surface, as indicated by the superposition relationship observed between the channel and the prominent moraine nearest the elongate plateau, rather than pooling and spreading laterally at the glacier edge, is also doubtful. Second, the channel and the flows may have formed almost simultaneously, allowing mixing and interaction between the fluvial streams and volcanic flows. This second interpretation is in conflict with the stratigraphic relationships between the line of cones and the pit craters along the strike of the elongate plateau (Fig. 4), unless (1) the apparent embayment relationship between the line of cones and the bifurcated flow is spurious or (2) there were significant delays between episodes of eruptive activity at the northern and southern ends of the two inferred dikes (e.g., Wilson and Head, 1988; Parfitt and Wilson, 1994), such that the cone-associated flows and elongate plateau were emplaced near-simultaneously but the cones farther north were emplaced before the pit craters north of the plateau. Our preferred explanation summarized above, i.e. that the flows associated with the cones were emplaced before the channel associated with the elongate plateau, is based on these considerations.

3.3. L-shaped ridge

The largest steep-sided flow within the Arsia FSD (Fig. 1a) is a broad, flat-topped ridge near the northern edge of the deposit (Fig. 1b, box 3.3). The ridge (Fig. 13), which is primarily L-shaped in plan view, is ~53 km long with a maximum width of ~4 km. Its tallest point lies more than 400 m above the surrounding terrain; with the exception of local highs at the bends in the ridge, its height decreases gradually downslope from this point (Figs. 13a and 14). The ridge is much steeper than the two plateaus at the western edge of the FSD, with some faces inclined at angles up to 38° (Fig. 13c). The distal branch is leveed (Figs. 13d and 14), and a small, flat mound extends from the terminus of the ridge. An approximately triangle-shaped lobe at the upslope end of the ridge (Fig. 13) has an area of ~160 km², with sides ~16 to ~28 km long, and stands ~150–200 m tall.

Similar elongate, cohesive, steep-sided ridges have been mapped in the fan-shaped deposits of Ascraeus Mons and Pavonis Mons (Zimbelman and Edgett, 1992; Scott et al., 1998), these have been interpreted to be englacial lava flows (Head and Wilson, 2002; Wilson and Head, 2002; Shean et al., 2005; Kadish et al., 2008). The Mazama Ridge extending from Mount Rainier (Fiske et al., 1963; Lanphere and Sisson, 1995; Lescinsky and Sisson, 1998; Lescinsky and Fink, 2000), while composed of more silicic lava than any known Tharsis flows, is also an instructive analog. Mazama Ridge is up to 450 m high and up to ~1 km wide. It
formed when lava flowed through a relatively thin corridor of glacial ice: the lava melted ice ahead of it, the melt was able to drain away, and the lava filled the void. Thicker ice at either side confined the flow, and a glassy, chilled layer formed at its sides and top while the interior remained molten and continued to flow downslope. When the flow was temporarily dammed, e.g. by thickness variations in the ice through which it traveled, the lava pooled and the ridge became locally thicker. Similar terrestrial features in basaltic eruptions have been described by Wilson and Head (2002).

On the basis of its steep sides, its increased thickness at bends, and the leveed appearance and thin lobe at the end, we interpret the L-shaped ridge as a subglacial lava flow. Due to the pinched-off morphology where the L-shaped ridge and triangular plateau meet, we propose that the ridge began as a breakout flow from the plateau and proceeded downslope, possibly following a basal tunnel eroded by melt from the eruption of the triangular lobe. The direction of least resistance may also have been set by a crevasse in the ice, a local minimum in the glacier thickness above the flow, or the slope of the underlying, pre-glacial topography. The edges of the flow, in contact with glacial ice, would have chilled much faster than the center.

The morphology of the ridge is consistent with a history in which, after proceeding downslope (approximately northward) for ∼15 km, the flow was temporarily stalled and pooled, resulting in a local maximum of ridge thickness. Eventually, a breakout flow emerged and continued approximately westward. After flowing west for ∼15 km, the flow stalled and pooled again, creating another tall and thick region of the ridge, then continued northwest along a sinuous path. As magma supply to the flow ceased, the still-molten lava in the interior of the ridge partially drained from the chilled exterior of the distal branch, and the ceiling of the empty tube would likely have collapsed under the weight of the glacier, leaving the channels now observed at the distal end of the ridge.

3.4. Hidden plateaus

Scott and Zimbelman (1995) noted the presence of two mountains buried in the knobby facies, as well as one in a region we have mapped as belonging to the Smooth Lower Western Flank unit (Fig. 1b, boxes 3.4). The MOLA and HRSC topographic data that has since become available revealed a third plateau in the knobby facies (Fig. 15b). Two of the buried mountains are similar in morphology to the Northwest plateau, with a small, steep mound elevated above a flat sheet (Fig. 15a and c). The third lacks a mound (Fig. 15b).

Any further distinguishing features are obscured by the superposed knobby facies. In light of the similarities between their
morphology, aspect ratio and size and those of the northwest and elongate plateaus, however, it is probable that the hidden plateaus were also emplaced subglacially, by a similar mechanism (Fig. 9). If so, they would have melted a proportional quantity of ice; the hidden plateaus represent at least 70 km$^3$ of additional subglacially emplaced lava. The presence of mounds on two of the plateaus suggests that these eruptions also reached an explosive phase; this could have been permitted by drainage of that melt at the edge of the glacier if the edge was close to the plateaus at the time of eruption, or if the eruption melted close enough to the glacier surface to significantly reduce the pressure above the vent. While no fluvial features are apparent near the hidden plateaus, channels the size of those at the northwest and elongate plateaus would easily be obscured by the superposed knobby facies, and any meltwater that remained confined by the glacier would have eventually refrozen in place without leaving any diagnostic landforms. A subaerial lava cap would be distinguishable in morphology and slope maps even through the knobby facies, if one had been emplaced (forming a tuya), but no such feature is apparent on either mound.

3.5. Volcano–ice interactions in the Lobate Facies

The Lobate Facies (Zimbelman and Edgett, 1992) is ∼7500 km$^2$ in area, located at the northeastern edge of the deposit (Fig. 1b, box 3.5), and consists largely of overlapping sinuous, leveed lava channels and chains of pit craters (Figs. 16 and 17). The unit is named for lobate, heavily channeled, steep-sided flow-like features that can be traced tens of kilometers from pit crater chains near the summit of the volcano (Fig. 16). These features have steep sides and stand 300 m or more above the surrounding terrain. Aside from being somewhat thinner, the lobate features at Arsia Mons are morphologically indistinguishable from similar features at Pavonis Mons and Ascreus Mons (Kadish et al., 2008; Shean et al., 2005).

In light of recent higher-resolution image and topographic data, the lobate flows at Ascreus Mons and Pavonis Mons have been interpreted as subglacially emplaced lava flows (Kadish et al., 2008; Shean et al., 2005); their stratigraphic position, broad, steep-sided morphology, and surficial channels are consistent with this origin, but not with earlier suggestions that they were eskers (Scott et al., 1998) or pyroclastic flows (Zimbelman and Edgett, 1992). The lack of fluvial features associated with the flows raised some questions about this hypothesis (Shean et al., 2005); Kadish et al. (2008) suggested, however, that melt quantities may have been minimal and that any melt could have locally ponded and re-froze. We note that terrestrial subglacial basaltic flows can melt a volume of ice comparable to or greater than their own volume (Höskuldsson and Sparks, 1997; Wilson and Head, 2002; Tuffen, 2007); although the lower temperature of ice on Mars would reduce the melt yield somewhat for these flows, the melt production would still have been of the same order of magnitude as the flow volume (see Section 5). We propose that the melt generated by these flows, which are far from the downslope FSD margin, re-froze in place without carving fluvial features.

Many of the steep-sided flows in the Arsia Mons FSD lie along the strike of pit crater chains farther upslope (Fig. 16). We propose that this morphology is consistent with the landforms originating from eruptions beneath the upslope margin of the glacier, with the transition between pit craters and steep-sided flows marking the transition from thin ice, firn, or snowpack in the overlying paleoglacier to thicker, more consolidated ice farther towards the center of the deposit. The thicker ice farther downslope would provide sufficient confining pressure over the vent to allow an effusive eruption at the downslope end of the rising dike. The more permeable snow, ice, or firn farther upslope would allow meltwater to interact with the rising magma at near-atmospheric pressure, causing explosive phreatomagmatic eruptions to form a chain of pit craters (Wilson and Head, 2007a,d).

Near the downslope extent of the Lobate Facies lies a steep-sided plateau, ∼13 km long and ∼3 km wide, tapering in height from 300 m above the surrounding terrain at its upslope extent to ∼100 m above the surrounding terrain at its terminus. Near the terminus, the center of the plateau has a sunken, leveed appearance, and a thin, delta-shaped lobe ∼4 km long by ∼5 km wide emerges at the distal end (Fig. 17). Based on its flat top and steep sides upslope, we suggest that this ridge emerged as a subglacial ice-confined lava flow similar to the L-shaped plateau. Like the L-shaped plateau, this feature has a leveed, sunken center near...
Fig. 10. The elongate plateau. (a) HRSC DEM superimposed on CTX images. (b) Sketch map showing location of topographic profiles in Fig. 11. (c) Slope map, derived from an HRSC DEM. Contacts between the ridged and knobby facies are shown in white; other geological features are outlined in black.
its terminus. We interpret this as resulting from part of the molten lava core draining downslope. The delta-shaped distal lobe could have formed from this late stage subglacial breakout and drainage of the steeply leveed flow, filling accommodation space provided by the melt cavity in the glacier.

3.6. The Smooth Lower Western Flank unit

The region mapped by Zimbelman and Edgett (1992) as the Smooth Lower Western Flank unit (Fig. 1a, yellow) is generally distinguished by flat lava deposits similar to those erupted subaerially (Fig. 1b, box 3.6). It also contains several steep-sided structures with morphologies not seen elsewhere in this deposit. It borders the Degraded Flank unit (Section 3.7), which is characterized by a series of “irregular scarps” (Zimbelman and Edgett, 1992).

One such feature (Fig. 18) is an exceptionally smooth-topped plateau, ~9 km long by ~5 km wide and ~150 m high (Fig. 18b and c), with sides inclined at angles ~12–23° where HRSC DTM coverage is available (Fig. 18d). Three small craters (200–500 m in diameter) lie along the center of the approximately equidimensional plateau; the plateau surface is otherwise as smooth as the surrounding, apparently subaerial, flows. This feature resembles a terrestrial subglacial volcanic landform known as a tuya, or table mountain.

Tuyas (e.g. Gudmundsson, 2003; Jones, 1968; Smellie, 2007) are landforms that begin as pillow mounds topped by hyaloclastite tephra, with eruptions proceeding in a manner similar to the Northwest, Elongate, and buried plateaus in the western part of

Fig. 11. (a) Topographic profile across the line A–A’ in Fig. 10b. (b) Topographic profile across the line B–B’ in Fig. 10b. (c) Topographic profile across the line C–C’ in Fig. 10b. Topographic data from HRSC-derived DEM.

Fig. 12. Channel and putative hyperconcentrated flow deposits emerging from the elongate plateau and embaying a line of spatter cones to the northwest of the plateau. (a) CTX image mosaic. Prominent moraines cross-cut by the channeled debris are marked with black arrows; the terminal moraine cross-cut by the hummocky flow is marked with a white arrow. (b) Sketch map.
the Arsia FSD (Fig. 1). If the eruption lasts long enough that the melt cavity penetrates to the surface of the ice sheet, the eruption will continue subaerially. A cohesive capping flow will form atop the tephra mound, and deltas of quenched lava will grow outward where the capping flow enters the water in the melt cavity, now an ice-confined subaerial lake. We interpret the exceptionally smooth plateau as a tuya. The shape of the feature in profile (Fig. 18c) strongly resembles those of mafic tuyas in Iceland, British Columbia and Antarctica, such as the Herðubreið tuya (e.g. Werner and Schminke, 1999), Tuya Butte (e.g. Mathews, 1947), and the Jonasen tuya (e.g. Smellie, 2006), respectively. The height of the plateau is somewhat less than those of terrestrial mafic tuyas with the same area (Fig. 19; Smellie, 2009). This could indicate that the explosive phase of the eruption was short, leaving a landform similar to a pillow mound with a subaerial cap; that the ice cover was thin while this plateau was built, allowing a subaerial phase to be reached without much vertical edifice construction; or that the ice-confined lake level, which determines the height of the lava deltas, was lower than in terrestrial settings. The largest “crater” on the surface of the plateau resembles a vent in the slope map (Fig. 18d). Since the plateau is so close to the summit of Arsia Mons (~200 km from the caldera center), the glacial ice covering it during the Arsia tropical mountain glacier period would have been comparatively thin, less than ~1.8 km even if the eruption occurred with the glacier at its maximum extent (Shean et al., 2005; Fastook et al., 2008). Depending on the time of emplacement, the comparatively thin ice in this region may have allowed the eruption to reach an explosive stage rapidly; the tephra would have melted the ice above it rapidly relative to effusive eruptions in deeper ice, quickly breaching the glacier surface and enabling the capping flow to form subaerially.

Another morphology common in the Smooth Lower Western Flank unit is that of the sunken-centered, steep-sided flow. These features resemble the L-shaped and smooth-topped plateaus in planform, height, and steepness, but are much lower in their interiors than at the margins, sometimes level with the surrounding
and lava withdrawal back into the vent.

Subsequent phases of flow emplacement, breakouts, distal drainages, and steep chilled margins of the flow forming during flow emplacement, growth and inflation. The decrease in topography and the steepness of its edges surpass those of subaerial flows, but are comparable to the hypothesized pillow effusions found throughout this region of the deposit. We therefore interpret the feature as a mound of pillow lava erupted from a subglacial fissure. East of this feature, also spanning the boundary between the Smooth Lower Western Flank and the Knobby Facies, lies a partially buried graben, similar in appearance to the Large and Small Grabsen and Aganippe Fossa (Fig. 22). The graben trends south-west-to-northeast, parallel to the Small Grabsen but in contrast to the predominantly north-to-south trend of other, similarly-sized graben in the deposit. West of this partially buried graben lies a steep, irregular scarp ~85 km long. At its southern end, a curved section of the scarp stands up to 400 m tall. We interpret this scarp as the edge of a subglacial flow predating the partially buried graben. A pair of sunken-centered flows superpose the graben and the material burying it.

Three plateaus in this region are associated with these graben (Fig. 1b). Two extend from the Large Grabsen, near the contact between the Degraded Flank and Smooth Lower Western Flank units (Fig. 23a). The larger of these two plateaus extends ~20 km upslope from the east side of the Large Grabsen. It is ~9 km wide, up to 600 m high, and steep sided, with slopes of 20–30° common on all faces covered by HRSC DEMs. It is incised by a deep, sinuous channel at least 8 km long; discontinuous ridges and hollows elsewhere may be sections of the same, partially buried, channel. The taller part of the plateau is smooth but covered in small dunes; the lower part has a hummocky appearance. Below the plateau and level with its base, the graben walls expose thin, horizontal layered flows with relatively even spacing (Fig. 23b and c). In CTX images stretched to enhance contrast at the edge of the plateau, the graben exposes several cohesive lava flows, but talus is abundant, especially higher in the section. Several layers within the talus are brighter than those above and below (Fig. 23b). The bright layers may correspond to compositional or grain size variations in the unconsolidated material. A few of these layers appear to be associated with outcrops of cohesive material.

Based on the morphology of the plateau and its abundance of unconsolidated material relative to the stratigraphically lower flows exposed by the graben (Fig. 23b), we propose that the plateau was built by explosive subglacial eruptions. If this is the case, the tephra is more abundant higher in the sequence because the confining glacial pressure suppressed explosive activity earlier in the eruption, as we interpret to have occurred in the plateaus to the west. We predict that further inspection of this plateau with higher-resolution images will reveal that the consolidated layers in the plateau are substantially thicker than those below it, indicating that they were emplaced in an ice-confined vault, and that they may contain pillow structures.

The smaller, knobby mound (Fig. 23a and d), extending ~4 km outwards from the south end of the Large Grabsen, is ~10 km wide comparable to those of the plateaus to the west. Some mounds are crosscut by local faults and others superimpose the faults (Fig. 21a and b). Due to their size, morphology, and morphometry, we interpret these features as sill-like pillow sheets and pillow mounds (e.g. Smellie, 2007; Cousins et al., 2013), the products of local subglacial eruptions.

One feature, located at the contact between the Smooth Lower Western Flank and Knobby Facies and partially obscured by Knobby Facies material (Fig. 22a and b), is much larger than the others: ~86 km long by ~8–30 km wide and ~200 m high, with an approximately circular superimposed mound about 16 km² in area and 100 m high. The long axis of the feature is approximately parallel to those of the Large Grabsen to the east and Aganippe Fossa to the west. Where HRSC DEMs are available, they show its edges to slope as steeply as 15°. One of the sunken-centered flows (Fig. 20b) lies on its southwestern corner. The thickness of this flow and the steepness of its edges surpass those of subaerial flows, but are comparable to the hypothesized pillow effusions found throughout this region of the deposit. We therefore interpret the feature as a mound of pillow lava erupted from a subglacial fissure.

Several low mounds (Fig. 21a and b), like those forming the base of the glacioclastic plateaus on the western edge of the FSD, are also present in the Smooth Lower Western Flank (Fig. 1a; yellow). The regional slope in this unit is significantly steeper than at the west of the FSD, and nearly all of the mounds are elongated in the downslope direction. Mound maximum heights range from ~100 to 200 m; some slope gently, and others have steep edges comparable to those of the plateaus to the west. Some mounds are crosscut by local faults and others superimpose the faults (Fig. 21a and b). Due to their size, morphology, and morphometry, we interpret these features as sill-like pillow sheets and pillow mounds (e.g. Smellie, 2007; Cousins et al., 2013), the products of local subglacial eruptions.

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Fig. 15. Plateaus buried in the knobby facies (left) and detrended profiles across the plateaus (right). The plateaus in (a and b) have steep mounds morphologically similar to those on the Northwest and elongate plateaus. MOLA elevation map superimposed on CTX images; profile data from 128° per pixel MOLA gridded dataset.
and ~150–200 m tall. Its surface texture is knobby, and its edges are not as steep as those of the larger plateau to the east. Thermodynamic relationships (Wilson and Head, 2007a) and observations of recent terrestrial eruptions (Belousov et al., 2011; Edwards et al., 2012) suggest that supraglacial lava flows can advance kilometers before melting through the snow or ice beneath them. If the flow is able to solidify before the snow or ice is completely removed, the flow will fracture into large blocks as it collapses (Edwards et al., 2012).

Based on this terrestrial analogue and the occurrence of a similar knobby texture at the edges of hypothesized ice-marginal flows elsewhere in the deposit (Section 3.7), our preferred explanation for the knobby mound south of the Large Graben is that it formed as a supraglacial effusive lava flow from a vent within the graben, and gained its current appearance when the ice beneath it was completely removed. The gently sloping sides of this mound could have resulted if the flow was thinner at the edges due to partial melting into the glacier during emplacement, or from mass wasting of loose rubble after the ice was removed.

Alternatively, the knobby mound could have been formed by the emplacement of volcanic tephra from an explosive eruption in the graben. Aganippe Fossa and the other morphologically similar graben in the Arsia Mons FSD, including the Large Graben, may have been formed by large-scale phreatomagmatic eruptions caused by the interaction between the rising magma and meltwater generated by dike emplacement into the glacier (Wilson and Head, 2007d). Hot tephra would have melted the ice below it (Wilson and Head, 2007e), potentially producing the observed uneven surface texture. If this was the formative mechanism for the knobby mound, however, the plan-view shape of the feature and the lack of similar features near other graben in the deposit are unexpected.

The Smooth Lower Western Flank unit (Fig. 1a, yellow) is generally distinguished by a lack of the glacial debris that forms the knobby facies and ridged facies at lower altitudes in the deposit. In this upslope region of the Late Amazonian Arsia Mons glacier, climate and glacial models indicate that snow accumulation exceeded sublimation for most of the glacier’s life span (Fastook et al., 2008), whereas sublimation exceeded accumulation in the downslope regions (ablation zone) now occupied by the ridged and knobby facies. Mouginis-Mark and Rowland (2008) studied High Resolution Imaging Science Experiment (HiRISE) imagery of the Small Graben in the Smooth Lower Western Flank unit.
and noted that no layers among the sequence of lava flows exposed in the graben wall were clearly glacial in origin. While they noted the presence of an unconformable layer that could be glacial sediment, as well as the possibility that the entire sequence exposed by the graben could have been emplaced between periods of high obliquity, we suggest that the lack of obvious glacial debris in the cross-section exposed by the Small Graben is a function of its location in this upslope region of the glacier, where little glacial debris is visible even at the surface.

3.7. Digitate cliffs

Except in the Lobate Facies, the upslope edge of the FSD (Fig. 1b, box 3.7) is marked by numerous overlapping, digitate, flat-topped, steep-sided protrusions (Fig. 24). A typical protrusion is a few kilometers wide and tens of kilometers long. Side slope angles, where HRSC DEM coverage is available, are as steep as 20°, though 10° slopes are more common. The protrusions farthest downslope are the lowest stratigraphically, and they appear subdued relative to the younger ones. The stratigraphically highest protrusions are superposed only by the youngest subaerial flows, pit craters, and concentric graben at the summit of Arsia Mons (Fig. 24a and b). The margins of some protrusions are similar in texture to the knobby mound, and in some cases these knobby margins appear to superpose graben that crosscut the surrounding terrain and the interior regions of the same flows.

Similar digitate features appear at the summits of Ascræus Mons and Pavonis Mons, but only on the sides of their summits in contact with their fan-shaped deposits (Kadish et al., 2008; Fig. 24c). Previous workers have interpreted these cliffs as erosive features caused by a mass flow or high-temperature lavas (Zimbelman and Edgett, 1992), or as analogous to the Olympus Mons basal scarp (McGovern and Solomon, 1993), though Scott and Zimbelman (1995) noted that they more nearly resembled individual flows emplaced prior to the flows above them. On Earth, the Hoodoo Mountain composite volcano in British Columbia is ringed by cliffs up to 200 m tall, whose steep edifices, glassy surface texture, and columnar jointing indicate that they formed as...
subaerial summit lava flows flowed downslope and chilled against a glacier that surrounded the mountain (Edwards and Russell, 2002; Edwards et al., 2002).

On the basis of their morphology and their exclusive association with the contact between subaerial summit lava flows and the edge of the FSD, we interpret the Tharsis Montes digitate cliffs as ice-margin chilled lava flows analogous to those at Hoodoo Mountain (Fig. 25). While the Hoodoo Mountain lavas, like the Mazama Ridge lavas we compare to the L-shaped ridge above, are more silicic than we expect Tharsis lavas to be, basalt flows also develop chilled crusts where they contact and melt ice, confining the flows and producing steep walls of pillow lavas (e.g. Edwards et al., 2009; Skilling, 2009). Higher-resolution images than are presently available for the Arsia Mons cliffs may reveal pillow structures or jointing that will allow more detailed investigation of the lava composition and cooling history; higher resolution DEMs will also provide more insight into the structure of these cliffs. We propose that the knobby margins of the cliffs are regions where the flow overrode the glacial ice, leaving rubble piles when the glacier receded. The superposition relationships of these margins can be explained if the tectonic events creating the graben occurred while ice was still present: when the ice eventually sublimated, the flows that had been supported by ice would have fractured, burying the segments of graben beneath them in rubble.

The digitate cliffs at Arsia form three discrete tiers, with the tier farthest downslope forming the contact between the Degraded Flank and Smooth Lower Western Flank units as we have mapped them (Fig. 1) and the tier farthest upslope forming the contact between the summit and Degraded Flank units; the middle tier and several discontinuous cliffs fall within the Degraded Flank unit. In terrestrial glaciers, ice can accumulate only above the equilibrium line altitude (ELA), the altitude where accumulation of snow into the glacier is equal to the ablation of the glacier by melting or sublimation (Benn and Evans, 2010). On Mars, the increase of sublimation with altitude is often greater than the decrease of temperature with altitude, leading to a second, higher-altitude ELA above which ice does not accumulate (Fastook et al., 2008).

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**Fig. 19.** The volume-to-area ratio of the smooth plateau (red triangle) lies below the trend of terrestrial tuyas (black diamonds) and is more similar to pillow mounds and “sheet-like sequences” (black circles and squares). Reproduced and annotated from Smellie (2009). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

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**Fig. 20.** Sunken-centered, steep-sided flows in the Smooth Lower Western Flank unit (a and b) and the Lobate Facies (c). HRSC DEM (where available) superimposed on CTX images.
We infer that each tier of digitate cliffs represents a series of Arsia Mons summit eruptions in a climate with a different upper ELA for the glacier. The ~4 km difference between the altitude of the upslope limit of the glacier when the highest and lowest tier were emplaced implies that the balance between temperature, atmospheric pressure, and precipitation changed significantly between the first and third summit eruptive episodes represented by the cliffs (Fig. 25), though the influences of these climate variables cannot be deconvolved \textit{a priori} (Fastook et al., 2008). The wide separation between each of the three layers is in keeping with other evidence that construction of the Tharsis Montes was episodic (Wilson et al., 2001).

4. Meltwater: volume and longevity

Interaction of ascending and intruding magma with glaciers and glacial ice deposits produces significant quantities of meltwater that can be stored and transported subglacially, influencing the flow of the overlying ice (cold-based to wet-based flow), and the meltwater can be released catastrophically in jökulhlaups. The quantity of ice melted by the subglacial eruptions recorded in the Arsia Mons fan-shaped deposit, and the longevity of the resulting liquid water bodies, are of primary astrobiological interest. We use thermodynamic relationships, validated by terrestrial measurements, to estimate the volume of meltwater produced in the construction of several Arsia Mons subglacial edifices, and the time scale over which this water remains liquid (e.g., Wilson and Head, 2002, 2007a, 2009; Gudmundsson, 2003; Head and Wilson, 2007; Tuffen, 2007; Wilson et al., 2013).

The heat energy $E$ released by the rapid chilling of a volume $V_m$ of magma (Gudmundsson, 2003) is given by:

$$E = \rho_m V_m c_g \Delta T_m,$$

where $V_m$ is the volume of magma, $\rho_m$ is the density of magma, $c_g$ is the specific heat capacity of the glass formed by rapid chilling of the magma, and $\Delta T_m$ is the temperature change the magma undergoes.
We use the specific heat of basaltic magma as a conservative estimate of the specific heat of basaltic glass (Gudmundsson, 2003). The heat energy $Q$ required to melt a volume $V_i$ of ice (Gudmundsson, 2003) is given by:

$$Q = \rho_i V_i [c_i (T_0 - T_1) + L_i + c_w (T_2 - T_0)],$$

where $\rho_i$ is the density of ice, $c_i$ is the specific heat capacity of ice, $T_0$ is the melting temperature of ice, $T_1$ is the initial temperature of the ice, $L_i$ is the latent heat of fusion of ice, $c_w$ is the specific heat capacity of liquid water, and $T_2$ is the temperature of the resulting meltwater.

The efficiency with which the magma can transfer its heat to the ice, $f_{\text{mi}}$, is:
Fig. 24. Steep-sided, digitate flows at the upslope edge of the Arsia Mons fan-shaped deposit. (a) THEMIS daytime IR image mosaic. (b) Sketch map. Steep-sided flows are outlined in black, pit craters are outlined in violet, impact craters are outlined in blue, tectonic graben are outlined in green, and the Small Graben is outlined in red. Knobby flow margins are shaded in gray, and generally appear to superpose tectonic features that cross-cut the steep-sided flows. (c) Comparison between digitate flows at the upslope edge of the Arsia Mons (left), Pavonis Mons (center) and Ascraeus Mons (right) fan-shaped deposits, plotted at the same scale for comparison. THEMIS daytime IR mosaics. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
We make the following assumptions: (1) the thermodynamic properties of lavas in the Arsia FSD were similar to those of terrestrial basalts (Table 1), (2) the internal temperature of the ice was similar to the modern martian equatorial surface temperature at high obliquity, (3) integrated over the course of the eruption, $f_{mi} \approx Q/E$, and (4) the final temperature reached by the water is 15 °C, typical of terrestrial eruptions (Gudmundsson, 2003; Tuffen, 2007). Then the volume of ice melted by a volume of magma $V_m$ is equal to:

$$V_m = \frac{f_{mi} \rho_m C \Delta T_m}{\rho_i [c_i (T_0 - T_1) + L_i + c_w (T_2 - T_0)]},$$

(4)

for a corresponding volume $V_w$ of meltwater:

$$V_w = \frac{f_{mi} \rho_m C \Delta T_m}{\rho_w [c_i (T_0 - T_1) + L_i + c_w (T_2 - T_0)]},$$

(5)

where $\rho_w$ is the density of water. Under these assumptions, Eq. (5) indicates that the formation of the Northwest plateau melted $\sim 39$–$46$ km$^3$ of ice, producing $\sim 36$–$42$ km$^3$ of liquid water: $\sim 22$ km$^3$ during the formation of the pillow sheet, and $\sim 14$–$20$ km$^3$ during the formation of the central tephra mound. The formation of the L-shaped ridge, for example, would have melted another $\sim 38$ km$^3$.

Due to the channel-forming jökulhlaup that occurred at the Northwest plateau, much of the water formed in the first phase of that eruption would have frozen or evaporated before the eruption ceased. Some of the water formed in the explosive phase is also likely to have escaped beneath the glacier before the glacier refroze to its bed. If most of the water from the explosive phase remained confined, it may have survived long enough to become inhabited by any extant surface life.

If, as an example, we model the melt cavity formed by the central mound of the Northwest plateau as a uniform planar sheet covering the same area as the mound, and use the more conservative estimate of meltwater generated by this phase of the eruption, the meltwater layer is $\sim 500$ m thick. The fact that the $\sim 150$ m tall central mound did not form a subaerial lava cap suggests that the meltwater layer was at least $150$ m thick. Though lakes should become increasingly saline as they freeze, we will conservatively assume fresh water for these calculations. Heat loss from the water to the cooling edifice below it is likely to be small compared to loss to the ice above (Wilson and Head, 2002), since conductive heat flux across a surface is proportional to the temperature gradient across that surface. Therefore, we consider only heat transfer from

$$f_{mi} = \frac{dQ}{dt} / \frac{dE}{dt},$$

(3)

and has been empirically estimated to equal $\sim 0.1$ for pillow lavas and $\sim 0.5$–$0.7$ for hyaloclastite tephra (Gudmundsson, 2003; Tuffen, 2007).

---

**Table 1**

Values used in ice melt calculations.

<table>
<thead>
<tr>
<th></th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northwest plateau, pillow sheet</td>
<td>$3.1 \times 10^{10}$ m$^3$</td>
</tr>
<tr>
<td>$f_{mi}$</td>
<td>0.1</td>
</tr>
<tr>
<td>Northwest plateau, tephra mound</td>
<td>$3.9 \times 10^{10}$ m$^3$</td>
</tr>
<tr>
<td>$f_{mi}$</td>
<td>0.5–0.7</td>
</tr>
<tr>
<td>L-shaped ridge</td>
<td>$5.2 \times 10^{10}$ m$^3$</td>
</tr>
<tr>
<td>$f_{mi}$</td>
<td>0.1</td>
</tr>
<tr>
<td>Basaltic lava</td>
<td>$&gt;1200$ J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$\rho_m$</td>
<td>2300 kg m$^{-3}$</td>
</tr>
<tr>
<td>$T_m$</td>
<td>1200 °C</td>
</tr>
<tr>
<td>Water</td>
<td></td>
</tr>
<tr>
<td>$L_i$</td>
<td>$3.35 \times 10^3$ J kg$^{-1}$</td>
</tr>
<tr>
<td>$\rho_i$</td>
<td>917 kg m$^{-3}$</td>
</tr>
<tr>
<td>$T_c$</td>
<td>$-60$ °C</td>
</tr>
<tr>
<td>$c_i$</td>
<td>2050 J kg$^{-1}$</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>1000 kg m$^{-3}$</td>
</tr>
</tbody>
</table>
the water to the ice above it. Under these simplifying assumptions, the time to freeze the melt layer is given by Kreslavsky and Head (2002):

\[
t(h) = \frac{L_r h^2}{2K_1(T_0 - T_i)}
\]

where \(L_r\) and \(\rho_i\) are the latent heat of fusion and density of ice; \(h\) is the thickness of the melt layer; \(K_1\) is the thermal conductivity of ice; 2.5 W m\(^{-1}\) K\(^{-1}\); \(T_0\) is the melting temperature of the melt water, 0°C for fresh water; and \(T_i\) is the surface temperature, −60°C for Amazonian equatorial Mars at high obliquity. Thus we estimate that a 150 m thick layer would not freeze through for ~370 years, or ~8000 years for a 500 m thick layer. Even the lower estimate would represent a sufficient time for many generations of microbes to have flourished in this melt body, if they existed to colonize it.

5. Conclusions

We conclude that glaciovolcanic landforms are abundant in the Arsia Mons fan-shaped deposit. These include landforms interpreted as subglacial pillow sheets larger than any known on Earth, hypabyssal mounds, large englacial and glacier-marginal ice-con fined flows, and a tuya. The presence of three well-separated strata of ice-marginal lava flows at the upslope margin of the deposit indicates that near-summit eruptions occurred in three distinct sets of climate conditions.

Glaciovolcanism in the deposit would have melted massive volumes of ice at the upslope margin, throughout much of the Lobate Facies, and at numerous eruptive centers throughout the rest of the deposit. The best-exposed of these eruptions, at the Northwest plateau, would have created an englacial lake that may have persisted for hundreds to thousands of years. This meltwater production and glaciovolcanic heat transfer caused local wet-based flow in the Arsia Mons glacier (Scanlon et al., in preparation, 2014), creating additional habitable environments in the Arsia Mons FSD.

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References
