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Terrestrial analogs and thermal models for Martian flood lavas

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Abstract. The recent flood lavas on Mars appear to have a characteristic "platy-ridged" surface morphology different from that inferred for most terrestrial continental flood basalt flows. The closest analog we have found is a portion of the 1783-1784 Laki lava flow in Iceland that has a surface that was broken up and transported on top of moving lava during major surges in the eruption rate. We suggest that a similar process formed the Martian flood lava surfaces and attempt to place constraints on the eruption parameters using thermal modeling. Our conclusions from this modeling are (1) in order to produce flows >1000 km long with flow thicknesses of a few tens of meters, the thermophysical properties of the lava should be similar to fluid basalt, and (2) the average eruption rates were probably of the order of 10^4 m^3/s, with the flood-like surges having flow rates of the order of $10^5 - 10^6$ m^3/s. We also suggest that these high eruption rates should have formed huge volumes of pyroclastic deposits which may be preserved in the Medusae Fossae Formation, the radar "stealth" region, or even the polar layered terrains.

1. Introduction

The objective of this paper is to provide initial qualitative and quantitative interpretations of new data on Martian flood lavas. The Mars Orbital Camera (MOC) on board the Mars Global Surveyor (MGS) spacecraft has returned spectacular new images of Mars [*Malin et al.*, 1998]. We were particularly interested in images of the Martian flood lavas in order to compare the emplacement of Martian and terrestrial flood volcanism.

Flood volcanism is one of the most significant crust-forming processes identified on Mars [e.g., *Greeley et al.*, 2000; *Greeley and Spudis*, 1981], and the extreme paucity of craters in some of the images of the flood lavas suggests that this style of volcanism might plausibly persist to this day on Mars [*Hartmann*, 1999; *Hartmann et al.*, 1999; *Hartmann and Berman*, this issue]. Understanding the eruptions that formed these flood lavas is a critical step in understanding the evolution of the surface, interior, and atmosphere of Mars.

Before proceeding, we should note that we use the term "flood lava" to denote a package of sheet-like lava flows that inundates a large region, producing a smooth plain without ~100 m tall constructs. This use of the word is in line with that presented by *Greeley and King* [1977] and its application to terrestrial continental flood basalts [*Bates and Jackson*, 1984] and is essentially identical to the use of the term "mare" on the Moon (but without the required low albedo). However, some other workers have used "flood lava" to simply refer to long lava flows with no clear channels or tubes [e.g., *Cattermole*, 1990].

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Some of the most recent and best preserved flood lavas lie in eastern Elysium Planitia (the Cerberus Formation of *Plescia* [1990]) and in a portion of Amazonis Planitia centered at 30° N, 160° W [*McEwen et al.*, 1999b]. These two regions are connected by Marte Vallis, which is also partially filled by flood lavas. The surface of the Cerberus plains records a complex history of volcanic, fluvial, impact, tectonic, aeolian, and possibly lacustrine and glacial processes [e.g., *Mouginis-Mark et al.*, 1984; *Tanaka and Scott*, 1986; *Plescia*, 1990; *Scott and Chapman*, 1991; *McEwen et al.*, 1998, 1999c]. In this paper we focus solely on the best preserved (and presumably youngest) volcanic features.

Well-preserved flood lavas of similar appearance (although probably older) are also present high on the Tharsis Bulge, near Tharsis Tholus. Layers exposed in the walls of Valles Marineris are also likely to be yet older flood lavas [*McEwen et al.*, 1999a]. Figure 1 shows the location of these flood lava exposures with respect to other major features on Mars. Crater counting and models suggest ages of about 10 Ma for the youngest of the flows in the Cerberus Plains [*Hartmann and Berman*, this issue]. However, MOC images also reveal that there are locations where the lava is eroding out from beneath the Medusae Fossae Formation, which may have protected the lava from small impact craters. While many processes can complicate the age estimates from crater counts, the crisp, undegraded morphology of the flows supports the idea that these flood lavas are geologically very young. Together, the recent flood lavas appear to cover an area of about 10^7 km^2, roughly the size of Canada [*McEwen et al.*, 1998]. The minimum thickness of the Cerberus deposits is estimated to be in the hundreds of meters, and the new MOC images support the estimate by *Plescia* [1990] that individual flow fronts are only of the order of 10 m thick.

These flood lavas have a distinctive surface morphology dominated by plates and ridges [*McEwen et al.*, 1999b] (Figures 2-9). The platy-ridged material is interpreted to be lava because it is bounded by lobate margins characteristic of lava flows (Figure 4). At high spatial resolution (2 to 20 m/pixel), these flows exhibit a complex surface morphology interpreted to include rare channels (Figures 3 and 4), common rafted crustal slabs (Figures 2, 3, 5, 6, 7, and 9), ubiquitous pressure ridges (Figures 2-9), some ponded surfaces (Figures 5-7), and possible squeeze-ups (Figures 2 and 5-7). Coherent slabs of crust appear to range in size from a few hundred meters to a few kilometers. Ridges are typically only tens of meters wide but up to kilometers in length. Most ridges are interpreted to have formed at the edges of plates as they are rammed into each other during emplacement (i.e., pressure ridges). Other ridges may have formed by liquid lava oozing up through extensional cracks in the plates (i.e., squeeze-ups).

These flows are remarkably flat, with about 400 m of vertical elevation over 1500 km of lateral distance (0.027% slope) [*Smith et al.*, 1999]. However, the region is radar bright, implying a rough surface on the decimeter scale [*Harmon et al.*, 1992, 1999].

Vents in the flood lava regions are difficult to identify, probably as a result of burial by the flows, their small size, and/or rapid removal of the relatively fragile vent structures. However, the Cerberus Rupes fractures and nine low shields on the western side of the Cerberus Plains were identified as likely vents from Viking data [*Plescia*, 1990; *Edgett and Rice*, 1995]. The most recent MOC images show small lava flows emanating from both sides of some of the Cerberus Rupes fissures, confirming that they indeed served as volcanic vents [*McEwen et al.*, 1999c]. However, there is still no conclusive proof that most of the large flood lavas were fed from Cerberus Rupes. Also, the Cerberus Rupes fractures cut all the lava flows, indicating that it has been tectonically active more recently than it has been a volcanic vent.

One of the ongoing limitations in working with the Martian flood lavas is our inability to trace individual flows. The Viking data do not have the resolution to allow flow margins to be confidently traced. The MOC data cover only a tiny fraction of the Cerberus Plains, allowing only extremely discontinuous sampling. Our best estimate for the extent of a single flow is shown in the lightest shading in Figure 1. This area corresponds to the brightest radar returns reported by *Harmon et al.* [1999]. It also follows the local topographic gradient away from the presumed vents along Cerberus Rupes. It is relatively dark in the Viking images, and the freshest looking lavas in MOC images fall within this area. Also, the radar bright area terminates at the same area as the flow terminus seen in Figure 4.

On a smaller scale the flows appear to have 100- to 1000-m-size lobes at the margins, but the larger sheets are generally not captured in any single MOC frame (3-10 km wide). From the small areas sampled by MOC and the subtle mottling in the Viking images, we tentatively infer that both the widths and lengths of the individual sheets are of the order of tens of kilometers. Only a few sheet-like lobes can be seen to be less than a kilometer wide (e.g., Figure 4). However, if previous suggestions for vent locations are correct, the lavas must have generally spread for >500 km and may have exceeded 2000 km in some cases. This implies that each flow is composed of a great number of smaller sheet-like lobes.

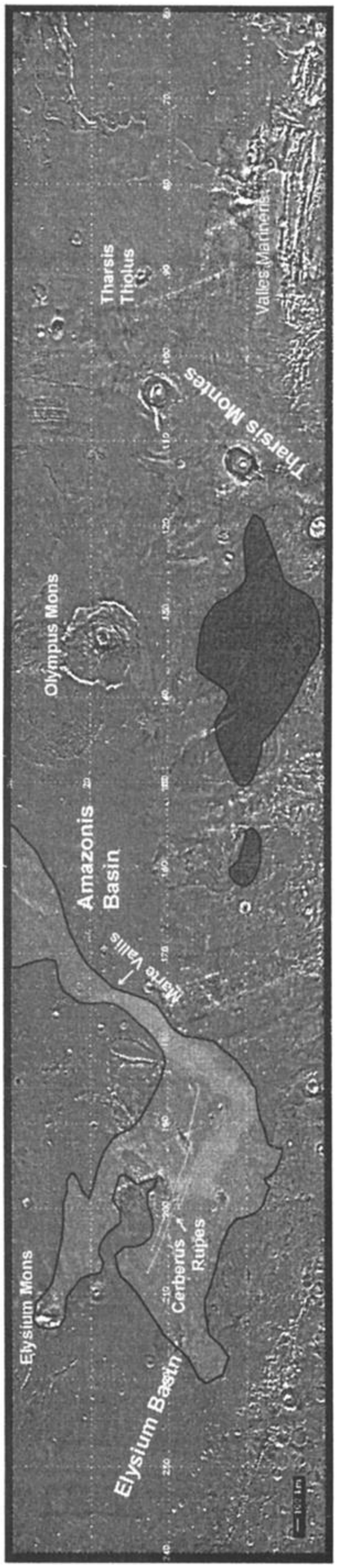


Figure 1. Location map of Mars. Recent, well-preserved flood lavas have been found in the Elysium and Amazonis Basins and Marte Valles which connects the two basins. Older flood lavas are exposed around Tharsis Tholus, and layers in Valles Marineris have been interpreted as possible ancient flood lavas [*McEwen et al.*, 1999a,b]. The shaded areas roughly outline the radar bright and radar dark (i.e., "stealth") regions seen by *Butler* [1994] and *Harmon et al.* [1992, 1999]. The brightest elongate patch corresponds to the brightest radar returns (roughest surface) and also contains the best preserved (and presumably youngest) flood lava surfaces. The Medusae Fossae Formation mostly lies between the radar bright and dark regions [*Scott and Tanaka*, 1986].

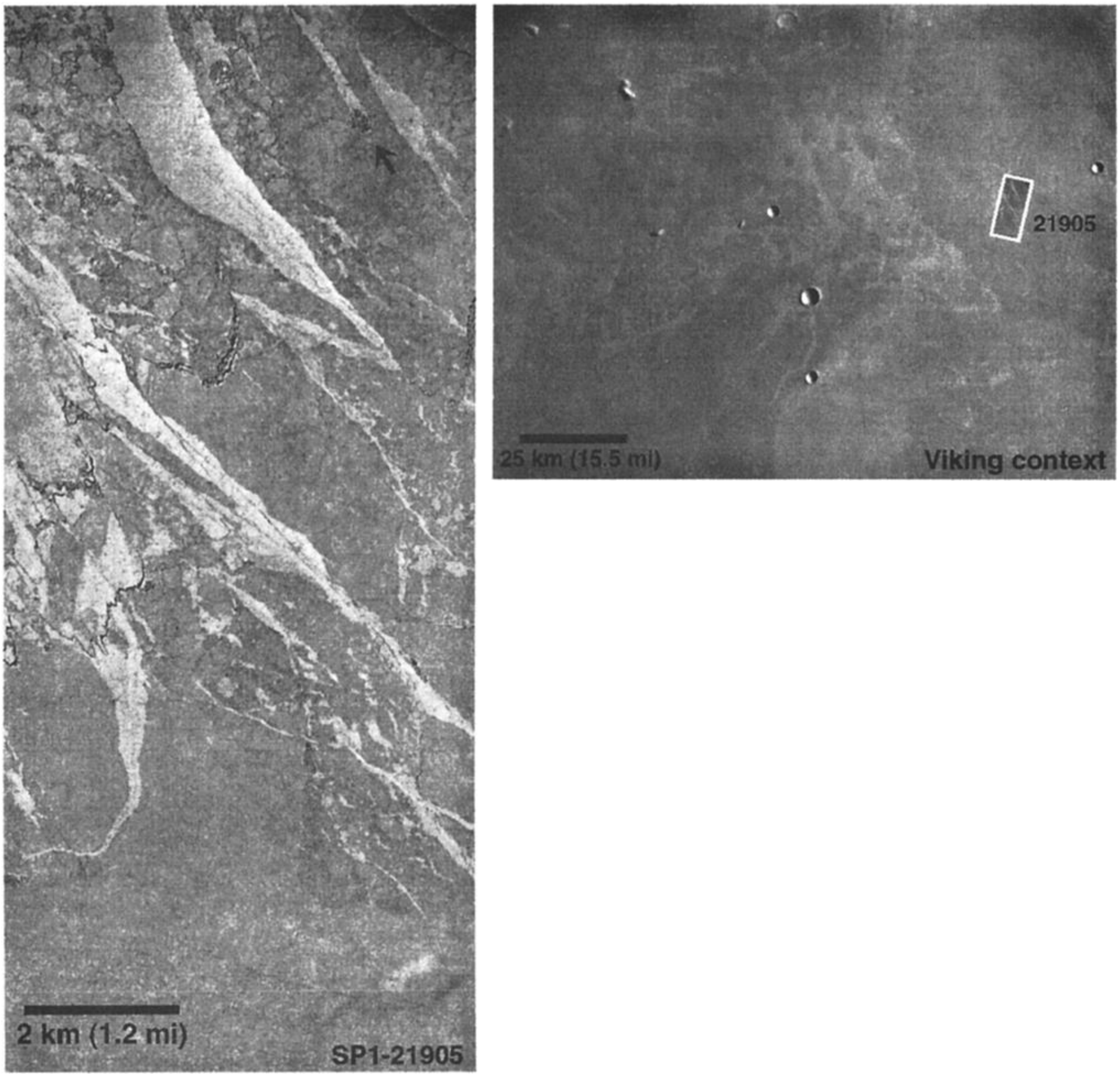


Figure 2. MOC image SP1-21905 and Viking context image from Elysium Planitia centered at 5.7°N, 209.0°W. In this and the following MOC images, north is up, and the Viking context is from a basemap mosaic [*Davies, et al.*, 1992]. Note rafted plates of lava that can be jigsaw fit in a pattern consistent with flow to the NW or SE. The wake behind an obstacle in the NE corner of the image (arrow) shows that the flow direction was actually to the NW. Larger, curving ridges in the plates are probably pressure ridges formed during the collision of plates. Thin, involute ridges within the brighter material between the plates are probably squeeze-ups of lava formed after the lava surface stagnated.

2. Terrestrial Analogs

Without well-studied terrestrial analogs to provide a qualitative model for the Martian flood lavas, it is impossible to pin down the physical processes to include in quantitative lava flow models. Current lava flow models require some basic assumptions about how the flow advances (e.g., tube-fed versus channel-fed). Even empirical relationships between eruption parameters and flow dimensions [e.g., *Walker*, 1973; *Pieri and Baloga*, 1986; *Kilburn and Lopes*, 1991] are critically affected by the style of emplacement. In this section we will show that portions of the 1783-1784 Laki lava flow in Iceland may be the best available analog to the Martian flood lavas.

2.1. Continental Flood Basalts

The Martian flood lavas derive their name from their obvious similarities to terrestrial continental flood basalts (CFBs). Continental flood basalt provinces typically cover more than 10^6 km^2 with a lava pile kilometers thick [e.g., *Macdougall*, 1988]. When not deeply eroded, they are also characterized by flat, plateau-like, topography [e.g., *Greeley and King*, 1977]. The best preserved and most extensively studied CFB province is the Columbia River Basalt Group (CRBG). The CRBG consists of 174,300 ±31,000 km^3 of dominantly basaltic-andesite lava covering 163,700 ±5000 km^2 of area. These lavas are divided into approximately 40 stratigraphic units that contain >300 individual "flows". The lavas have radiometric ages between 17 and 6 Ma, but the majority of the lava was erupted between 16.5 and 15.3 Ma [*Tolan et al.*, 1989]. Comparison to other CFB provinces shows that most CRBG lava flows are similar to the majority of lava flows found in other flood basalt provinces [e.g., *Hooper*, 1982; *Self et al.*, 1998].

Although continental flood basalt provinces have been studied extensively, their flows are not sufficiently well preserved or exposed for the original surface morphology to be directly observed. Instead, the surface morphology must be inferred from observations made in cross sections along natural and artificial cliffs. Such examination of the CRBG lava flows indicates that they were primarily emplaced as inflated pahoehoe lava flows [*Self et al.*, 1997, 1998]. However, it is important to note that only one unit, the Roza Member of the Wanapum Formation, has the combination of exposure and preservation that has allowed the details of an entire flow field to be described [e.g., *Shaw and Swanson*, 1970; *Swanson et al.*, 1975; *Martin*, 1989; *Thordarson*, 1995; *Thordarson and Self*, 1998]. Most other CRBG flow fields are either too poorly exposed or too extensively eroded for such detailed study. Thus there is a continuing debate about the role of inflation in the emplacement of other CRBG lava flows [e.g., *Reidel*, 1998]. Other terrestrial flood basalt provinces (including the Deccan, Entendeka, Kerguelen, Parana, and Keewena) also appear to be dominated by inflated pahoehoe [*Self et al.*, 1997, 1998; *Jerram et al.*, 1998; *Keszthelyi et al.*, 1999; *Frey et al.*, 2000]. However, a substantial fraction of flows in the CRBG and other flood basalt provinces have brecciated flow tops that do not fit in the aa versus pahoehoe categorization [*Macdonald*, 1953; *Self et al.*, 1998; *Frey et al.*, 2000].

Based on the surface morphology of large inflated pahoehoe flows in Hawaii, New Mexico, Australia, and elsewhere [cf., *Theilig and Greeley*, 1986; *Keszthelyi and Pieri*, 1993; *Hon et al.*, 1994; *Stephenson et al.*, 1998; *Self et al.*, 1998], we infer that a substantial portion of the surface of CFB lavas consists of a convoluted set of lobes, tumuli, and plateaus, all inflated to about the same level. Such a surface has a distinctive mottled appearance at 1-10 m/pixel, especially when partially infilled by aeolian deposits. Figure 10 shows this surface morphology of terrestrial inflated pahoehoe lava flows in a variety of settings. This morphology is diagnostic of inflation; no other style of emplacement produces this morphology.

Do we expect inflated flows to be common on Mars? Inflated flows require the presence of a coherent upper crust

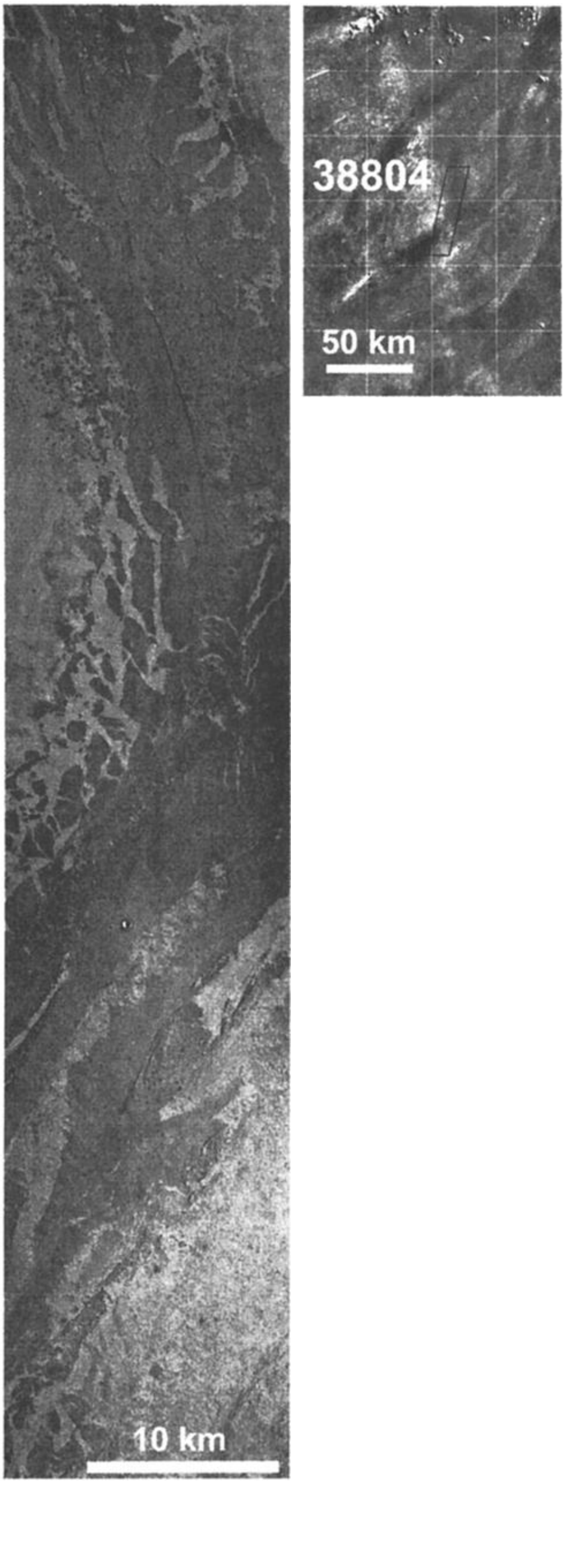


Figure 3. MOC image SP2-38804 and Viking context from Marte Vallis centered at 6.9°N, 183.0°W. Note plates in the lava flow consistent with the center of the a channel winding North. The parallel ridges in the top of the image appear more similar to compressional pressure ridges formed by the collision of solid plates than to levees formed by overbank deposits of liquid lava. The streamlined feature SE of the center of the image appears to be sheets of crust piling up against a topographic obstacle, leaving a wake in the crust. This type of wake is seen on a much smaller scale in Hawaii when channelized lava piles against a tree. The grooves seen in the Laki flow field in Iceland are more similar in scale (see Figures 12 and 13).

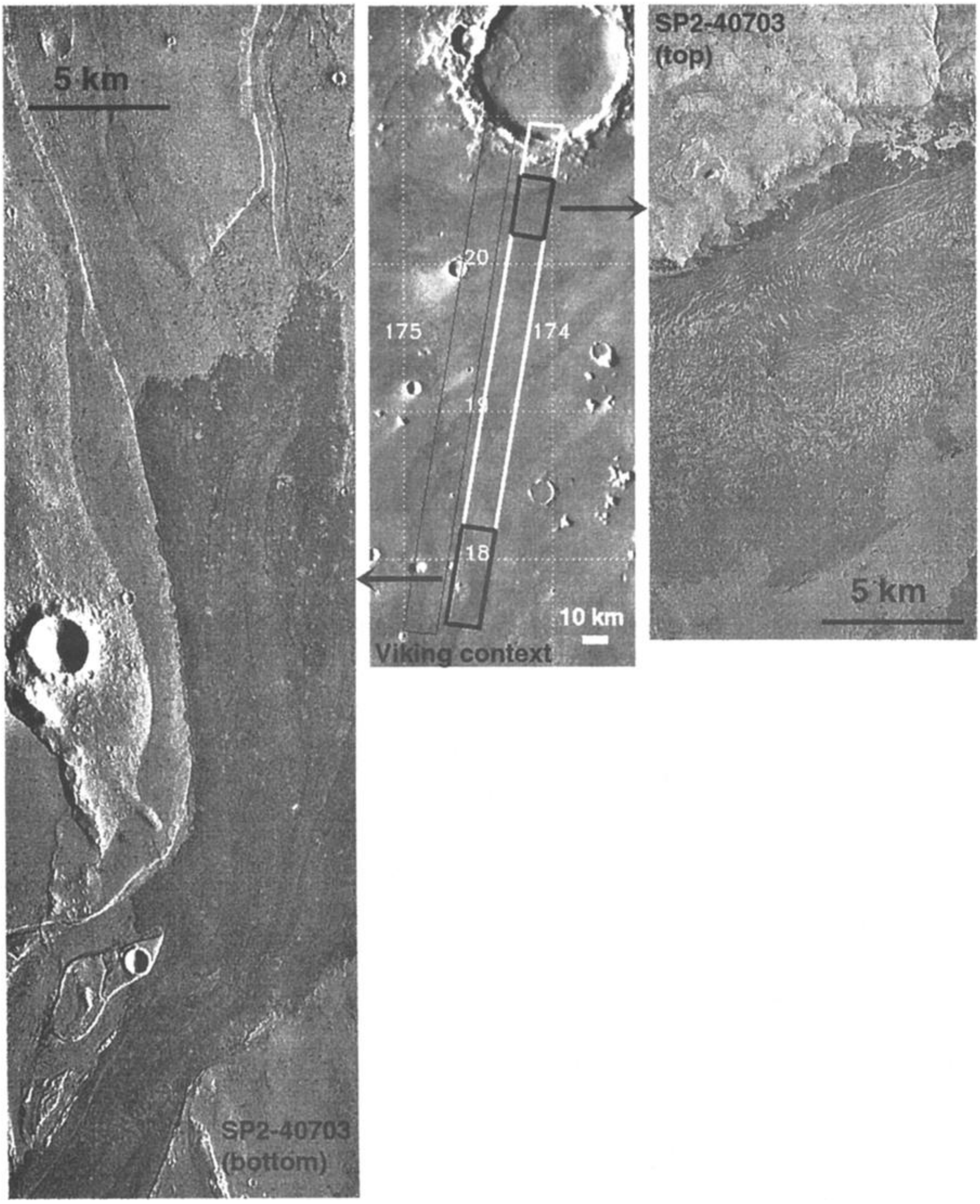


Figure 4. Cutout windows from MOC image SP2-40703 and Viking context. Image centered at 19.2°N, 174.6°W inside Marte Vallis. Lower (left) cutout window shows the terminus of a flow lobe that was confined to preexisting channels presumed to be fluvial in origin. The margins of the flow are typical for basaltic lava flows. The albedo markings within the flow are suggestive of multiple channels that did not have the opportunity to drain. The flow margins and surface appear to be only minimally modified by postemplacement geologic processes. The upper (right) cutout window shows the margin of a flood lava flow with smooth sheet-like channel overbank deposits and disrupted channel surface consistent with flow to the NE. Also of interest are the small fluvial channels that the lava embays. The margin of the flow is partially covered by fluvial and/or aeolian sediments. Numbers on context image are latitude and longitude.

and conditions under which it is easier for the lava to raise the crust than to advance forward. The major difference between the Earth and Mars is the lower gravitational acceleration on Mars. This decreases both the force driving the advance of the flow and the force holding the crust down. Thus the reduced gravity should significantly enhance the tendency of lava flows to inflate [*Keszthelyi and Self*, 1998a]. The other major difference between the Earth and Mars is the ~100

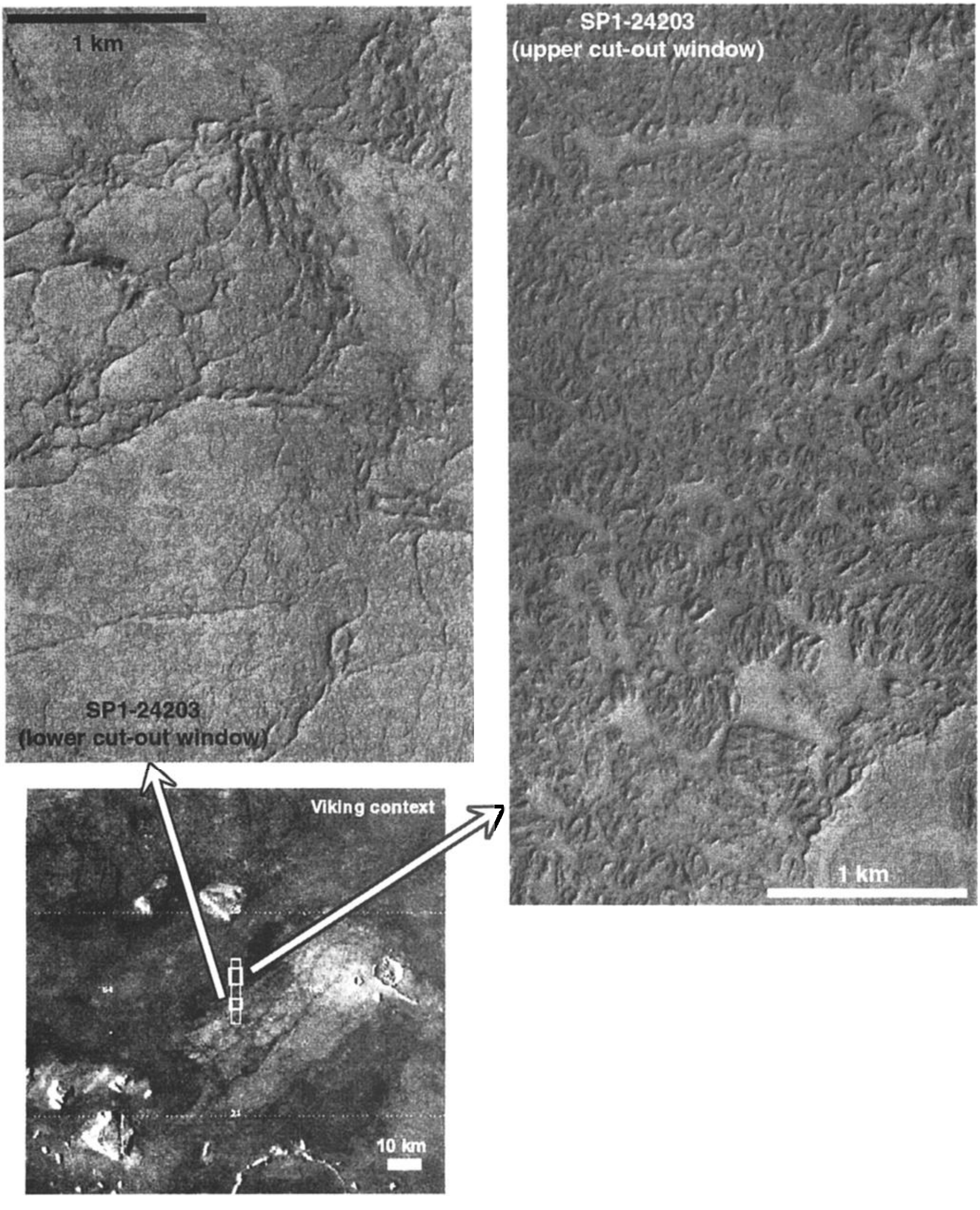
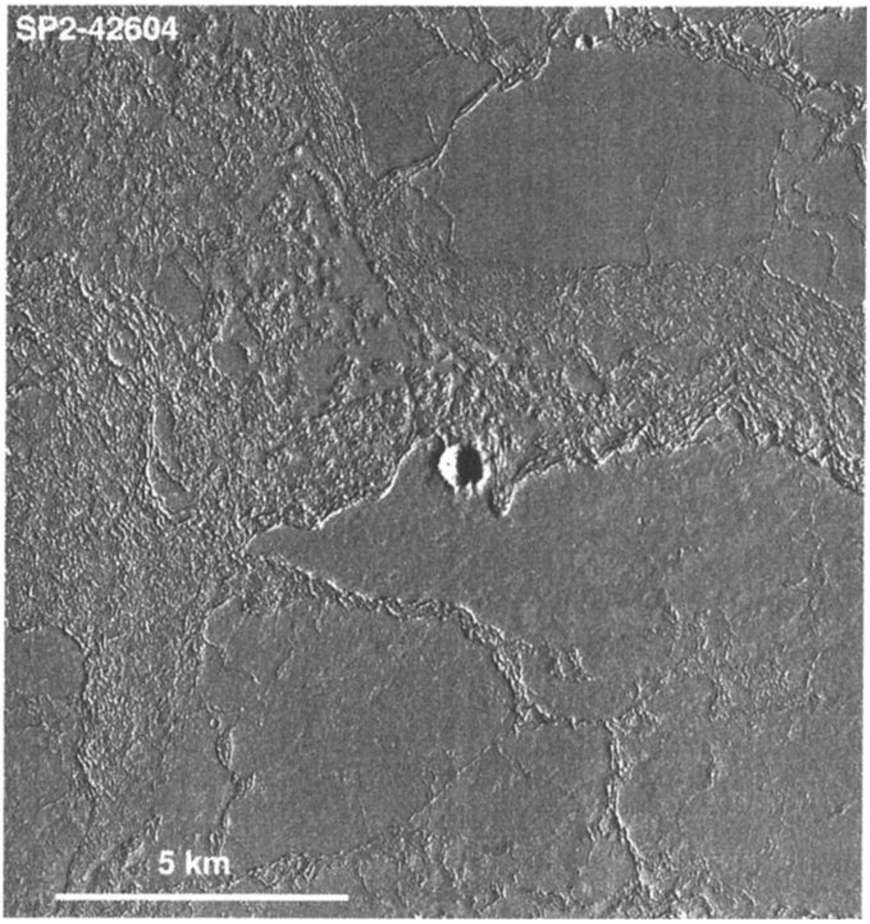


Figure 5. Cutout windows from MOC image SP1-24203 and Viking context from Amazonis Planitia. Image centered at 24.6°N, 163.6°W. In the upper (right) cutout window, note the ridges on the plates on the lava surface at the top of the image. Our interpretation of the formation of this platy-ridged surface is the following: First, the ridges formed as compressional hummocks perpendicular to flow direction. These ridges do not appear to be inflation features (see Figure 10) and instead are similar to compressional pressure ridges formed in brecciated flow tops (slab pahoehoe, aa, etc.). We suggest that below the brecciated flow top, the lava solidified into a coherent slab. A later surge in lava flux then rafted pieces of this crust, translating and rotating blocks with ridges on their surface (see Figure 15). In the lower (left) cutout window the same flow exhibits a much smoother appearance. The random orientation of the web of thin convoluted ridges on the smooth lava surface suggests that they are squeeze-ups of molten lava coming up between cracks in the crust on a ponded area. Pressure ridges would be expected to show a clear preferred orientation perpendicular to flow.



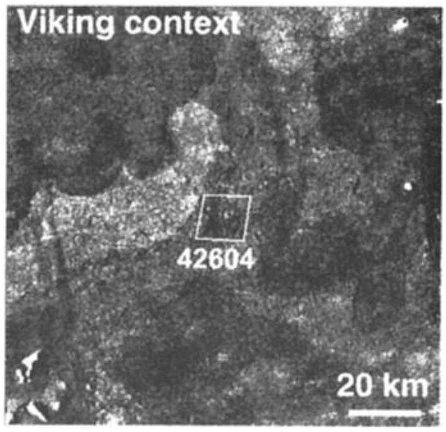


Figure 6. MOC image SP2-42604 and Viking context from Amazonis Planitia centered at 25.7°N, 167.0°W. Groups of parallel ridges are probably compressional pressure ridges of disrupted crust, and more sinuous, isolated ridges are probably large squeeze-ups in the flat ponded surfaces. Local flow direction is ambiguous, but motion to the SW is most consistent with the ridge orientations.

times lower atmospheric pressure on Mars. This should lead to reduced convective cooling of the lava surface and slower crust growth on a lava flow. However, the initial formation of a crust is governed by thermal radiation, which will be very similar on Earth and Mars [e.g., *Wilson and Head*, 1994]. Overall, the conditions on Mars appear to be very conducive for the formation of inflated flows. What remains to be seen is if the eruptions on Mars were also in the regime that forms inflated flows.

Locally high strain rates (relative to the lava viscosity) lead to tearing and disruption of the crust [*Peterson and Tilling*, 1980]. In Hawaii, aa flows with disrupted crusts begin to form at effusion rates above about 15 m^3/s [*Rowland and Walker*, 1990]. However, the Roza Member of the CRBG is an inflated pahoehoe flow that was emplaced at approximately 2000 m^3/s [*Thordarson*, 1995; *Thordarson and Self*, 1998]. In the case of the Roza eruption, the lava front was tens of kilometers wide, so the much higher volumetric effusion rate did not translate into rapid flow rates or high strain rates. Thus high volumetric effusion rates, by themselves, do not rule out emplacement as inflated pahoehoe sheet flows.

The layered deposits seen in the walls of Valles Marineris appear similar to the layers of lava flows commonly seen in CFB provinces [*McEwen et al.*, 1999a]. However, we lack the resolution needed (<<10 cm/pixel) to identify diagnostic features (tumuli, inflation clefts, lobe boundaries, vesicle distribution, etc.) in the Martian cliffs. Instead, we make the assumption that most of the Martian flood lavas through time have had surfaces like the recent flood lavas. Our examination of the MOC images of the exposed surfaces of the well-preserved Martian flood lavas suggests that only isolated segments of their margins might be inflated pahoehoe. The rafted plates that characterize the Martian flood lavas are not seen on inflated terrestrial pahoehoe flows and instead suggest a disrupted crust that translates laterally during emplacement. Thus, while the scale and gross physiography of the terrestrial and Martian flood lavas appear similar, the more detailed flow morphology revealed by the MOC images suggests that they formed by significantly different processes. Therefore we look for other terrestrial analogs to provide the needed insight into the emplacement processes of Martian flood lavas.

2.2. The 1783-1784 Laki Eruption

The eruption in 1783-1784 at Laki, Iceland, produced a large flow field that has many similarities to CFB lava flows, but also has areas with a morphology that is strikingly similar to the Martian flood lavas. The Laki eruption is the largest

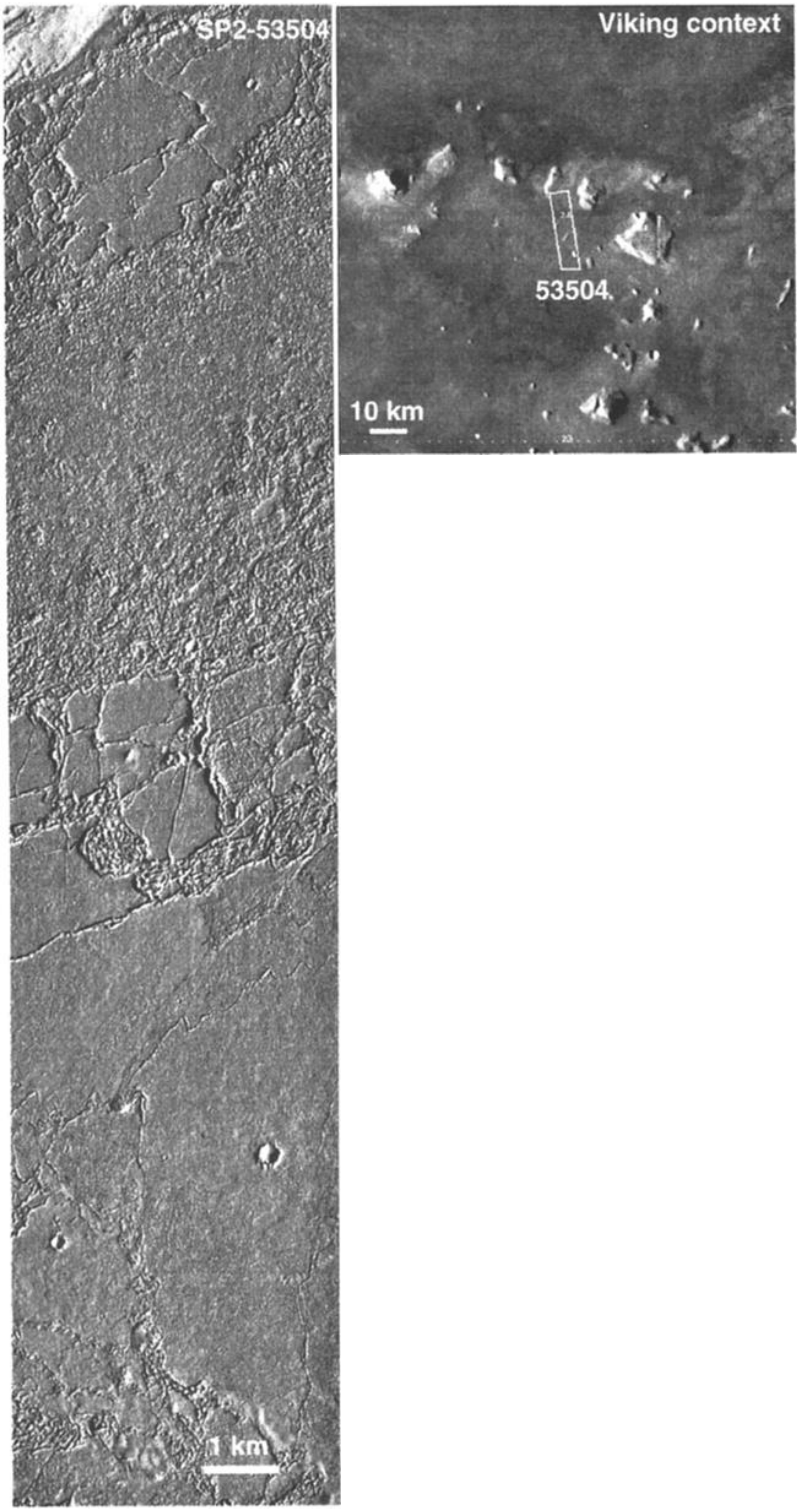


Figure 7. MOC image SP2-53504 and Viking context from Amazonis Planitia, centered at 23.9°N, 164.3°W. Rough ridged areas and the most prominent linear ridges appear to be pressure ridges of brecciated material. Smooth areas appear to be ponded lava. In general, the rough areas match the dark mottles in the Viking context images. The margin of the flow, where it abuts the hills to the north, is not morphologically distinct from the rest of the lava. Dunes along the flow margin and the sparse craters suggest that this flow might be slightly older than some of the Elysium Planitia flows. Flow direction is ambiguous, with no clear interaction between the lava flow and the older hills.

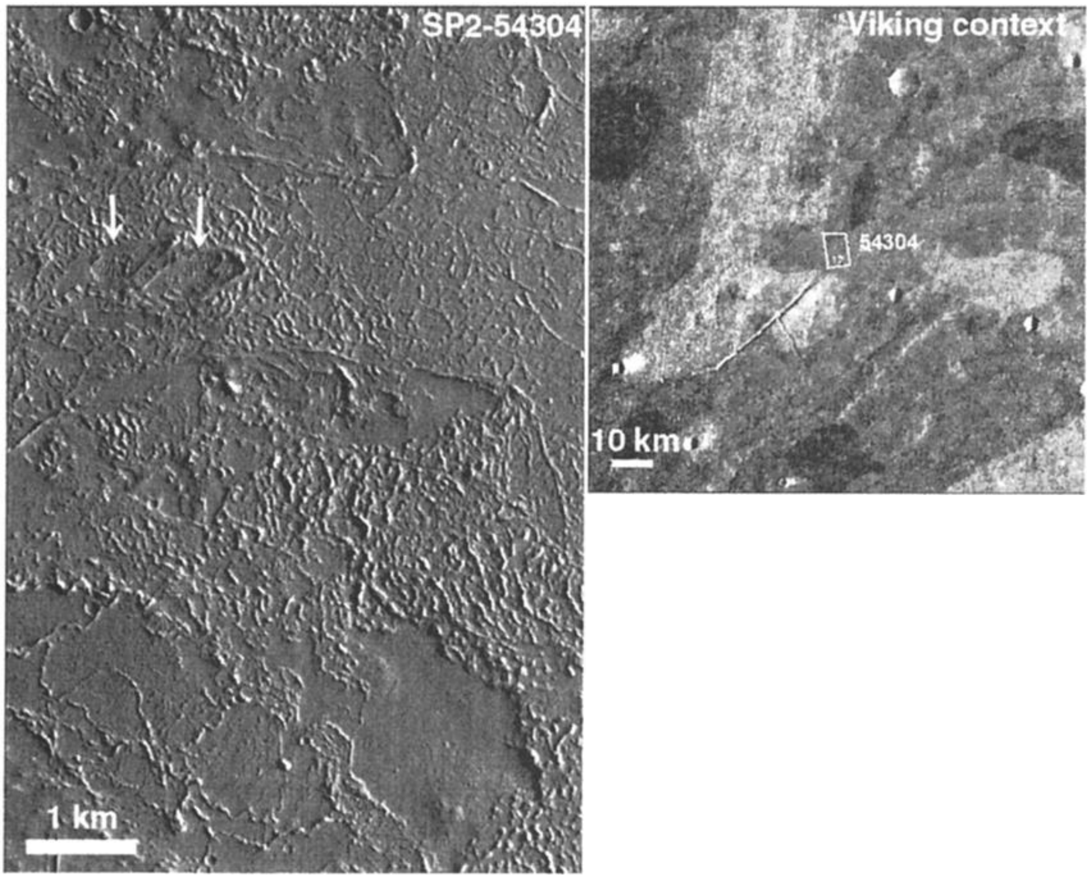


Figure 8. MOC image SP2-54304 and Viking context from lava plains near Tharsis Tholus, centered at 17.0°N, 84.1°W, showing older lava surface with ridges and grooves, similar to those seen on the recent flood lavas in Elysium and Amazonis. Despite significant aeolian cover, including >100-m-long dunes (arrows), these ridges are readily recognized. However, the margins of plates are generally obscured.

historical eruption and the detailed written records of the eyewitness observations of this eruption (e.g., J. Steingrimsson, written communications, July 4, 1783 and November 24, 1788; O. Stephensen, written communication, August 15, 1783) are particularly helpful in understanding the eruptive history and behavior of the lava. (Written eyewitness reports are published in Icelandic by *Einarsson et al.*, 1984 and the account of Jon Steingrimsson is translated into English [*Steingrimsson*, 1998]). Also, while the Laki flow has been carefully examined by several Icelandic workers [e.g., *Jakobsson*, 1979; *Oskarsson et al.*, 1982; *Thordarson*, 1990; *Thordarson and Self*, 1993], little is published on the flow morphology. Much of the following description is based on previously unpublished field work by T. Thordarson.

The eruption started on June 8, 1783, and lasted until February 7, 1784, erupting 15.1 km^3 of basalt (dense rock equivalent (DRE)) that advanced >60 km from the vent [*Thordarson and Self*, 1993; *Thordarson et al.*, 1996] (Figure 11). About 10% of the mass (and >50% of the volume) was deposited as pyroclastics. Of the lava flows, 60% (by mass and volume) was erupted in the first 1.5 months and 90% in the first 5 months [*Thordarson and Self*, 1993]. The average effusion rate for the entire eruption was 730 m^3/s [*Thordarson and Self*, 1993], not much higher than historical Mauna Loa eruptions [*Rowland and Walker*, 1990]. However, based on the eyewitness accounts of the rate of advance of lava surges, effusion rates appear to have reached values as high as 8700 m^3/s for short periods during the early part of the eruption [*Thordarson and Self*, 1993].

Ten eruptive episodes fed fissure segments 1.6-5.1 km in length along a fissure system with a 27 km total length [*Thordarson and Self*, 1993]. Generally, only one fissure segment was active at any given time, but for a short period sections of three fissure segments were active over a total distance of ~10 km [*Thordarson and Self*, 1993]. The eruption released 350 Mt of CO_2, 240 Mt of H_2O, 120 Mt of SO_2, 15 Mt of HF, and 7 Mt of HCl [*Thordarson et al.*, 1996]. The fluorine in particular was deadly to the local livestock (primarily sheep), which led to a massive famine that killed approximately 20% of the population of Iceland [*Thorarinsson*, 1979]. The tops of the buoyant eruption columns penetrated the tropopause, and SO_2 was distributed across the Northern Hemisphere [*Woods*, 1993]. The Laki eruption was associated with an unusually cold winter, with the Danube River freezing over at Vienna and rafted ice filling the Mississippi River at New Orleans [*Thordarson*, 1995; *Thordarson et al.*, 1996].

The lavas flowed away from the fissure system and passed through two river gorges before fanning out on the coastal plains (Figure 11). Field mapping and examination of aerial photographs show that the majority of the lava formed inflated pahoehoe. However, one section of the distal end of

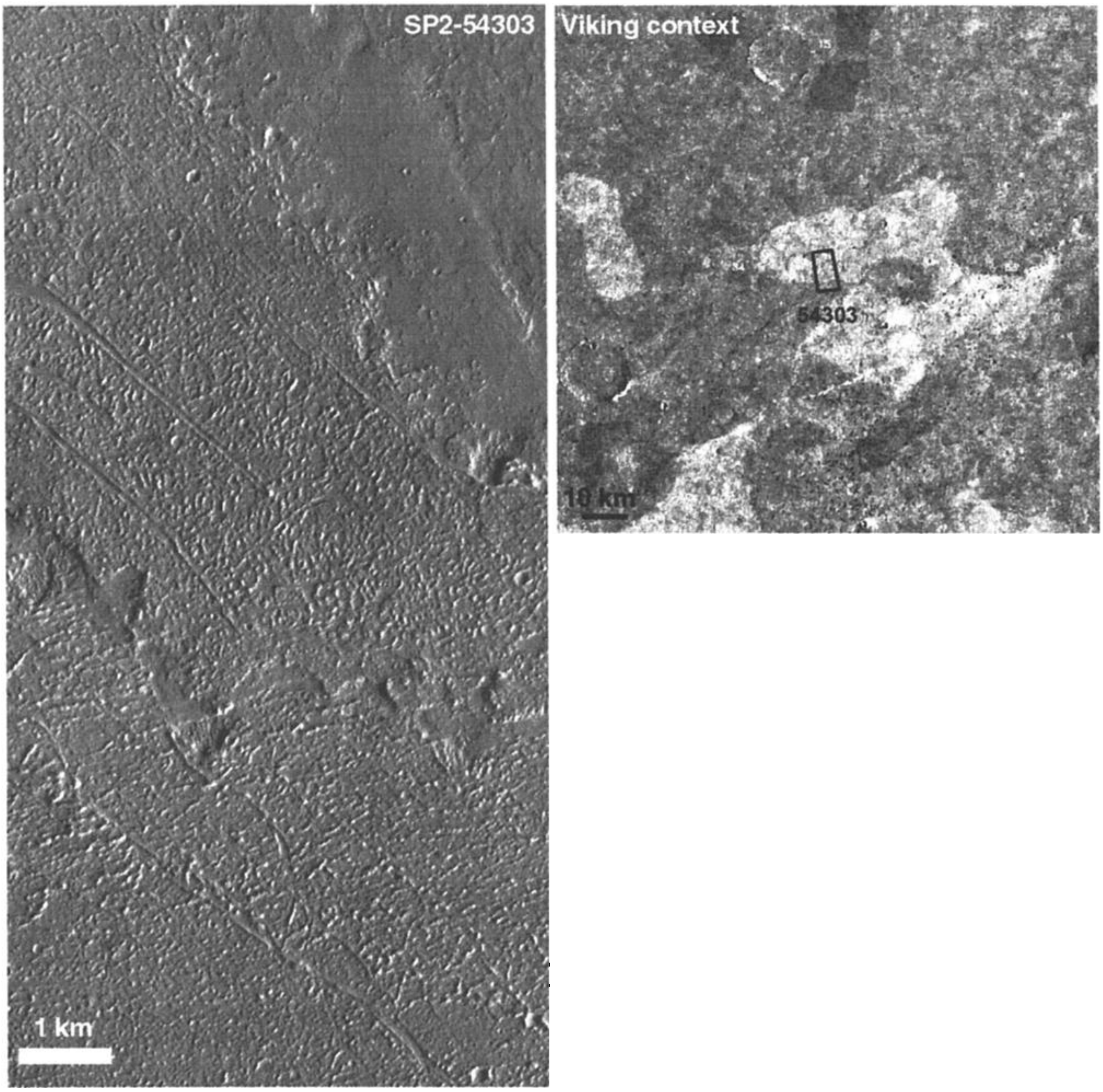
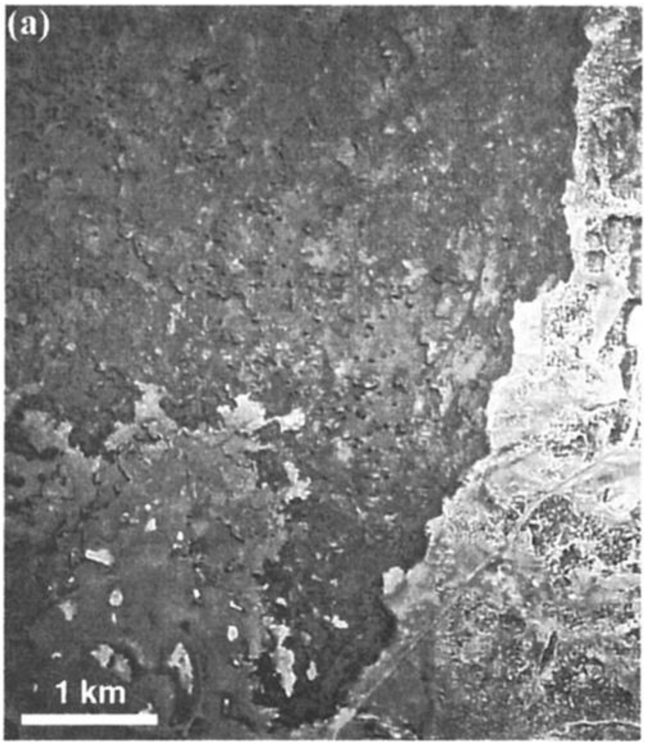


Figure 9. MOC image SP2-54303 and Viking context from lava plains near Tharsis Tholus, centered at 14.2°N, 83.7°W, showing older, degraded example of flood lava with similar ridged morphology as the more recent Elysium and Amazonis examples. Grooves in the crust suggest flow to the NW during the initial emplacement. However, the crack indicates opening to the NE, suggesting that the flood-like breakout from this lobe moved in that direction. Note that in these images, there is no correlation between the lava morphology and the albedo mottling in the Viking image.

the flow field is composed of arcuate pressure ridges, rafted crustal plates, and longitudinal grooves (Figures 12 and 13). The lava in this area is covered by a few centimeters of moss, but the underlying flow top can be seen to be brecciated, forming a lava surface unlike any common examples in Hawaii. The closest Hawaiian analog to this disrupted crust is slab pahoehoe, a form of lava transitional between pahoehoe and aa [*Macdonald*, 1953].

We infer that the rafted crustal plates initially formed on a relatively stagnant flow and were disrupted and transported by a later surge in the flow rate. This is directly analogous to the manner in which the ice cover on a partially frozen river breaks up during spring floods ("undanhlaups" in Icelandic). Recorded observations of the flows advancing through the areas shown in Figures 12 and 13 include reports of the lava surging forward at 2-5 times the average rate for several hours [*Steingrimsson*, 1998].

Based on field work and examination of the aerial photographs (Figures 12 and 13), we conclude that the grooves in the flow surface are wakes left in the brecciated crust as it translated past isolated obstacles. The arcuate ridges appear to be locations where the brecciated crust was compressed as it ponded before overtopping topographic ridges. This part of the Laki flow field is underlain by



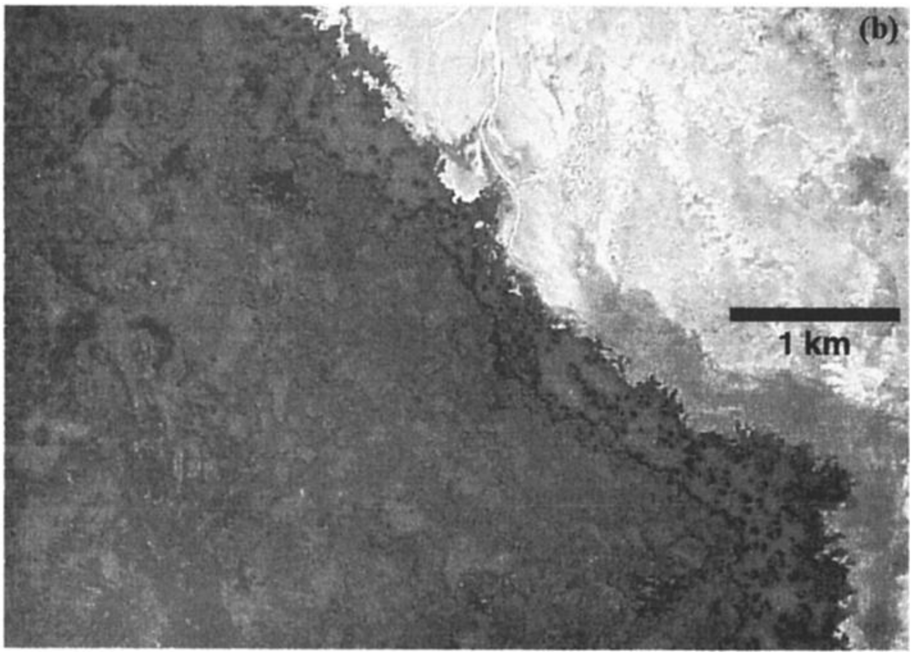


Figure 10. Examples of terrestrial inflated pahoehoe flow surface morphology. (a) Aerial photograph of SE corner of the Macarthy's flow field, New Mexico. Note large inflation plateau in center of image with inflation pits. Smaller inflation features on hummocky pahoehoe surfaces appear as mottling at this resolution. (b) Aerial photograph of SE portion of Carrizozo flow field, New Mexico. Note the large inflation plateau along the flow margin with inflation pits and the more mottled appearance of the hummocky portions. Dark delta-shaped features are small patches of slab pahoehoe breaking out from the inflated pahoehoe lobes. (c) Aerial photograph of SE portion of the Laki flow field, Iceland. Note the large inflation plateau in the center of this lobe and the more hummocky textures. Highway 1 crosses the lava in this image. (d) Oblique aerial photograph looking south over the west end of the Pisgah flow field, California. Note mottled pattern of the hummocky pahoehoe surface made more visible by partial filling with aeolian sands. Trucks on Interstate 40 are shown for scale. Photo courtesy of Bruce C. Murray.

rootless cones in the older (934 A.D.) Eldgja flow field. These cones occur both in lines and in isolated clusters of cones, making them the only likely candidate for the topographic obstacles that the Laki lava encountered. The raised edges and streamlined shape of the grooves suggest that the brecciated top of the lava had a significant yield strength, while pahoehoe breakouts show that the interior remained fluid.

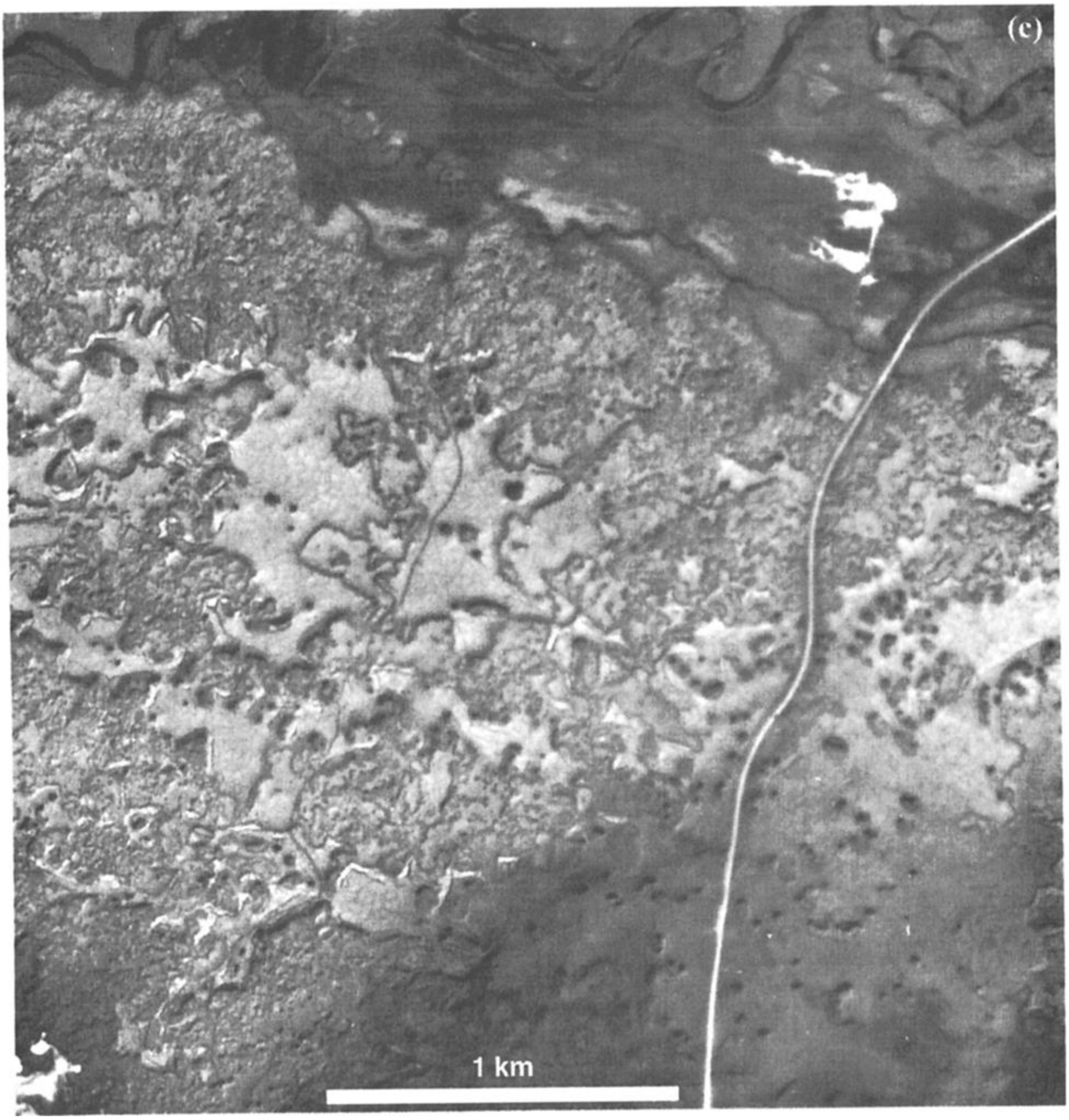


Figure 10. (continued)

This surging and autobrecciating mode of emplacement has not been described in Hawaii, but may apply to the ~20% of the flows in the Columbia River flood basalt province that have rubbly flow tops and smooth pahoehoe-like flow bases [*Self et al.*, 1998]. It may also help explain the enigmatic brecciated flows seen on the Kerguelen Plateau [*Frey et al.* 2000].

In Hawaii, longitudinal grooves and pressure ridges are largely absent within lava flows. However, we have found some pieces of rafted crust that are tens of meters in scale. Rafted pieces of crust are visible along the Southwest Rift Zone of Kilauea and in the summit caldera of Mauna Loa (Figure 14). The rafts are mostly confined to the near-vent facies of lava flows (i.e., the broad sheets of highly vesicular lava within a few hundred meters of the vent that formed before narrow channels or tubes evolved within the flow). These near-vent flows are characterized by high local effusion rates and hot, gas-rich lavas with relatively low viscosities. Where the flows have ponded, their surfaces are also characterized by very shallow slopes ($<<1°$).

3. Modeling

The observations from Iceland, and to a lesser extent, from Hawaii and continental flood basalt provinces, strongly suggest that the Martian flood lavas were emplaced as sheet flows with intermittent, surging advances. In this section we describe this qualitative emplacement model in more detail and then attempt to place quantitative constraints on eruption parameters. Surging emplacement is inherently difficult to

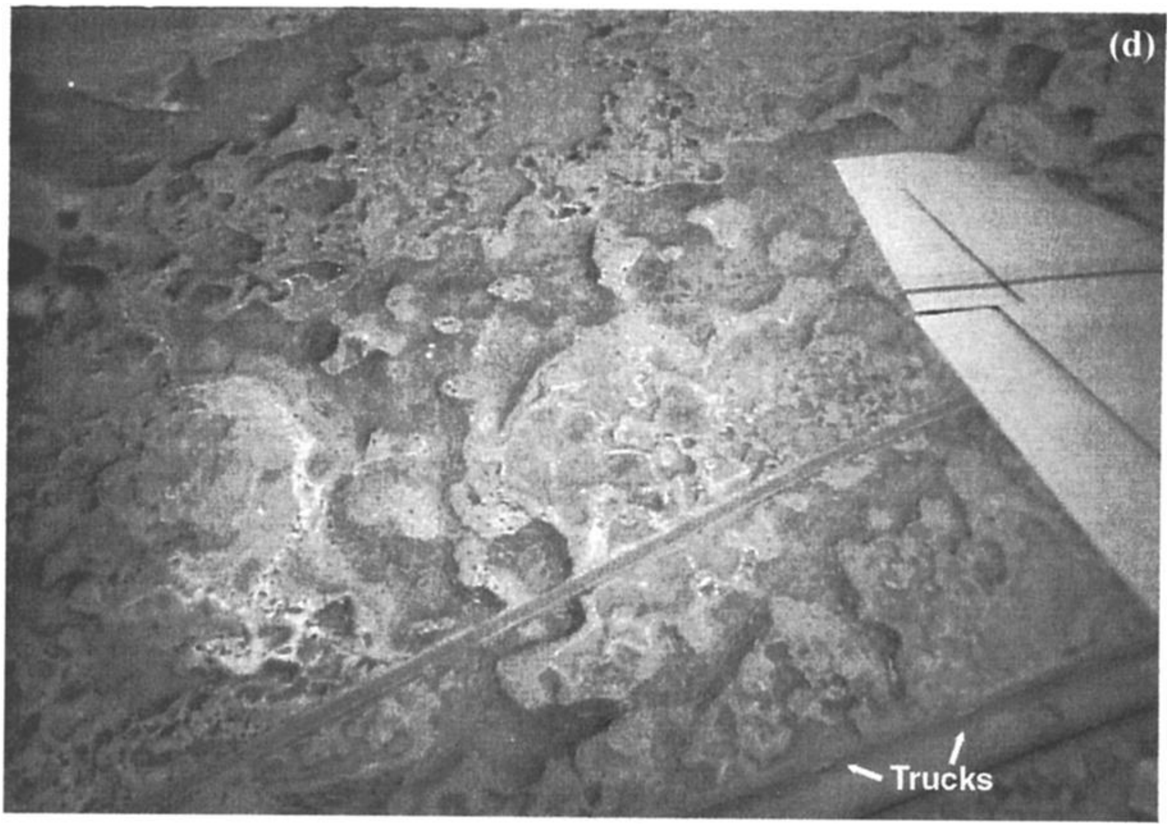


Figure 10. (continued)

model and we lack critical measurements such as the length and width of individual outpourings. Therefore it is currently impossible to produce a quantitative model replicating the emplacement of a specific Martian flood lava flow. However, it is possible to show how the flows were not emplaced. Elimination of the physically impossible provides surprisingly useful constraints on the formation of the flows.

3.1. Qualitative Model

The broken rafts of crust and the observations of the Laki flow field suggest that the Martian flood lavas have disrupted flow tops. The observed rafts of crust are highly suggestive of surges traveling through the flow, inflating it to the point that a sudden breakout carries away large pieces of previously solidified crust (Figure 15). The crust riding the flood is broken up as it is sheared along the margins of the flow and as the flow moves between and around topographic obstacles. As the surge wanes, the base of the disrupted upper crust may become coherent via solidification of the liquid lava matrix, allowing inflation to resume. While there is some evidence for the formation of marginal levees, based on the images available to date, the flow does not seem to be restricted to narrow channels. The disrupted crust on these open sheet flows should have experienced a steady process of production and destruction as it is churned by the flow past bends, constrictions, and other types of obstacles.

3.2. Quantitative Thermal Model

The surging flow model implies that the lava was transported in two very different regimes. The first flow regime involved flow under a thick, stationary, insulating crust. The second involved open sheet flow with a disrupted and mobile crust. The thermal models from *Keszthelyi and Self* [1998b] are readily modified to Martian conditions and can describe each of these two distinct flow regimes.

3.2.1. Insulated sheet flow. The thermal model for insulated sheet flows presented by *Keszthelyi and Self* [1998b] is a simple modification of the thermal model for lava tubes presented by *Keszthelyi* [1995]. Both models involve steady state conductive heat loss through the upper crust and heat generation via steady state viscous dissipation. Flow velocities within the sheet flow are calculated using the equation for flow between two parallel plates and the assumptions of laminar flow, a Newtonian rheology, and shallow slopes. The resulting set of equations is given by *Keszthelyi and Self* [1998b] as

$$\langle v \rangle = \rho\, g\, \theta\, H^2 / 8\, \eta, \tag{1}$$

$$Q_{visc} = \rho\, g\, H \langle v \rangle\, \theta\,, \tag{2}$$

$$Q_{cond} = k\,(T - T_a) / Hc, \tag{3}$$

$$\partial T/\partial x = (Q_{visc} - Q_{cond}) / (\langle v \rangle\, \rho\, C_p^*\, H), \tag{4}$$

where $\langle v \rangle$ is the average flow velocity, ρ is the bulk lava density, g is the gravitational acceleration, θ is slope, H is the thickness of the liquid lava, η is viscosity, Q_{visc} is the heat generated by viscous dissipation, Q_{cond} is the heat loss by conduction, k is thermal conductivity, T is the temperature of the liquid lava, T_a is the temperature of flow surface (assumed to be ambient), Hc is the thickness of the solid crust, $\partial T/\partial x$ is the cooling per unit distance of flow, and C_p^* is heat capacity including the effect of latent heat (dominated by the crystallization of olivine). This model is very crude, and results should be considered accurate to no better than a factor

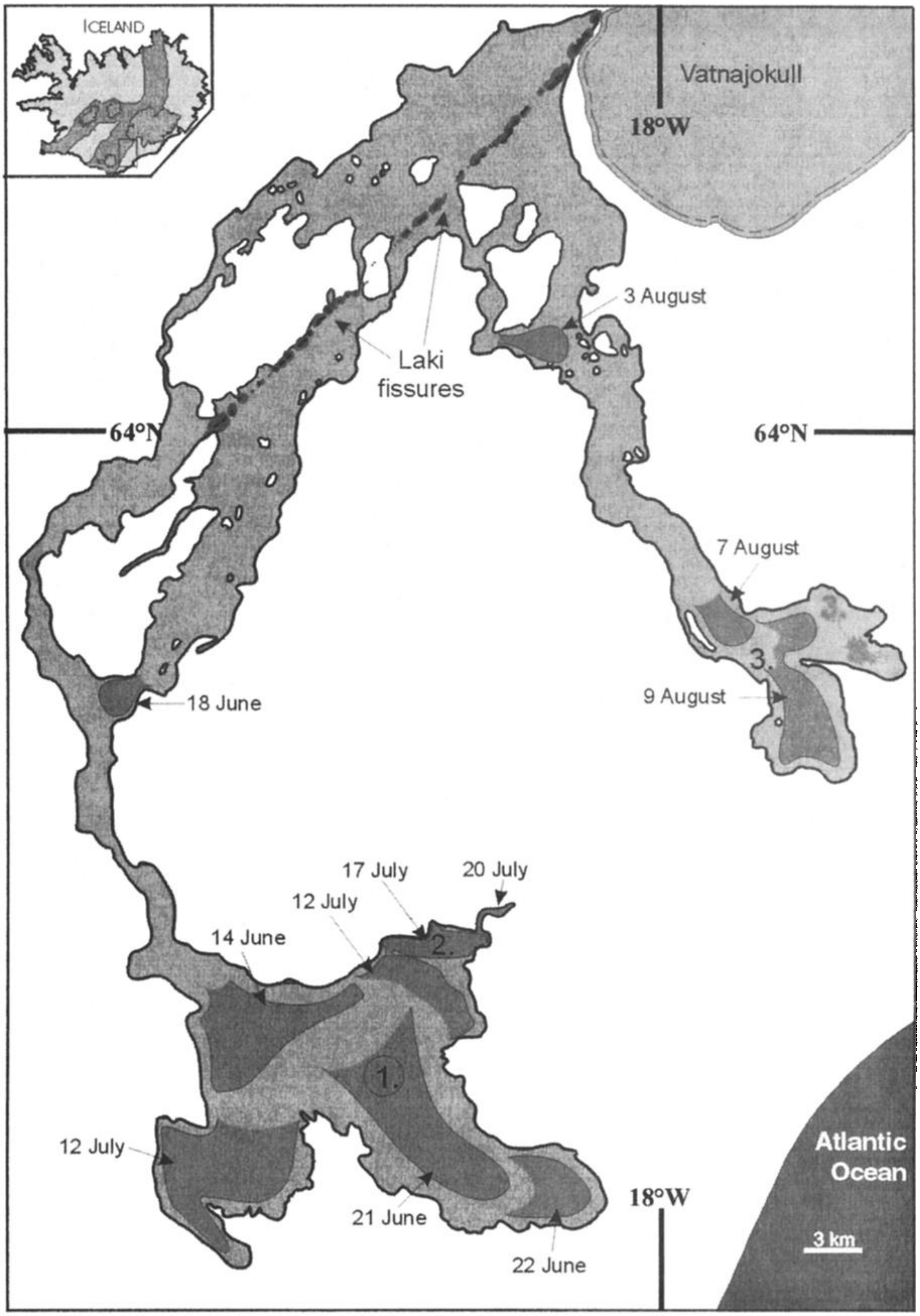


Figure 11. Map of Laki flow field, Iceland. Positions of the flow front at known dates are shown. Large, flood-like breakout with ridged surface and rafted plates (Figure 13) is indicated by location 1. Smaller flood-like breakout with rafted plates (Figure 12) is shown by location 2. Classic inflated pahoehoe structures (Figure 10c) are shown by location 3.

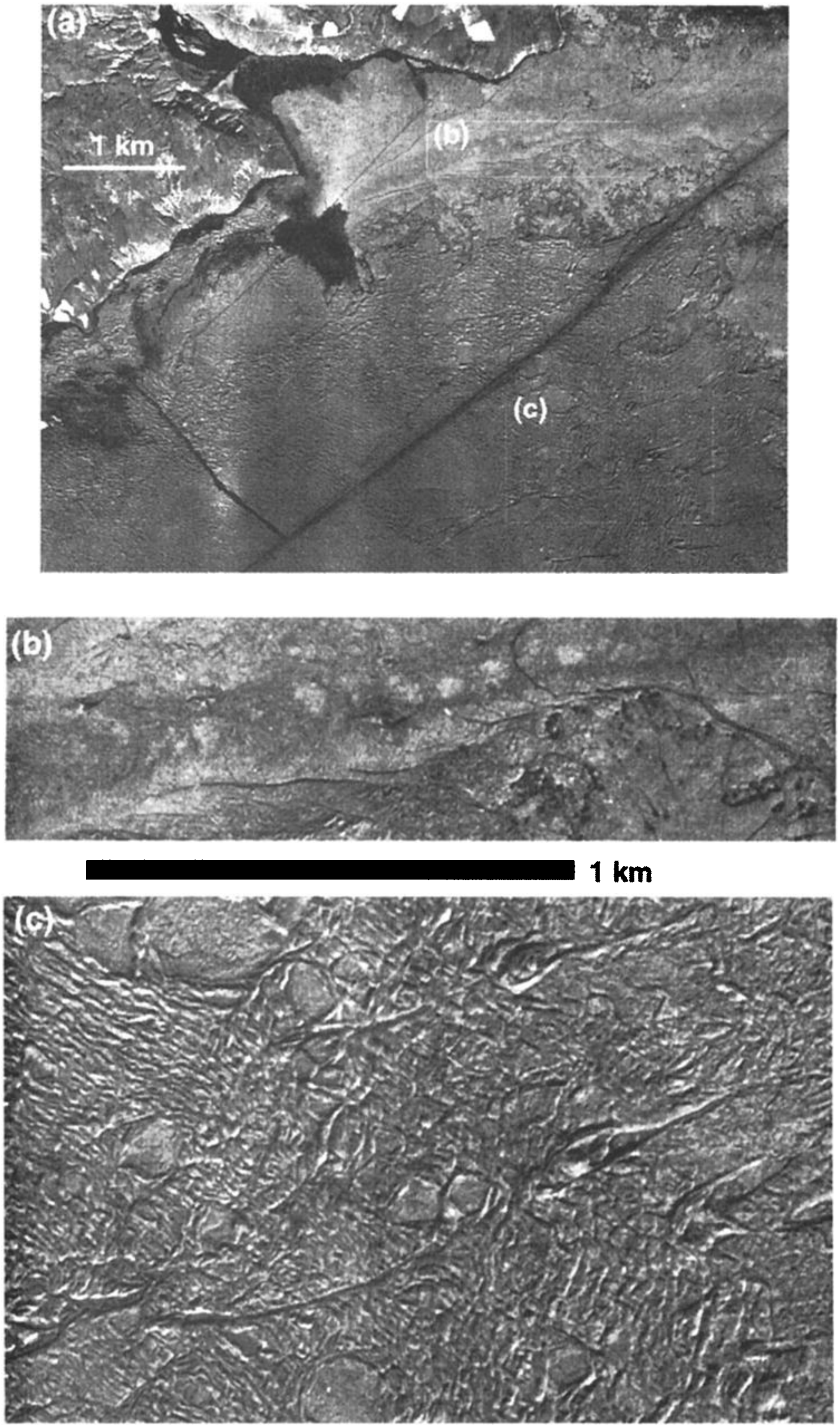


Figure 12. Aerial photograph of portion of the Laki flow field, Iceland (labeled 2 in Figure 11). (a) Context showing the Skafta River, Highway 1, and farms as well as the locations of Figures 12b and 12c. (b) Close-up of the northern channel which was active on July 17, 1783, showing 100-m-scale rafts of lava crust transported in an open channel. This channel broke out from a lava front that had stagnated around July 12, 1783. (c) Close-up of this earlier lava, showing that it formed a broad sheet with pressure ridges, plates, and grooves. The ridges are piles of breccia, the plates are pieces of more intact crust, and the grooves are "wakes" left in the brecciated top as it moved past topographic obstacles.

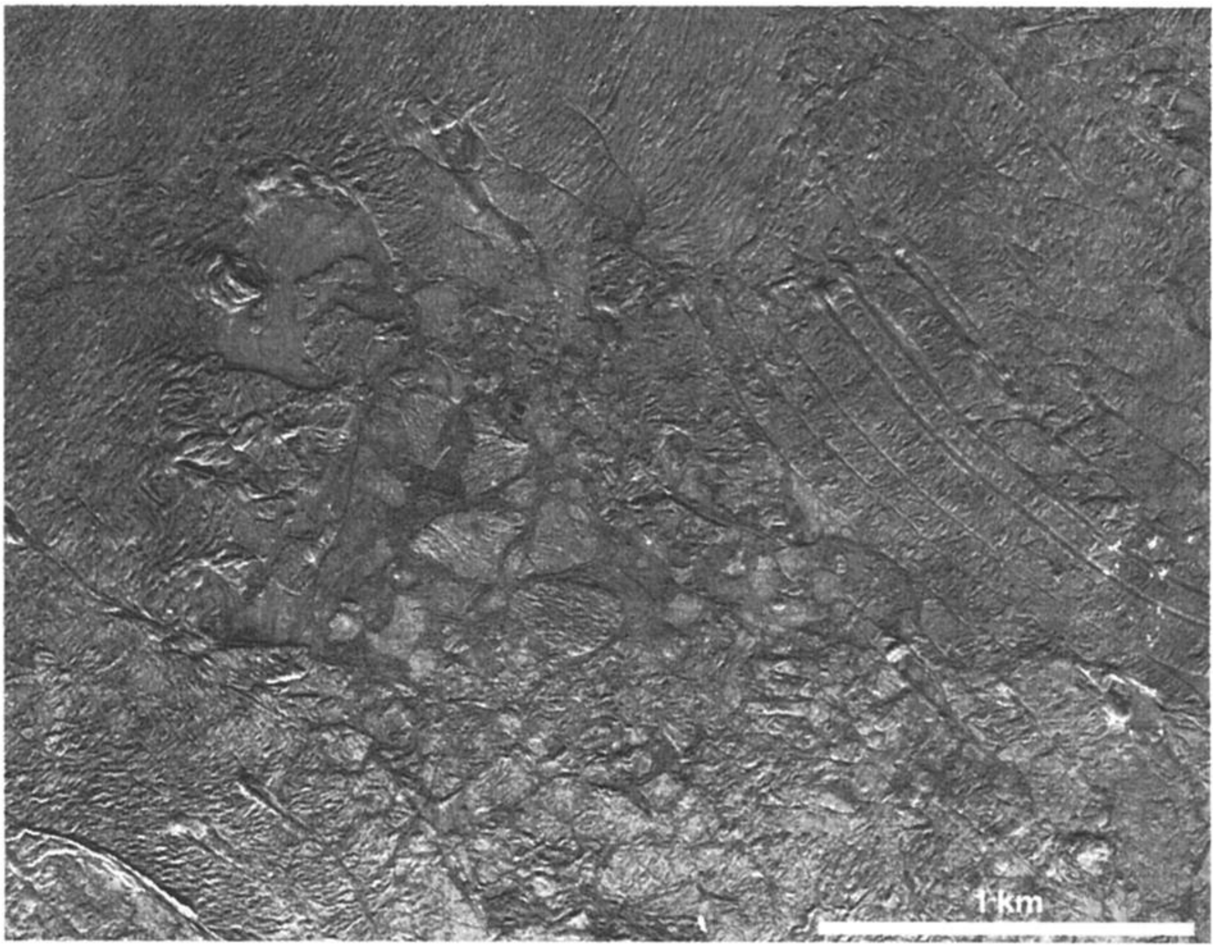


Figure 13. Aerial photograph of portion of the Laki flow field, Iceland (labeled 1 in Figure 11), showing part of the large flood-like breakout that formed in mid-June, 1783. Note the large rafts of ridged crust torn from the earlier formed flow top, as well as long grooves in the ridged lava. The ridged lava surface consists of pressure ridges in the brecciated flow top, not tumuli as suggested by *Theilig and Greeley* [1986]. The grooves formed as this brecciated flow top translated over topographic obstacles.

of 2. However, more sophisticated numerical models [e.g., *Sakimoto and Zuber*, 1998] do not produce significantly different results, only more precise ones.

The major modifications required to adjust the model presented by *Keszthelyi and Self* [1998b] to the Martian flood lavas are the reduction in both slope and gravity. Ambient temperature is also reduced to –30°C, but this parameter has very little effect on the model results. Also important is that the Martian flood lavas appear to have advanced as much as 2000 km from the presumed vent areas, while *Keszthelyi and Self* [1998b] only examined the requirements for a flow to extend ~100 km.

In examining the Martian flood lavas, we chose to investigate the effect of lava viscosity and the thickness of the insulating crust on the minimum flow thickness and flow velocities needed for a flow to travel 1000 km without freezing. As in the work of *Keszthelyi and Self* [1998b] we assume that a temperature drop of 50°C within the lava transport system would lead to significant crystallization and the flow stopping. Thus the maximum cooling rate allowable for a 1000-km-long flow would be 0.05°C/km. Table 1 lists the model runs and results for the Mars Observer Laser Altimeter (MOLA) derived slope of 0.025%. Viscosities of 10, 100, and 1000 Pa s for the fluid lava are considered, roughly corresponding to typical values for ultramafic, basaltic, and basaltic-andesite flows.

Several points in Table 1 require some discussion. First, irrespective of the chosen viscosity, a flow that can extend 1000 km under a 1-m-thick crust will be turbulent because of the high flow rates required. Because it is difficult to envision a stationary insulating crust surviving over a turbulent flow, we regard this scenario as unlikely. In the case of the ultramafic (10 Pa s) lavas, the crust must be ~20 m thick before flow velocities drop into the laminar regime. While the ~25 m total thickness required by this scenario may be compatible with some observed flow thicknesses, it suggests that it might not be possible to produce insulating sheet flows with extremely fluid lavas, even with low gravity and remarkably shallow slopes. This argues against ultramafic compositions for the Martian flood lavas. However, we note that recent studies suggest that mildly turbulent terrestrial komatiite lava flows may have been emplaced as insulated pahoehoe sheet flows [*Hill et al.*, 1995].

For a viscosity of 1000 Pa s, the minimum model flow thicknesses are of the order of 60 m. This already exceeds the measured flow thicknesses, so we can confidently rule out more evolved and viscous compositions (e.g., andesites) for these flood lavas. We conclude that physical properties

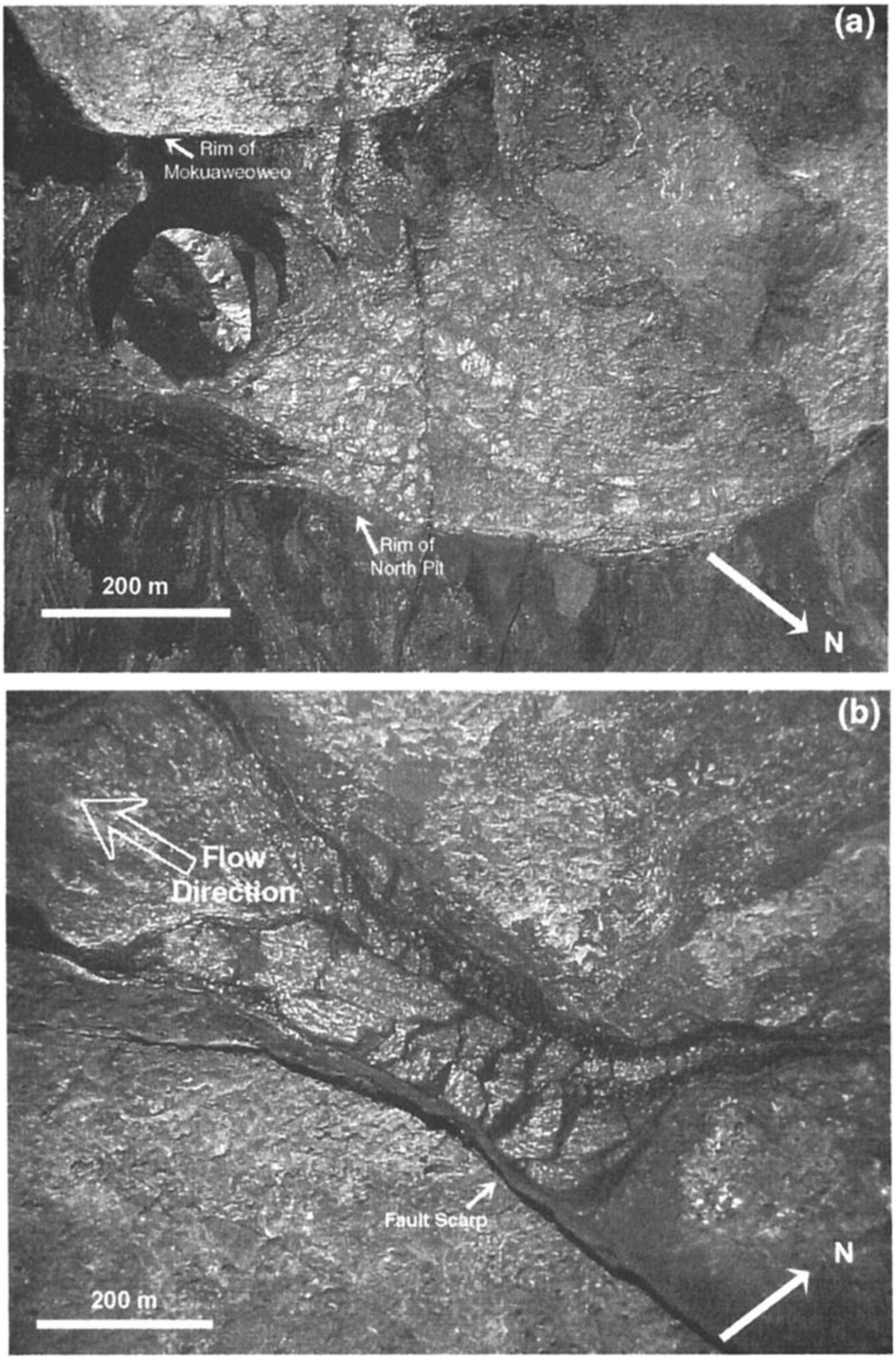


Figure 14. Aerial photographs of platy lava in Hawaii. (a) Ponded flows at the northern end of Mauna Loa's summit caldera (Mokuaweoweo) and in North Pit. In this case, there is no clear direction of flow recorded in the orientations and shapes of the crustal rafts. (b) The 1974 Kilauea Southwest Rift Zone open channel flow. Lava ponded against the cliff face and formed a solid crust before the flow surged onward. Both these areas are characterized by high flow rates and shallow slopes.

similar to terrestrial basalts are probably the most appropriate for further modeling of the Martian flood lavas.

Since the model constrains both flow velocity and flow thickness, we can also make quite robust conclusions about minimum flow rates per unit width of flow. This must be at least 1-5 m^2/s for the lava to be transported >1000 km without freezing. The available images suggest that the sheets are generally on the order of 10 km wide, requiring average effusion rates on the order of $1\text{-}5\times10^4$ m^3/s or greater. This is approximately 1 order of magnitude higher than estimated for

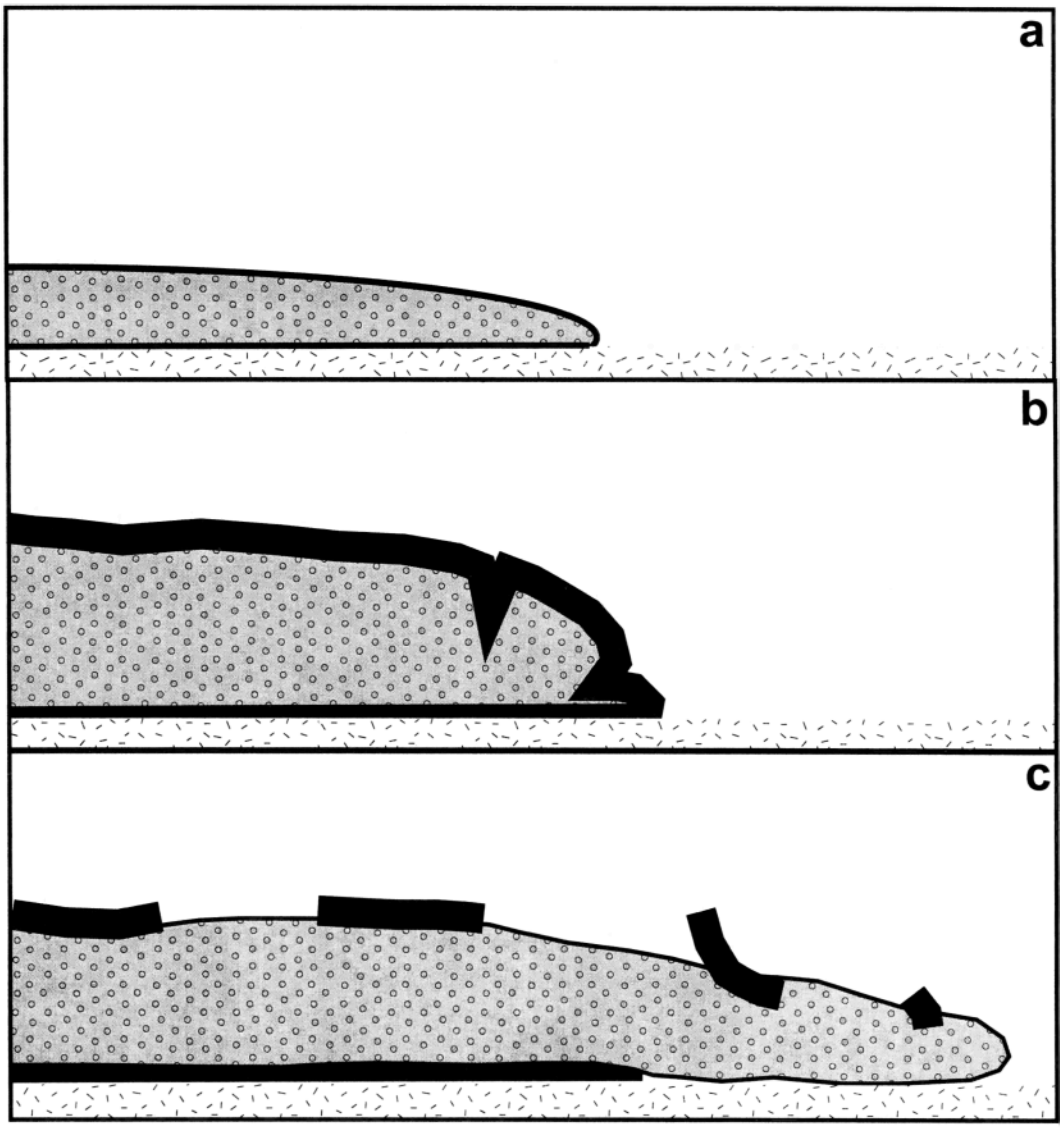


Figure 15. Cartoon of formation of rafted plates on Martian flood lavas. (a) Lava enters the area, most likely as a rapid flood. (b) The flow front stagnates and a large volume of liquid lava is stored within the confines of a thickening, insulating crust. Lava is fed into the area under a thick insulating crust. (c) The flow front fails and lava breaks out in a runaway flood, rafting away pieces of crust. The sequence is repeated as the flood wanes and slows, and an insulating crust begins to grow again.

Table 1. Results From Insulated Sheet Flow Model

Viscosity (Composition) Pa s	Crust Thickness m	Total Flow Thickness m	Average Flow Velocity m/s
10	1	(turbulent)	
(ultramafic)	5	(turbulent)	
	10	(turbulent)	
	20	24	0.25
100	1	(turbulent)	
(basaltic)	5	27	0.75
	10	26	0.39
	20	32	0.21
1000	1	(turbulent)	
(basaltic andesite)	5	70	
	10	56	0.33
	20	53	0.18

Insulated sheet flow not considered physically possible with turbulent flow.

terrestrial flood basalt eruptions [*Thordarson*, 1995; *Self et al.*, 1997, 1998; *Thordarson and Self*, 1998].

As an aside, the *Keszthelyi* [1995] thermal model for lava tubes indicates that 1000 km long flows could be formed on the Martian plains with an effusion rate of only 33 m^3/s and a tube diameter of 22.7 m. However, since we have not found any evidence for narrow lava tubes in the Martian flood lavas, and because such lava tubes are not expected on these shallow slopes [*Keszthelyi and Self*, 1998b], we find it very unlikely that the Martian flood lavas seen in the recent MOC images were emplaced at low effusion rates.

3.2.2. Open sheet flow. In order to quantify the more rapid, surging, stage of our hypothesized emplacement model for the Martian flood lavas, we require a different set of thermal models. *Crisp and Baloga* [1994] developed a model for the open channel aa flows during the 1984 Mauna Loa eruption that is generally applicable to the open sheet flows we postulate. The *Crisp and Baloga* [1994] model calculates the cooling within the well-mixed core of the flow, including the effects of reduction in radiative cooling by a disrupted crust, mixing of cold crust back into the flow, and the latent heat of crystallization. *Keszthelyi and Self* [1998a,b] presented a minor modification of the *Crisp and Baloga* [1994] model to convert the inputs to parameters that are somewhat easier to constrain for flows that are not observed while active. In particular, the increase in crystallinity is calculated from the cooling rather than using petrological data from samples collected from the active channel. Atmospheric convective cooling and viscous heating were also added, with flow rates calculated assuming a Newtonian rheology while the flow is laminar and using the formulas presented by *Goncharov* [1964] when the flow is turbulent. Equations (5)-(10) describe the *Keszthelyi and Self* [1998a,b] modification of the *Crisp and Baloga* [1994] model:

$$\partial T/\partial x = \{Q_{visc}-Q_{rad}-Q_{atm}-Q_{entr}\}/\{\langle v\rangle\ \rho\ C_p{*}\ H\}, \tag{5}$$

where T is the core temperature, x is the down flow direction, Q_{visc} is the viscous heating, Q_{rad} is radiative heat losses, Q_{atm} is atmospheric heat loss, Q_{entr} is the heat lost from the core of the flow via entrainment of the crust, $\langle v\rangle$ is the average flow velocity, ρ is lava density, $C_p{*}$ is heat capacity including latent heat, and H is the flow depth. Viscous heating is described by equation (2), and radiative heat losses can be expressed as

$$Q_{rad} = \varepsilon\sigma f(T^4 - T_a^4), \tag{6}$$

where ε is emissivity (~0.95), σ is the Stefan-Boltzmann factor, f is the fraction of core exposed, and T_a is the ambient temperature [*Crisp and Baloga*, 1990]. The atmospheric heat loss has a similar expression,

$$Q_{atm} = \mathbf{h}\ f(T - T_a), \tag{7}$$

where $\mathbf{h}$ is the atmospheric heat transfer coefficient, which has an approximate value of 70 W m^{-2} K^{-1} on the Earth [*Keszthelyi and Denlinger*, 1996]. This value should be directly proportional to atmospheric density and heat capacity [e.g., *Arya*, 1988], suggesting a value of ~1 for Mars. Heat transfer due to entrainment of the crust is

$$Q_{entr} = \rho\ C_p{*}\ H_c\ (T - T_c)\ /\ \tau, \tag{8}$$

where H_c is the thickness of the crust, T_c is the average temperature of the crust, and τ is the average time a piece of crust survives before being entrained back into the core of the flow.

The sheet flow is assumed to be laminar while the Reynolds number (Re) is less than 500. We note that the value of Re marking the transition to turbulent flow depends on both the exact version of the Reynolds number used and the geometry of the flow. The various permutations of Re and the critical Re that permeate the published literature were discussed by *Keszthelyi and Self* [1998b]. In this paper we use

$$Re = \rho\ r_h \langle v\rangle\ /\ \eta, \tag{9}$$

where r_h is the length scale of the largest physical perturbation that can propagate down the length of the flow (i.e., the flow thickness for a sheet), $\langle v\rangle$ is the average velocity, and η is viscosity.

While laminar, average flow velocity for Newtonian sheet flow is given by [e.g., *Bird et al.*, 1960]

$$\langle v\rangle = \rho\ g\ \theta\ H^2\ /\ 3\ \eta. \tag{10}$$

There are many different formulations for turbulent flow, but *Keszthelyi and Self* [1998b] explained why the one presented by *Goncharov* [1964] appears to be the most applicable to lava flows. Thus for moderately turbulent flow we use

$$\langle v\rangle^2 = g\ H\ \theta\ /\ C_f, \tag{11a}$$
$$C_f = \lambda/2, \tag{11b}$$
$$\lambda = \{4\ \log_{10}[6.15((Re'+800)/41)^{0.92}]\}^{-2}, \tag{11c}$$
$$Re' = 2Re, \tag{11d}$$

where C_f is a friction factor and λ is an empirically derived fit to experiments of fluid flow that are moderately turbulent [*Goncharov*, 1964].

Combining these equations, the *Keszthelyi and Self* [1998a,b] model calculates the flow rate and cooling rate for the core of a sheet or wide channel flow with a disrupted crust. However, it must be noted that the model does have several critical shortcomings when applied to the Martian flood lavas. First, the pulsating nature of the proposed model is not incorporated into this simple model. However, by examining the two flow regimes separately, we are able to make some reasonable inferences about the overall flow emplacement. Second, even if the liquid lava is approximately Newtonian, a thick brecciated crust can impart the bulk flow with a significant yield strength [e.g., *Booth and Self*, 1973; *Hulme*, 1974; *Fink and Zimbelman*, 1986]. However, inputting a reasonable viscosity of 500 Pa s for the Laki lava, our flow model produces flow velocities (1-4 m/s) and eruption rates (1500-9000 m^3/s) that are consistent with the eyewitness observations. This suggests that the crust riding on this type of flood might have a relatively minor effect on the bulk flow rheology. Third, the model cannot be used to examine the temporal and spatial evolution of the flow. Still, the success of the original *Crisp and Baloga* [1994] model suggests that this modified version should be more than adequate to provide broad constraints on the eruption parameters. More rigorous modeling of the flow behavior, as in the work by *Glaze and Baloga* [1998], will become applicable only when individual flows and breakouts can be mapped out.

The key inputs into this model are the thermophysical properties of the lava (initial temperature, bulk viscosity, heat capacity including latent heat and bulk density) and the crust (average temperature, survival timescale, and thickness). Slope is the single most important factor [*Keszthelyi and Self*, 1998a], but is well constrained for the most recent flood lavas by the new MOLA data at about 0.025% [*Smith et al.*, 1999]. The thermo-physical properties are related to the assumed composition of the lavas. We have determined that the open sheet flow model does not help further constrain the thermophysical properties of the lava and thus only discuss runs using the basaltic composition suggested by the insulated sheet model.

The properties of the crust are poorly constrained. The only published observational constraints on the crust parameters are from *Crisp and Baloga* [1994] for the 1984 Mauna Loa open channel aa flow. After examining a range of possible crustal parameters, we have chosen to report just three examples which illustrate the main results. We call

Table 2. Crustal Properties for Open Sheet Flow Model

Crust Type	τ, s	f, %	Hc, m	T-Tc, °C
Thin	100	50	0.05	100
Medium	1000	10	1	500
Thick	100,000	1	10	700

these three cases "thin" crust, "medium" crust, and "thick" crust. The parameters used for each case are listed in Table 2 but require some justification. In general, we assume that a thinner crust will be both hotter and more rapidly destroyed. In the "thin" crust case, we examine the possibility that the crust is only centimeters thick. Such a thin crust would be expected to be disrupted and reincorporated into the flow in a matter of minutes. The extensively disrupted crust would also expose a relatively large portion of the interior of the flow. The values listed in Table 2 for the thin crust case are for a crust marginally thinner, hotter, and more disrupted than that reported by *Crisp and Baloga* [1994] for the early part of the 1984 Mauna Loa eruption. The "medium" crust case examines parameters related to a crust approximately a meter thick. Crustal blocks might survive for tens or hundreds of minutes and would be substantially cooler than the liquid interior of the flow. The "thick" crust is taken to be about 10 m thick with blocks surviving for roughly a day. The thick crust case is likely to be the best representation of the rafting lava on Mars, where pieces of crust are strong enough to form rafts kilometers across. While these three cases do not cover the full range of plausible crust properties, they do provide important insight into the dynamics of the surging flood lavas.

Despite the shortcomings of the model noted earlier, several points are clearly demonstrated by the model results shown in Table 3. First, we can rule out certain types of flow behavior with a high degree of confidence. If the disrupted crust is vigorously broken up and mixed into the interior of the flow, the flow will be remarkably thermally inefficient. In fact, the "medium" thickness crust, with a survival timescale of 15 minutes, cools almost an order of magnitude faster than the "thin" crust that loses heat via essentially uninhibited thermal radiation. This counterintuitive result is actually easily explained. The crust riding the flood-like breakout is the product of days or weeks of cooling. When this material is mixed into the interior of the flow, the cumulative heat loss from the many days of cooling is suddenly transported into the previously insulated flow interior. Not even wholesale radiative heat loss from the surface of the flow can match this rate of cooling.

The modeling suggests that in this most thermally inefficient mode, the flood should not have been able to advance more than 1-12 km before losing >50°C from the interior of the flow and being brought to a halt by the resulting crystallization. The fact that we see the sheet lobes with rafted pieces of crust extending for tens of kilometers strongly indicates that this thermally inefficient style of emplacement was not the norm for the Martian flood lavas. This rapid mixing of the disrupted crust and freezing of the lava would be most likely if the flow was turbulent, again arguing against very fluid ultramafic lavas.

The model runs with the "thick" crust are our current best attempt to mimic the undanhlaup-like phase of the Martian flood lavas. Even though the thick crust case has a crust that is stable on the timescale of a day, the occasional entrainment of blocks of crust leads to substantial cooling. In fact, to extend more than a few tens of kilometers before cooling >50°C, the flow must be >20 m thick. At such a flow thickness the calculated flow velocities will be of the order of a few meters per second, and the local flow rates will be >40 m^2/s per unit flow width. For sheets ~10 km wide this translated to minimum local volumetric flow rates of ~4 x 10^5 m^3/s. If individual flood-like breakouts traveled closer to 100 km, the volumetric flow rates are likely to be ~10^6 m^3/s and flow thicknesses had to be >40 m. However, it is likely that the final flow thickness dropped as the flood stage of the emplacement waned, explaining why >40-m-tall flow margins have not been observed in this region.

4. Conclusions and Discussion

Based on comparison with the morphology of large terrestrial lava flows and thermal modeling, we conclude that the recent Martian flood lavas are broad sheet flows that were emplaced by alternating between two flow regimes: insulated sheet flow and flood-like breakouts. The closest terrestrial analog we have identified is part of the 1783-1784 Laki lava flow in Iceland that was subjected to large surges in lava influx.

Modeling results suggest that the flows should have thermo-physical properties close to basalts. More evolved compositions (i.e., andesites) would require flow thicknesses greater than those observed. Ultramafic lavas are not likely, because they would require the formation of a stable, insulating crust over a turbulent flow.

Average eruption rates for basaltic Martian flood lavas were probably ~10^4 m^3/s, about 1 order of magnitude higher than those estimated for typical terrestrial flood basalt eruptions. However, the large surges that rafted away pieces of crust are likely to have involved short-lived volumetric flow rates 1 to 2 orders of magnitude higher (10^5 – 10^6 m^3/s). This style of emplacement is significantly different from that inferred for the majority of terrestrial flood basalts. However, if typical Martian flood lava eruptions produce 10^3-10^4 km^3 of lava at ~10^4 m^3/s, the eruption durations are expected to be of the order of a year to a decade, very similar to those estimated for some terrestrial flood basalt eruptions [e.g., *Thordarson*, 1995; *Thordarson and Self*, 1998].

These conclusions naturally raise a host of new questions. Perhaps the most obvious is, Why are the styles of eruption of flood lavas different on the Earth and Mars? We speculate

Table 3. Results From Open Sheet Flow Model

Flow Thickness m	Flow Velocity m/s	Flow Rate m^2/s	Crust Type	Cooling Rate °C/km
			"thin"	71
5	0.19	9.5	"medium"	520
			"thick"	73
			"thin"	8.8
10	0.78	7.8	"medium"	65
			"thick"	9.1
			"thin"	1.8
20	1.8	36	"medium"	14
			"thick"	1.9
			"thin"	0.57
40	3.0	120	"medium"	4.2
			"thick"	0.58

that this may be the result of the thicker lithosphere on Mars. *Wilson and Head* [1983, 1994] have shown that larger volumes of magma must accumulate to break through the thick lithosphere on Mars. The resulting dikes should be wider and produce higher eruption rates. The gas phase in the propagating dike is expected to be concentrated in the upper part of the dike [e.g., *Spence et al.*, 1987], leading to an initial vigorous fountaining phase followed quickly by a waning phase as the volatiles are rapidly expended. Each phase of extending the fissure system could lead to a pulse in the flux of lava traveling through the flow, as in the early part of the Laki eruption. Such cycles are common for terrestrial basaltic eruptions [*Wadge*, 1981], but might have been even more dramatic on Mars.

The hypothesis that the Martian flood lavas were fed by fissure eruptions with repeated very vigorous phases leads to predictions that may be testable with current and future missions to Mars. The hypothesized style of eruption should produce very little in the form of vent structures but should result in extensive pyroclastic deposits. The lava flows without extensive vent structures seen around Cerberus Rupes [*McEwen et al.*, 1999c] are consistent with our hypothesized style of eruption.

The prediction of large volumes of pyroclastics is also testable. If we assume that the products of the lava fountains on Mars are similar to those of high basaltic lava fountains on the Earth, we estimate that 10% of the mass of the erupted lava will be in the form of pyroclastics [*Thordarson and Self*, 1993]. The pyroclasts are expected to include a wide range of materials including larger droplets (Pele's tears), open framework foam (reticulite), and fine ash [e.g., *Mangan and Cashman*, 1996]. These materials are typically 10-100 times less dense than the lava flows. This suggests that the total volume of pyroclastics should be approximately 1-10 times the volume of the lava, implying about 10^6-10^7 km^3 of recent pyroclastic deposits.

It is likely that much of this pyroclastic material is preserved on the surface of Mars. The denser fraction, including the Pele's tears and pieces of spatter, should be concentrated nearest to the Elysium and Amazonis Planitia regions, while the extremely porous (>>99% void space) reticulite would have traveled farther, perhaps reaching the base of the Tharsis volcanoes. The fine-grained ash might have been transported globally.

The Medusae Fossae Formation (MFF) is characterized by a weakly lithified material that is readily carved by the wind [*Ward*, 1979; *Scott and Tanaka*, 1982, 1986]. The MFF is mostly located between the radar bright fresh flows of the Cerberus Plains and the "stealth" region [*Butler*, 1994; *Edgett et al.*, 1997] (see Figure 1). A small degree of welding or alteration of basaltic pyroclastic deposits may provide the appropriate degree of lithification for the MFF.

It is intriguing that our estimated volume of pyroclastics is similar to the estimated volume of the Medusae Fossae Formation [*Sakimoto et al.*, 1999]. The MFF partially overlaps the stealth region, but the stealth region is generally farther to the east [*Muhleman et al.*, 1991; *Butler*, 1994; *Harmon et al.*, 1999]. The stealth region is of the order of 10^6 km^2 and has a radar- absorbing cover at least several meters thick [*Butler*, 1994]. Extremely porous reticulite is a prime candidate for this surface cover, but in situ sampling will probably be necessary to confirm this hypothesis [*Butler*, 1994; *Edgett et al.*, 1997].

Another location where the predicted recent pyroclastic deposits might be preserved is in the polar regions. Again, detailed in situ observations will probably be needed to distinguish ash fall layers formed by an eruption from dust-rich layers caused by dust storms. Such observations coult enable us to observe the interaction between volcanic eruptions and climate, in the same manner as the terrestrial ice cores [*Zielinski et al.*, 1994].

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