An igneous origin for features of a candidate crater-lake system in western Memnonia, Mars

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Received 12 January 2004; revised 18 April 2004; accepted 29 April 2004; published 11 June 2004.

[1] The association of channels, inner terraces, and delta-like features with Martian impact craters has previously been interpreted as evidence in favor of the past existence of crater lakes on Mars. However, examination of a candidate crater-lake system in western Memnonia suggests instead that its features may have formed through igneous processes involving the flow and ponding of lava. Accumulations of material in craters and other topographic lows throughout much of the study region have characteristics consistent with those of volcanic deposits, and terraces found along the inner flanks of some of these craters are interpreted as having formed through drainage or subsidence of volcanic materials. Channels previously identified as inlets and outlets of the crater-lake system are interpreted instead as volcanic rilles. These results challenge previous interpretations of terrace and channel features in the study region and suggest that candidate crater lakes located elsewhere should be reexamined. *INDEX TERMS:* 6225 Planetology: Solar System Objects: Mars; 5415 Planetology: Solid Surface Planets: Erosion and weathering; 8450 Volcanology: Planetary volcanism (5480); 5470 Planetology: Solid Surface Planets: Surface materials and properties; *KEYWORDS:* channel, craters, terrace

Citation: Leverington, D. W., and T. A. Maxwell (2004), An igneous origin for features of a candidate crater-lake system in western Memnonia, Mars, *J. Geophys. Res.*, *109*, E06006, doi:10.1029/2004JE002237.

1. Introduction

[2] There are numerous examples of Martian channels that appear to have once carried inflow to or outflow from impact craters [e.g., *Forsythe and Zimbelman*, 1995; *Cabrol and Grin*, 1999, 2001; *Ori et al.*, 2000; *Cabrol et al.*, 2001]. Channel and crater systems, along with inner terraces interpreted as wave-cut features [e.g., *Forsythe and Zimbelman*, 1995] and flat-surfaced accumulations interpreted as Gilbert-type deltas [*Ori et al.*, 2000], are collectively considered to represent "unambiguous" [*Cabrol and Grin*, 1999] and "unequivocal" [*Ori et al.*, 2000] evidence in support of the past existence of crater lakes on Mars. Purported lacustrine and fluvial systems have been used to support propositions regarding the evolution of Martian climate, and have been identified as locations where the search for evidence of past life on Mars might be carried out [e.g., *Ori et al.*, 2000; *Cabrol and Grin*, 2001; *Cabrol et al.*, 2001]. However, as described below, the results of an investigation of a well-known "crater-lake" system in western Memnonia suggest instead that its channels and terrace features formed and evolved through igneous rather than fluvial and lacustrine processes.

[3] Prior studies of possible crater-lake systems in the study region lacked access to topographic data and high-resolution images of the surface, limiting the amount of information that could be used in interpretations. In addition to Viking Orbiter images, the present study used data generated by the Mars Orbiter Camera (MOC), Mars Orbiter Laser Altimeter (MOLA), and Thermal Emission Imaging System (THEMIS) instruments, providing an improved basis for reexamination of the study region. While parts of the Memnonia study region contain relatively ancient terrain features such as large valleys and elevated highland ridges, some of which display large-scale fluting and gullying that could be related to an early episode of fluvial activity, we concentrate in this paper on relatively young fill materials of craters and intercrater plains, as well as channels inset in those materials. This investigation has yielded unanticipated evidence for a younger, volcanic origin for these features.

2. Study Region

[4] The crater and channel system examined in this study is centered in a ~250 × 350 km study region located in western Memnonia, an area of heavily cratered Noachian-aged highlands characterized by a wide range of degradation [*Mutch and Morris*, 1979; *Scott and Tanka*, 1986] (Figure 1). The primary crater of interest in this system (referred to below as the "central crater") is centered at 174.8°W and 14.6°S, and has a diameter of 45 km. This crater has a maximum rim elevation of ~1650 m and a minimum floor elevation of ~375 m. The crater has an almost continuous inner terrace that is tilted toward crater center with convex form [e.g., *Forsythe and Zimbelman*, 1995; *Ori et al.*, 2000], with maximum radial widths of ~5 km (Figure 2). In places, discontinuous segments of additional terrace-like features are also present. Wrinkle

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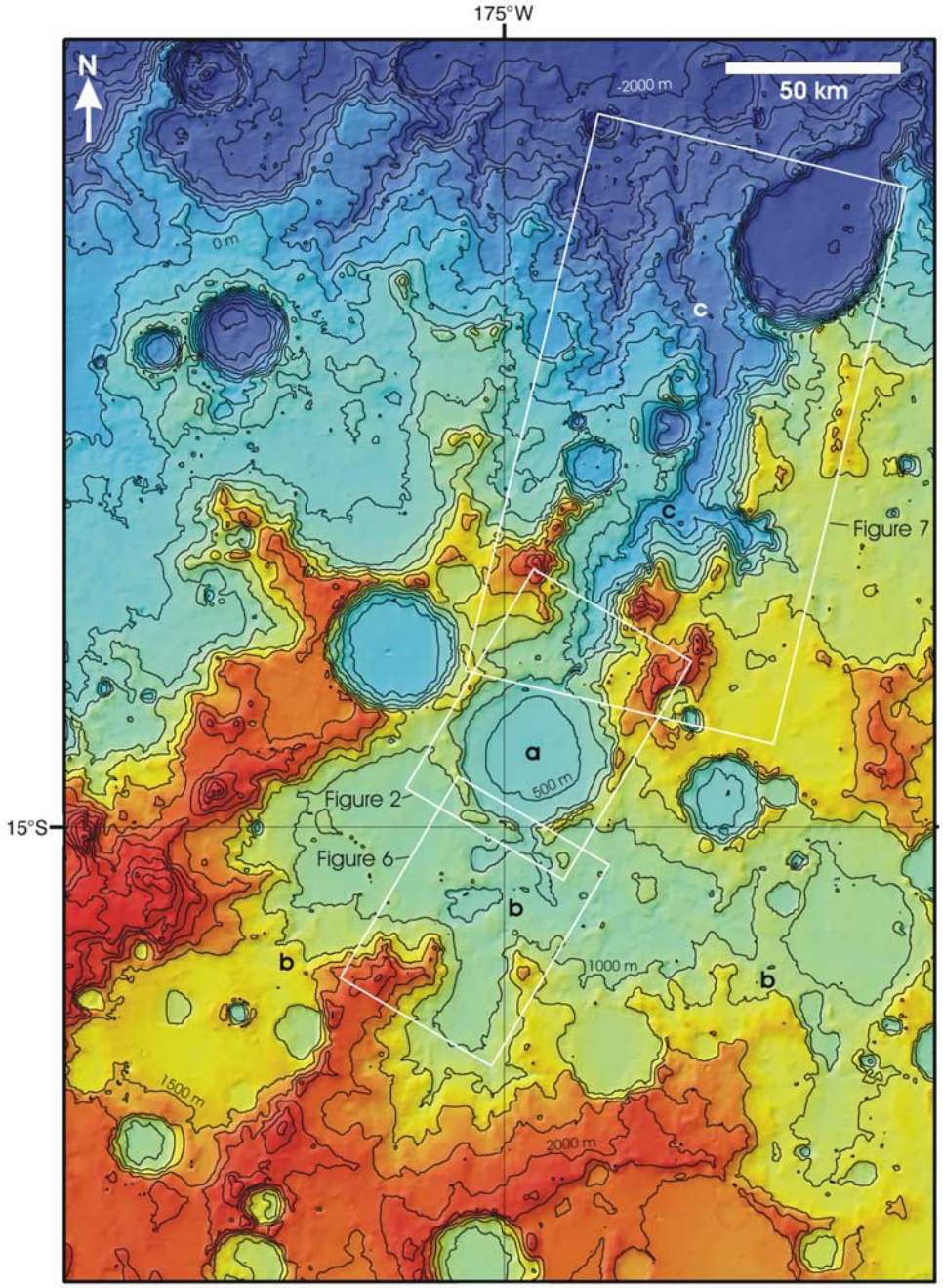


Figure 1. Map of MOLA elevation data superimposed on shaded relief of the study area, showing the locations of (a) the main terraced crater, (b) the inlet basin, and (c) the outlet valley. Contour interval is 250 m. The locations of Figures 2, 6, and 7 are indicated. Topographic data after MEG128 model of *Smith et al.* [2003].

ridges and lobate margins not prominent in Viking data are present in materials that have accumulated in the crater (Figure 3); these features, as well as moat-like features peripheral to scarps formed by units of interior crater fill, are also present in the fill materials of numerous nearby craters (Figure 4). The materials that comprise the uppermost fill units of the main crater, including both the surfaces of the terrace and the inner crater, preserve a common surface texture and a cratering record that differs from that of adjacent highlands for craters with diameters of less than ~300 (Figure 5). Highland material in this region displays a softened appearance relative to crater fill.

[5] The inlet channel of the central crater (Figure 6) is sinuous and flat-floored. At its mouth, the channel has a width of ~3.5 km and a floor elevation of ~620 m. The main inner terrace of the crater extends more than 10 km up the inlet valley [e.g., *Forsythe and Zimbelman*, 1995; *Ori et al.*, 2000]. To the south, the inlet channel has a width of ~800 m, and forms part of a low-order system of relatively narrow and low-relief channels (with typical widths of

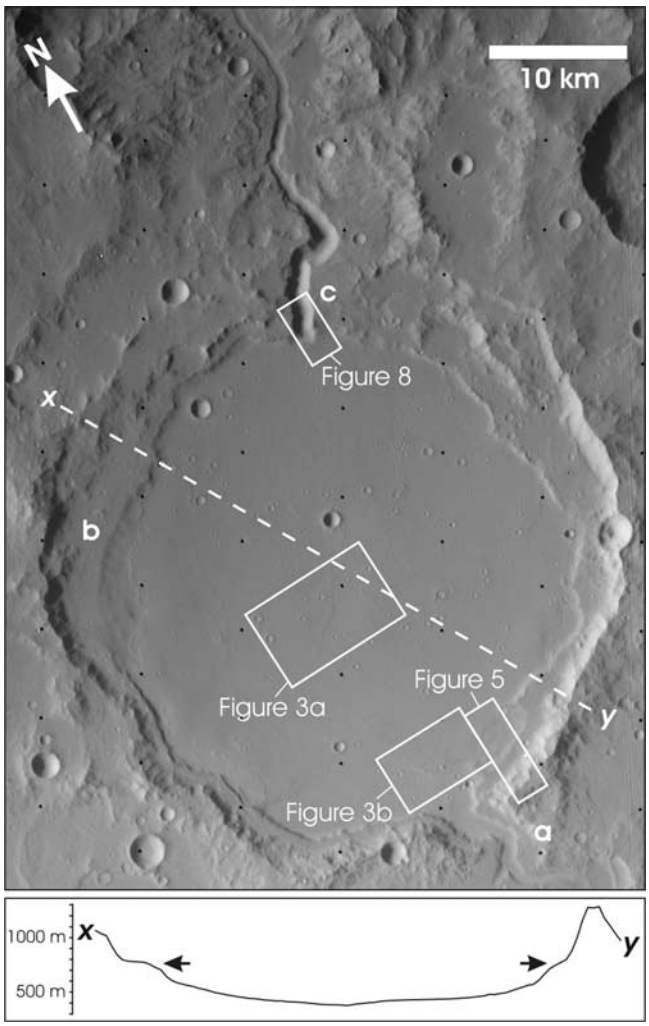


Figure 2. The large terraced crater at the center of the study region (crater diameter = 45 km). Features are labeled as follows: (a) inlet, (b) terrace, and (c) outlet. The locations of Figures 3a, 3b, 5, and 8 are indicated. Elevation profile across line *x-y* is shown at bottom [after *Smith et al.*, 2003]; note the prominent terrace along the inner crater flanks. Image location is given in Figure 1. Viking Orbiter frame 438S12.

~800 m and depths of ~10–25 m) that, together with the craters to which some of these channels are connected, extends throughout much of the inlet basin (Figure 1). The terrain of the inlet basin is mostly rounded and subdued, with the steepest and most dissected slopes mainly confined to relatively small areas at higher elevations.

[6] The outlet channel of the central crater extends northeastward into a 4 to 6 km wide valley that terminates at the highland-lowland boundary [*Scott and Tanaka*, 1986], ~160 km from the northern perimeter of the main terraced crater (Figure 7), and immediately south of deposits of the Medusae Fossae Formation [e.g., *Bradley et al.*, 2002]. The outlet channel is explicitly distinguished here from the "outlet valley", into which the outlet channel is nested. The channel begins at a blunt amphitheater-like headwall (Figure 8) at an elevation of ~570 m and with a maximum channel width of ~1.8 km. The channel sinuously extends northeastward with a gradual decrease in depth corresponding to a reduction in local slope, fading into relatively flat-floored valley fill about 40 km from the channel head. The channel's deepest elevation relative to its rim is ~325 m. A distal segment of the channel appears further down the valley where slope increases, widening to ~3 km before narrowing and fading into deposits of the northern lowlands. The outlet valley has one major tributary valley that intersects it from the southeast, about 50 km from the mouth of the outlet channel (Figure 8). Crater ejecta covers part of the valley floor near this intersection. The full length of the main outlet valley is ~190 km. The gradient of the outlet channel ranges between ~6 and 18 m/km, and the overall gradient of the outlet valley is ~12 m/km.

3. Reevaluation of the Lacustrine Model

[7] Previous workers have suggested that the past existence of lacustrine and fluvial environments on Mars can account for the association of channels, inner terraces, and delta-like features with impact craters [e.g., *Forsythe and*

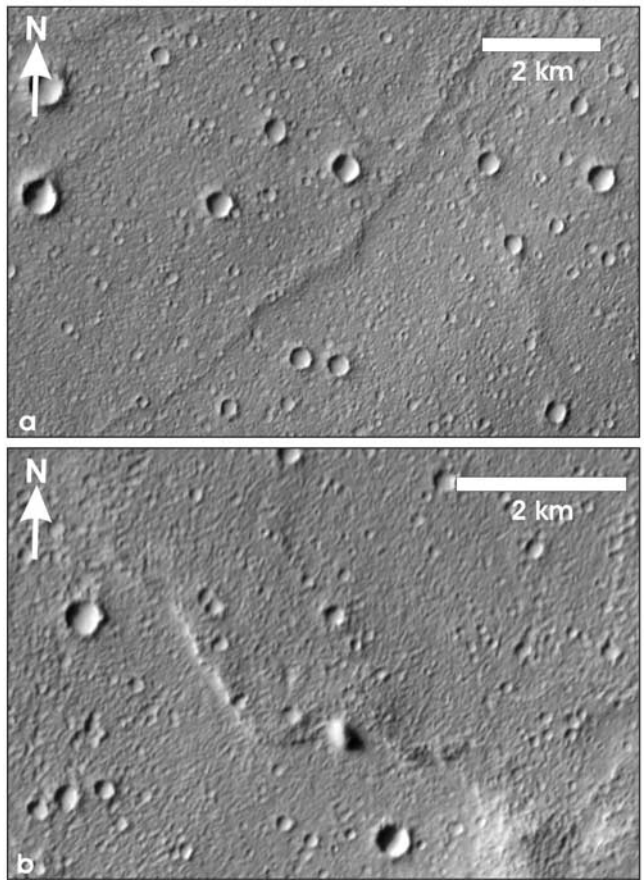


Figure 3. Floor materials of the terraced crater showing (a) wrinkle ridges and (b) lobate margins. Floor materials with similar features are found inside many craters and topographic depressions in the region. Image locations are given in Figure 2. Themis frame V04688002.

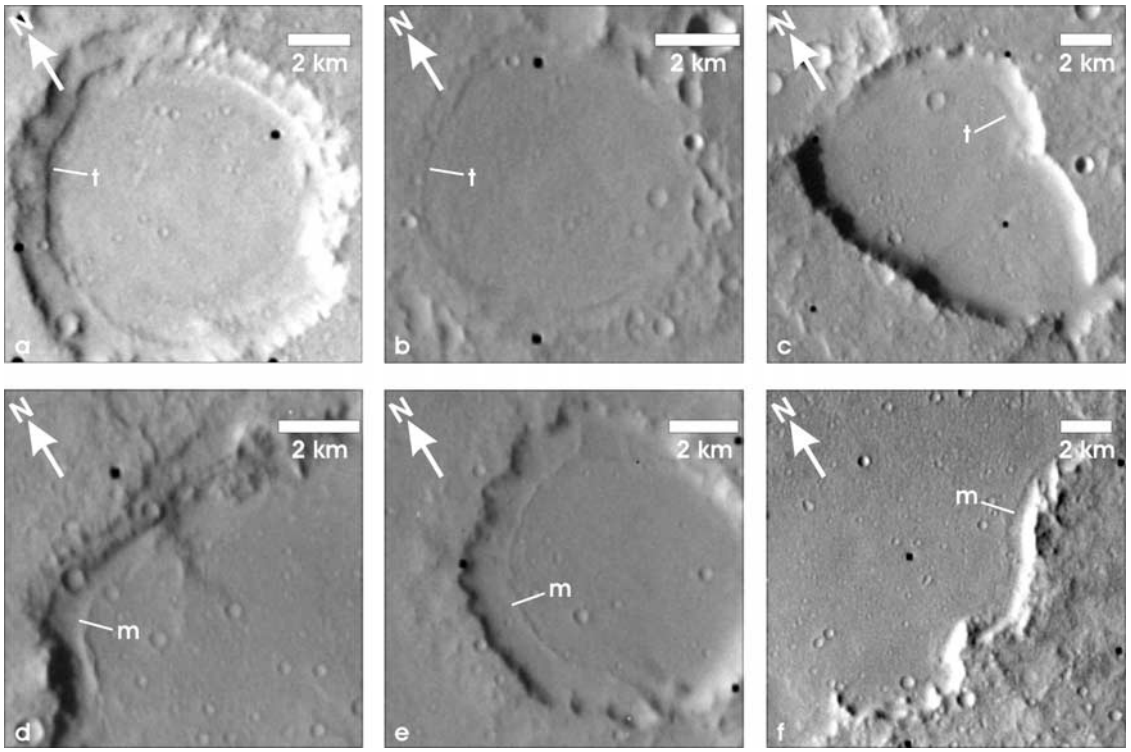
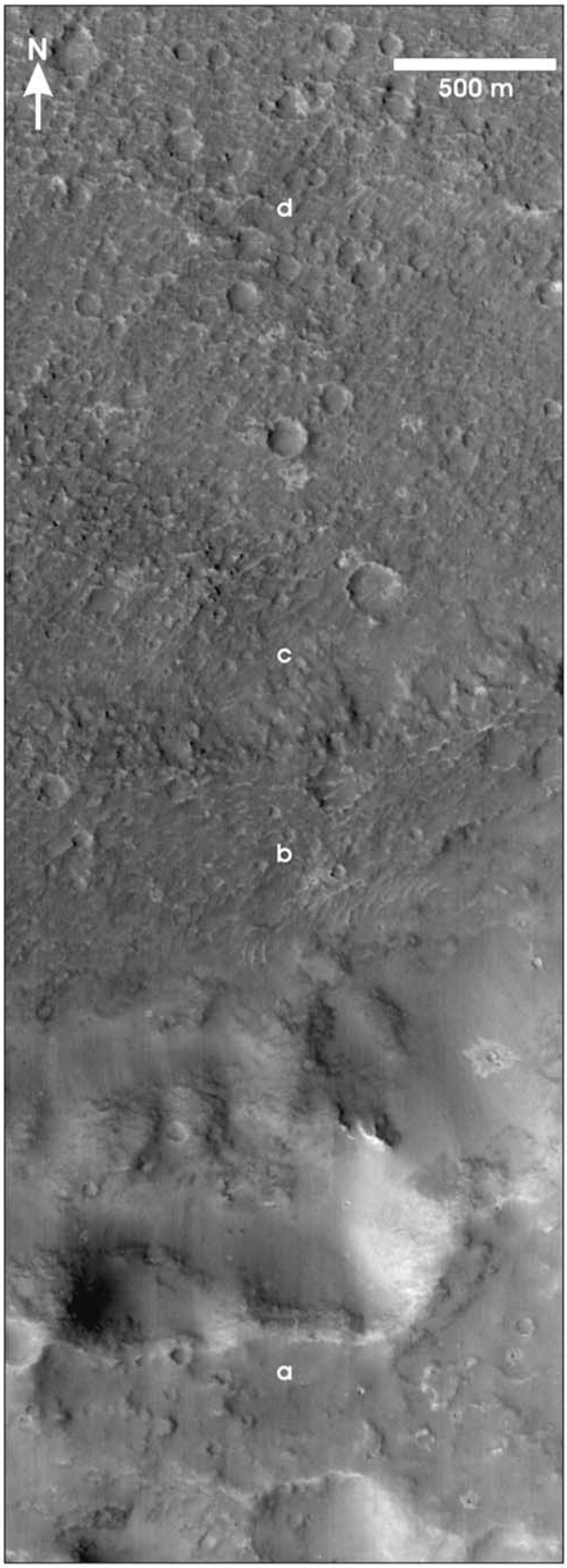


Figure 4. Terraces (t) and moat-like features (m) that are peripheral to fill deposits in craters of the inlet basin (also see Figure 6). The peripheral moats are exterior to and partly formed by narrow adjacent rises. Viking Orbiter frames: (a) 437S14, (b) 437S20, (c) 437S19, (d) 437S14, (e) 436S17, and (f) 437S14.

Zimbelman, 1995; *Cabrol and Grin*, 1999; *Ori et al.*, 2000; *Cabrol et al.*, 2001]. However, recognized modes of formation of lacustrine terraces and channels cannot easily account for the nature of the candidate crater-lake system examined here. For example, extensive lateral dimensions are typically required in order to develop the fetch necessary for significant wave-cutting action. It is unlikely that a maximum fetch of only 45 km in the central crater could have allowed wave action to form terraces that are orders of magnitude larger than most terrestrial lacustrine counterparts [e.g., *Gilbert*, 1885, 1890; *Brophy*, 1967; *Matmon et al.*, 2003]; attenuation of wave energy by a growing terrace and shelf should have further inhibited widening of these features [see, e.g., *Trenhaile*, 1983, 2000, 2002]. The terrace of the central crater is large relative even to most terrestrial marine terraces [e.g., *Keraudren and Sorel*, 1987; *Massari et al.*, 1999; *Polenz and Kelsey*, 1999; *Tortorici et al.*, 2003; *Maeda et al.*, 2004; *Yamaguchi and Ota*, 2004]. The possible origins of the terrace materials themselves are also not compatible with the lacustrine hypothesis. For example, if the material that originally comprised the crater's inner wall contributed the terrace material, then it is not logical for the terrace to extend 10 km up the inlet valley. The absence of a delta at the mouth of the inlet is inconsistent with formation of the terrace by sediments transported through the inlet from the inlet basin.

[8] Both the inlet and outlet channels display inconsistencies with a crater-lake scenario. The outlet channel begins abruptly at full width, with no local development of a nick point or plunge basin as would be expected with continued overflow from a crater lake. The rim of the outlet channel and the fill of the outlet valley are not fluvially dissected, yet such modification would be expected under the same environment that would have maintained a crater lake. Formation of the outlet channel by sapping is not compatible with formation of channels of the inlet basin by substantial water flow, and with ponding of these waters to form a crater lake drained by the outlet channel. Maintenance of lake level at roughly the same elevation over a sufficiently long period to form the terrace, only for lake level to subsequently fall as a result of substantial outlet incision, cannot be accounted for without invoking unlikely scenarios involving catastrophic flood events that are inconsistent with the nature of the central crater and the inlet basin.

[9] Aeolian, erosional, and structural models for the origin of the terrace of the central crater are, as with the lacustrine model, problematic. For example, formation of the terrace by aeolian infilling followed by selective aeolian exhumation of materials in the crater center is not consistent with the lack of similar deposits in numerous nearby craters, nor would such exhumation necessarily be expected to produce a roughly continuous and symmetric terrace that also continues into the crater's inlet. Formation of the terrace following impact or through subsequent crater erosion is not likely, given the absence of evidence of listric-normal slope failure [e.g., *Ori et al.*, 2000]; the extension of the terrace into the inlet valley also weakens this hypothesis, in that structural or mass-wasting



features might be expected to be related to wall failure, and oriented roughly symmetrically about the center of the crater.

4. An Igneous Origin for Crater Fill, Terraces, and Channels in the Study Area

[10] The nature of the central crater, the terraces and channels of the inlet basin, and the channel of the outlet valley are inconsistent with formation and evolution by fluvial and lacustrine processes. However, the morphologies of and interrelations between these features are remarkably consistent with formation by extrusive igneous processes, with interior crater fill having accumulated through deposition and subsidence of materials erupted at the surface during effusive volcanic activity maintained by a system of feeder dykes, and with channels having formed as volcanic rilles that transported lava from overflowing sources and other filled depressions. It is important to note that, while this igneous hypothesis is cited here as a means to account for the origin of fill material and landforms previously considered strong evidence in support of the past existence of lacustrine environments, it is not used to account for the earlier origin of features such as the outlet valley and the gross form of the topography of the inlet basin. These features could have formed through a variety of processes that would have been responsible for general denudation of topography [e.g., *Craddock and Maxwell*, 1993] in the region prior to volcanic resurfacing.

[11] Evidence in support of an igneous origin for the system under study is grouped into three categories: (1) crater fill and associated terraces; (2) outlet channel; and (3) inlet basin.

4.1. Crater Fill and Associated Terraces

[12] Accumulations of material in the interiors of many of the craters of the study region are associated with (1) wrinkle ridges that are suggestive of subsidence and lateral compression of layered volcanic units; (2) surface units with lobate margins and peripheral moats that are suggestive of formation through flow and subsidence of volcanic materials; (3) textural, crater-preservation, and thermal characteristics that are collectively consistent with volcanic materials that are relatively dense and consolidated; and (4) inner crater terraces that are suggestive of subsidence or drainage of volcanic materials.

[13] Wrinkle ridges are present in the fill of the central crater as well as that of craters of the inlet basin (Figure 3a). These features have widths of $\sim$1–2 km, lengths that in some cases exceed 10 km, and heights of meters to tens of meters; these dimensions fall within the size range of small Martian wrinkle ridges [e.g., *Watters*, 1988; *Schultz*, 2000; *Montési and Zuber*, 2003]. Wrinkle ridges are sinuous and elongate topographic highs that are found on Mercury, the

Figure 5. Surface texture of floor materials of the terraced crater: (a) crater rim, (b) terrace, (c) bottom of terrace scarp, and (d) crater floor. Fill materials preserve a cratering record that for diameters <300 m is not preserved in adjacent highlands. Image location is given in Figure 2. Mars Orbiter Camera frame M0903914.

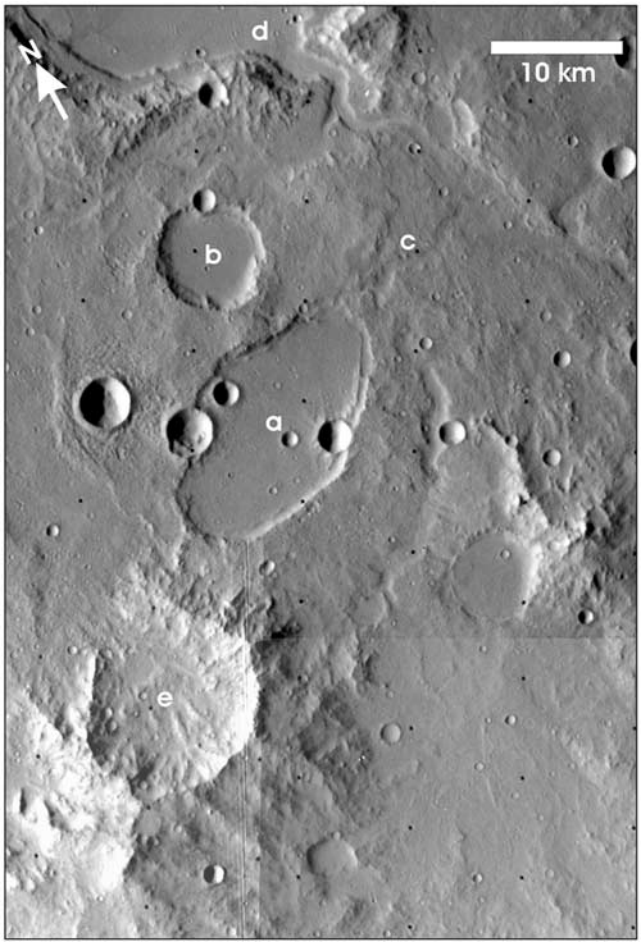


Figure 6. Inlet channel of the main terraced crater, and terraced craters associated with subtle channel features that feed into the inlet channel. Features are labeled as follows: (a) elongate terraced crater with peripheral moat, (b) circular terraced crater, (c) channel extending from crater a, (d) inlet area of main terraced crater, and (e) breached crater. Image location is given in Figure 1. Viking Orbiter frames 437S14, 437S15, and 437S16.

Moon, Venus, and Mars, and are believed to form through horizontal shortening of near-surface layered deposits [e.g., *Watters*, 1988, 1991; *Golombek et al.*, 1991; *McGill*, 1993; *Schultz*, 2000]. Although the occurrence of wrinkle ridges on the terrestrial planets is not in principle restricted to materials with a volcanic origin [see, e.g., *Schultz*, 2000], most or all clear examples of these features are found in materials interpreted to be layered and volcanic in origin [*Watters*, 1988]. On this basis, and on the basis that wrinkle ridges are very commonly formed in lunar and Martian basaltic deposits [e.g., *Muehlberger*, 1974; *Maxwell*, 1978; *Solomon and Head*, 1979; *Watters*, 1988; *Head et al.*, 2002], the existence of wrinkle ridges in the study region is interpreted as suggestive of a volcanic origin and layered nature for materials that have accumulated in topographic depressions.

[14] Many geological units deposited in craters and other depressions in the study region have lobate margins that typically have local elevations of <10 m, but can reach ~10–30 m in height (Figure 3b). The thicknesses and appearance of these units are comparable to, for example, volcanic flows of the Martian highlands and lowlands [e.g., *Schaber*, 1980; *Baloga et al.*, 2003; *Ivanov and Head*, 2003] and prominent lunar volcanic flows in Oceanus Procellarum and Mare Imbrium [e.g., *Schaber*, 1973; *Schultz*, 1976; *Zimbelman*, 1998; *Hiesinger et al.*, 2002] (Figure 9). Some lobate units in the study region form prominent scarps that are peripheral to crater-fill deposits (Figures 4e–4g and 6) and that can resemble the peripheral flow features of lunar lava lakes [e.g., *El-Baz and Roosa*, 1972; *Schaber et al.*, 1976] and lunar craters containing subsided accumulations of volcanic materials [*Schultz*, 1976] (Figure 10). The similarity between the nature of the lobate units in the study region and those of volcanic

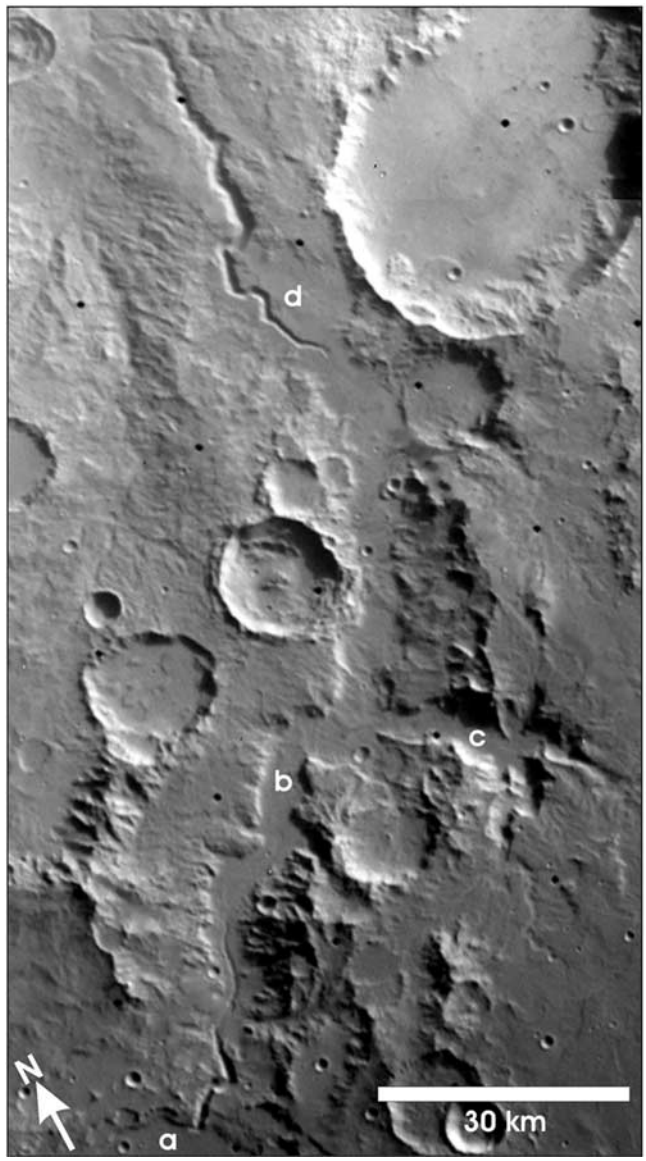


Figure 7. Outlet valley of main terraced crater. Features are labeled as follows: (a) outlet head, (b) outlet valley floor, (c) tributary valley, and (d) distal outlet channel. Image location is given in Figure 1. Viking Orbiter frames 599A70 and 599A72.

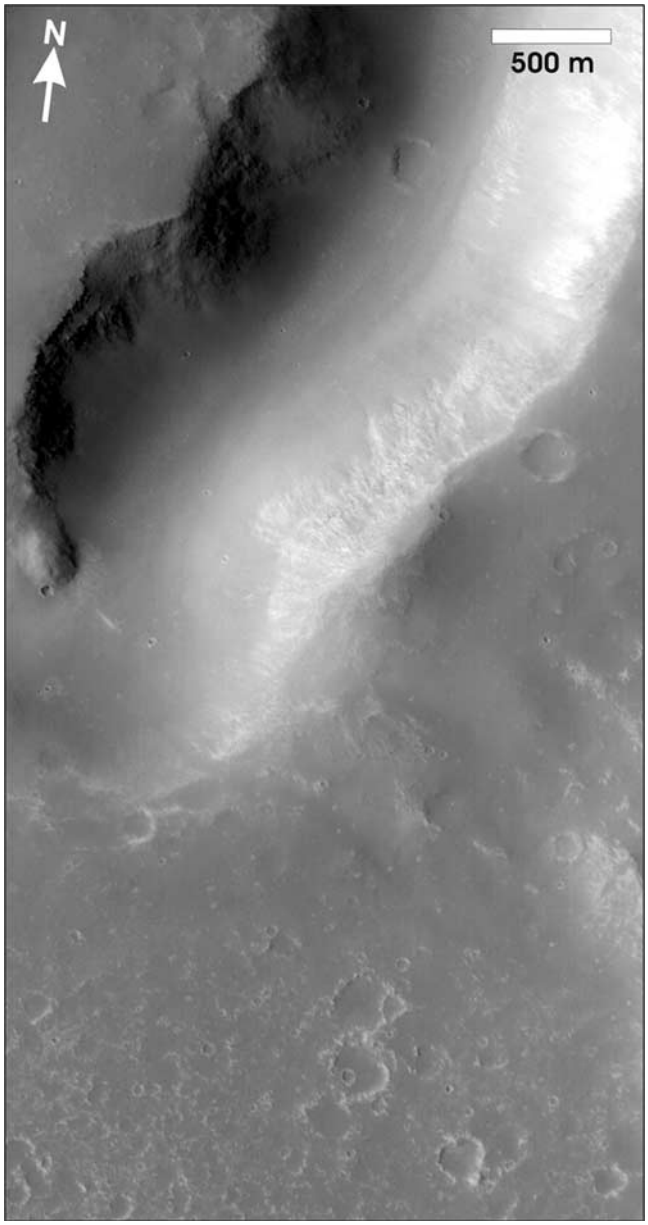


Figure 8. Head of outlet channel. Image location is given in Figure 2. Mars Orbiter Camera frame M0702911.

units located on Mars and the Moon is consistent with a volcanic origin for these units.

[15] The surfaces of the terrace and inner fill materials of the main crater have an identical surface texture as viewed in high-resolution MOC images (Figure 5). This texture is similar to that of materials that have accumulated on the flanks of Martian volcanoes such as Hadriaca Patera, and suggests that both the terrace and inner fill materials may have had a volcanic origin. A volcanic origin of these materials is supported by preservation of a cratering record not seen in adjacent uplands (Figures 5 and 8); on the Moon, relatively young volcanic deposits similarly preserve cratering records that are not found in surrounding older but less-consolidated materials (Figure 11). The nighttime thermal properties of crater fill of the central terraced crater and numerous other craters in the study region are suggestive of surface materials that are relatively dense and consolidated compared with materials in surrounding areas (see, e.g., Themis frame I06841006).

[16] Terraces that may not have formed through mass-wasting or tectonic processes are found on the inner flanks of numerous craters in the study region, and include the relatively narrow terraces of craters in the inlet basin (Figure 4a) and the relatively broad terrace associated with the central crater (Figure 2). The formation of terraces and moats inside impact craters and along the perimeters of fill accumulations through subsidence of volcanic materials is believed to have occurred on the Moon [e.g., *Holcomb*, 1971; *El-Baz*, 1972; *Greeley*, 1976; *Schultz*, 1976; *Young*, 1976; *Greeley and Spudis*, 1978] (Figures 12, 13, and 14), perhaps involving such well-known terrestrial processes as degassing or cooling of subsurface magma [see, e.g., *Francis et al.*, 1993], or involving subsidence through loading. The lunar terraces are morphologically similar to the narrow terraces found in the study region, and are suggestive of formation through analogous igneous processes.

[17] While the broad terrace of the central crater could have also formed through subsidence of volcanic materials, other mechanisms are possible. The cooling and crystallization of a lava lake, taking place along all exterior surfaces of the lava body [e.g., *Rüpke and Hort*, 2004], could, depending on associated degassing processes, result in terrace-forming subsidence of the cooled surface. Alterna-

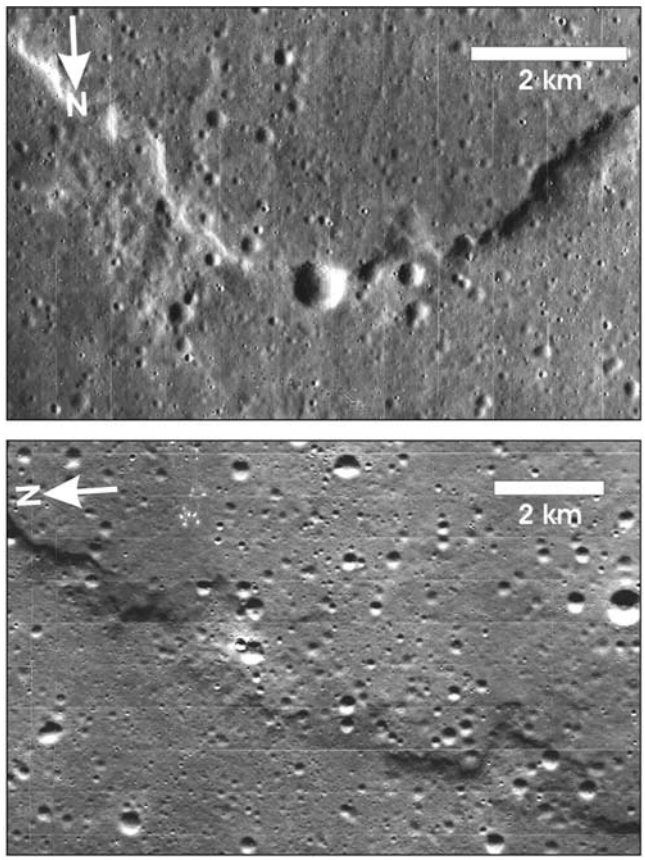


Figure 9. Examples of flow fronts in Mare Imbrium. Lunar Orbiter frames V-160-H2 and V-161-H2.

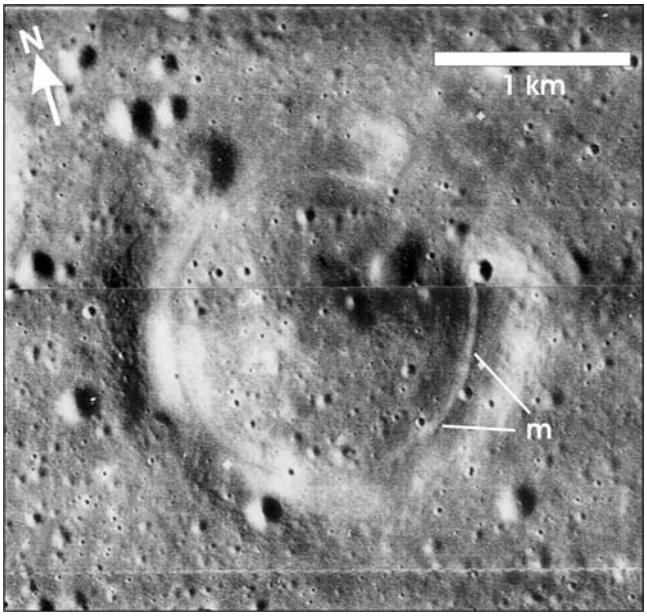


Figure 10. Moat (m) believed to have been formed through subsidence of volcanic deposits that once covered this crater [*Schultz*, 1976]. Lunar Orbiter frame III-150-M.

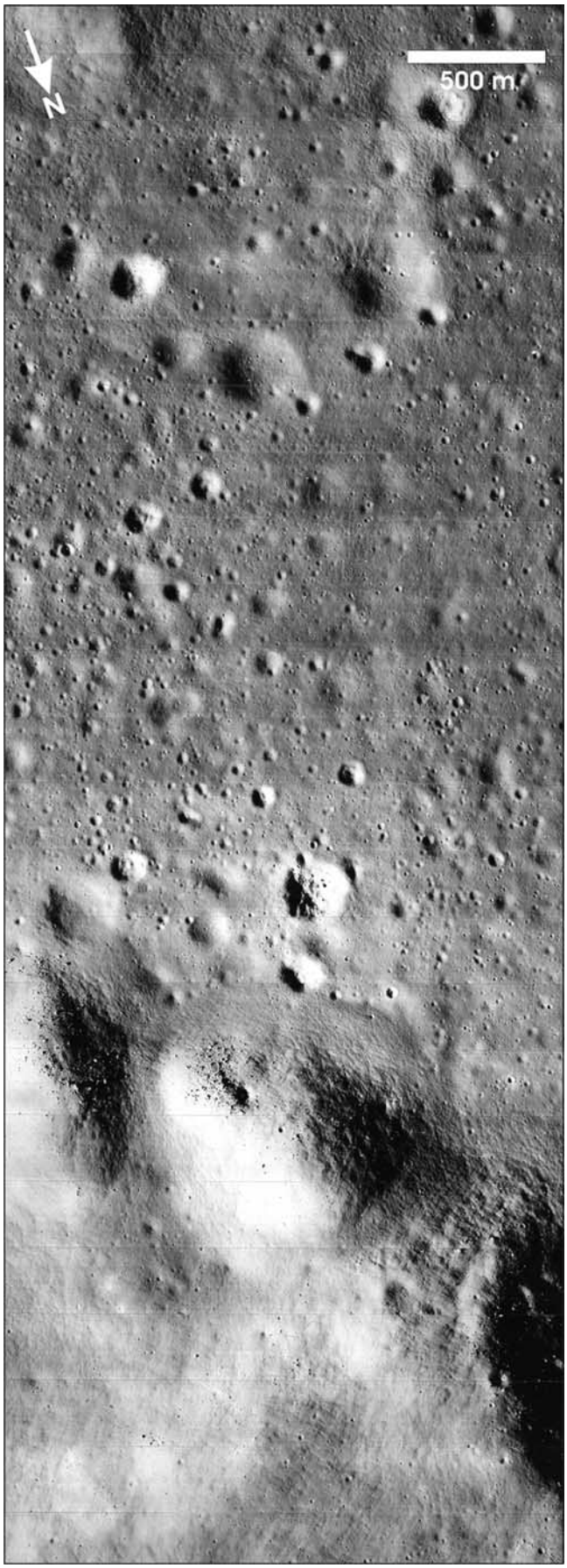
tively, there are abundant terrestrial examples in which inner terraces have formed in calderas by substantial collapse caused by evacuation of subsurface magma chambers by degassing, lateral intrusion, or drainage events [e.g., *Young*, 1976; *Francis et al.*, 1993; *Geshi et al.*, 2002; *Gottsmann and Rymer*, 2002]. Repeated filling and drainback of terrestrial lava lakes, driven by cyclic processes such as vesiculation within the magma column beneath the source vent, can also produce distinct terraces [e.g., *Burgi et al.*, 2002; *Barker et al.*, 2003]. On the basis of these terrestrial analogs, it is possible that the broad terrace in the main crater formed through the occurrence of at least one drainage event of a lava lake located inside the crater. If such an event took place, then the terrace in this crater would correspond to an earlier, higher filling level, and the outlet channel would have acted as a conduit for this and subsequent releases of lava. Lava terraces are present in Bowditch, a 35 km long lunar crater located northeast of Mare Australe (Figure 14). These terraces are up to 3.5 km wide and 50 to 200 m high [*Young*, 1976], and are believed to mark the high level of a lava lake that breached the crater rim [*El-Baz*, 1972; *Young*, 1976]. A large patch of mare-like material (Lacus Solitudinis) extends ~120 km in a southeastward direction from the breach. Crater Bowditch and its terraces are of comparable size, morphology, and appearance to the central crater of the Memnonia study region (Figure 2). The similarity between the appearance, dimensions, and geological context of these features is consistent with their common formation by breach and outflow of lava lakes.

4.2. Outlet Channel and Valley

[18] The outlet channel is a sinuous single channel that heads at full width and gradually tapers in width in a downchannel direction, has sharp and undissected rims, and progressively shallows until it disappears into relatively flat-floored valley materials before re-appearing at the steeper slopes that extend onto the volcanic plains of the northern lowlands. As such, the morphological characteristics of this channel closely match those of many sinuous rilles on the Moon [e.g., *Schubert et al.*, 1970; *Greeley*, 1971a; *Schultz*, 1976; *Guest and Murray*, 1976; *Strain and El-Baz*, 1977; *Hulme*, 1982] (Figure 15). The full length of the outlet channel, including the flat-floored section where surface expression of the channel is absent, is ~190 km; the greatest channel widths range between ~1.8 and 3 km, and maximum depth is ~325 m. These dimensions are generally consistent with those of such lunar rilles as Hadley, which is ~135 km long and averages 1.2 km in width and 370 m in depth [*Greeley*, 1971b].

[19] Lunar sinuous rilles are believed to have formed in a manner at least partly analogous to terrestrial volcanic channels or lava tubes [e.g., *Oberbeck et al.*, 1969; *Greeley*, 1971b; *Cruikshank and Wood*, 1972; *El-Baz and Roosa*, 1972; *Young et al.*, 1973; *Carr*, 1974], which form as conduits for lava streams in effusive flows [see, e.g., *Stephenson et al.*, 1998; *Calvari and Pinkerton*, 1998; *Kauahikaua et al.*, 1998]. Levees often form along active flow margins of lava channels, and formation of fully enclosed lava tubes can result from processes such as accretion on levees or aggregation of floating crustal rafts [e.g., *Peterson et al.*, 1994; *Kauahikaua et al.*, 1998]; formation of crusted roofs can dramatically decrease cooling rates by insulating the core of the flow from direct radiative and convective cooling, allowing basaltic lava flows to form with lengths of hundreds of kilometers [e.g., *Keszthelyi and Self*, 1998; *Sakimoto and Zuber*, 1998; *Harris and Rowland*, 2001]. The roofs of lava tubes can partially or completely collapse following drainage of parent flows [e.g., *Papson*, 1977; *Calvari and Pinkerton*, 1998], sometimes producing the appearance of discontinuous channels. Steeper segments of terrestrial and lunar lava tubes typically form no or weak roofs, while segments with low slopes can form stronger roofs and are less likely to drain [e.g., *Greeley*, 1971b; *Sakimoto and Zuber*, 1998]; as a result, there can be preferential preservation of the roofs of tubes along segments of low slope, a situation that may apply to the morphology of the outlet channel in the Memnonia study region. The absence of obvious accumulations of material that flowed out of the outlet channel (Figure 7) is a quality that is typical of lunar sinuous rilles [e.g., *El-Baz et al.*, 1972], and in the case of the Memnonia outlet channel could be related to burial by later volcanic flows. The absence of fluvial dissection of the margins and interior of the outlet channel is consistent with the channel's origin as a volcanic feature that formed under relatively dry climatic conditions that have persisted until the present.

[20] The edges of the outlet channel near the channel head are at the same level (and are composed of) the main crater terrace (Figures 2 and 8). Even if the outlet channel did act as a conduit for lava, initial incision of the crater rim near the present outlet head may have taken place during an earlier stage of fluvial activity. It is alternatively conceivable that the present morphological relationship between the rim and terrace of the main crater and the head of the outlet channel may have developed after lava overtopped the



crater rim, causing incision. Or, northward lava flow from the main crater may have instead been initiated below the surface of the crater rim along structural or other zones of weakness, causing subsequent collapse. The capacity of the flow of molten materials to penetrate crater rims and other geological features, through possible processes such as simple incision (downcutting and sidecutting) into regolith or through exploitation of preexisting sub-surface areas of weakness, is illustrated by, for example, breaches of the rims of the Elysium caldera on Mars (Figure 16) and crater Krieger on the Moon (Figure 17).

[21] Could the flow of lava have breached the surface or subsurface of the rim of the central crater? Volcanic rilles usually form as constructional features that emplace lava flows, and incision is often not important in their formation [e.g., *Young et al.*, 1973; *Carr*, 1974; *Greeley*, 1977]. Furthermore, the morphologies of deep lunar lava channels can in some cases simply be attributed to factors such as preexisting topography and constructional processes associated with the formation of levees [e.g., *Sparks et al.*, 1976; *Greeley*, 1987; *Spudis et al.*, 1988] (indeed, the morphological characteristics of the full length of the Memnonia outlet channel north of the crater breach are consistent with that of a constructional channel formed during flow and infill of lava in a preexisting valley). Nevertheless, while there is much that is not understood about the environmental and eruptive conditions necessary to support the formation of erosional channels [e.g., *Fagents and Greeley*, 2001], thermal and mechanical processes of erosion by flowing lava have been recognized as important through both theoretical studies [*Huppert et al.*, 1984; *Huppert and Sparks*, 1985; *Kerr*, 2001] and terrestrial field studies of active and ancient flows [e.g., *Peterson and Swanson*, 1974; *Barnes and Barnes*, 1990; *Greeley et al.*, 1998; *Kauahikaua et al.*, 1998, 2002, 2003; *Williams et al.*, 2004]. A basaltic erosive rate of 0.1 m/day over a period of 60 days has been measured for a relatively small lava stream with laminar flow at Kilauea Volcano in Hawaii, suggesting a capacity for local-scale basaltic erosion to incise tens of meters into a basaltic substrate over a period of less than one year [*Kauahikaua et al.*, 1998]. The erosive capacity of lava increases with higher temperatures and volumes, lower viscosities, and increased turbulence [see, e.g., *Greeley et al.*, 1998; *Williams et al.*, 1998], suggesting a far greater capacity for erosion for lunar and Martian flows associated with large rilles. Capacity for erosion may be still greater for cases in which incision involves a regolith substrate rather than bedrock.

[22] On the basis of theoretical and field studies, thermal and mechanical processes of erosion associated with lava streams are expected to have been an important geomorphological process in a number of regions of the Moon, Venus, Earth, and Mars [e.g., *Carr*, 1974; *Bussey et al.*, 1995; *Williams et al.*, 1998]. It is therefore reasonable to conclude that lava could have had the capacity to breach the rim of the central crater of the Memnonia study region,

Figure 11. Greater preservation of cratering record by younger and more consolidated lunar fill materials relative to surrounding older materials. Compare with Figures 5 and 8. Lunar Orbiter frame III-200-H2.

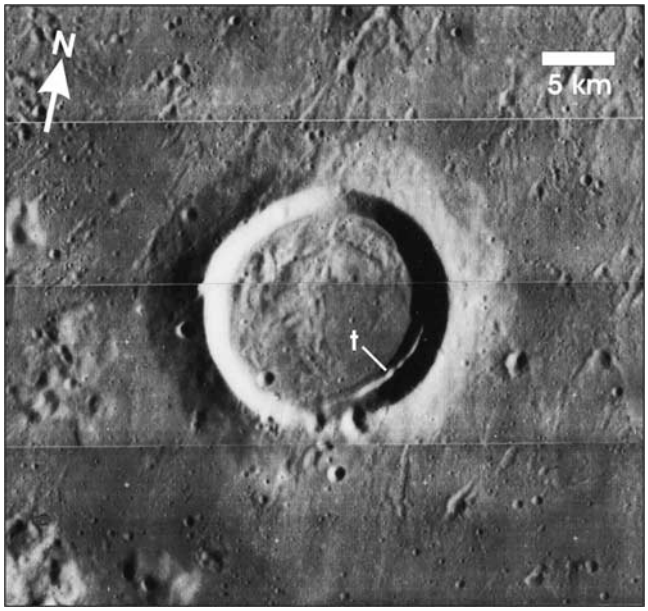


Figure 12. Lunar impact-crater terrace believed to have been formed through subsidence of volcanic fill [*Schultz*, 1976]. Compare with Figures 4 and 6. Lunar Orbiter frame IV-195-H2.

and to form the outlet channel and valley fill to the north. The relation of the outlet valley itself to its flat-floored fill and central sinuous rille is similar to Valles Alpis (Alpine Valley) on the Moon, a lava-flooded breach in the highlands located between Mare Frigoris and Mare Imbrium [e.g., *Schultz*, 1976; *Cook and Hiesinger*, 1996] (Figure 18). While the two valleys likely formed through very different processes, the characteristics of the flat-floored valley fill and central channel of the Memnonia outlet valley (Figure 7) are, by analogy with those features in Valles Alpis, further suggestive of formation through infilling and flow by extrusive volcanic materials.

4.3. Inlet Basin

[23] The surface of the inlet basin (Figure 1) is crossed by shallow channels (Figure 19), typically of roughly constant widths, that connect craters and other depressions that have been filled with deposits (Figures 4 and 6; see also section 2) interpreted above as volcanic in origin (sections 4.1 and 4.2). The distribution and morphology of these channels are consistent with formation by overtopping of confining rims of topographic lows into which volcanic materials accumulated, perhaps in a manner analogous to the formation of channels by constricted overflow of lava lakes on the Moon [e.g., *Schaber*, 1973]. Channels in the inlet basin ultimately converge on the inlet of the central crater (Figure 5); the extension of the crater terrace into the inlet valley is suggestive of a terrace origin that post-dates formation of the crater [*Ori et al.*, 2000], and is consistent with formation by lava flow through the inlet during periods that followed reductions from earlier, higher levels. The inlet breach in the crater wall could conceivably have formed by lava erosion, if lava ponded at the lowest part of the inlet basin, although initial formation of the inlet breach through earlier fluvial processes is also possible.

[24] Lava flows have the capacity to follow routes defined by the gradients of preexisting topography [e.g., *Swanson and Wright*, 1978; *Reidel*, 1998; *Branca*, 2003], to incise channels (section 4.2) [see also, e.g., *Kauahikaua et al.*, 1998], and to flow hundreds of kilometers at very high effusion rates [e.g., *Wilson and Head*, 1994]. However, while there are documented sources of volcanic materials in Memnonia and surrounding regions (one of which is less than 50 km east of the outlet valley) [*Scott and Tanaka*, 1986], there are no geomorphological features yet recognized in the immediate study region that can be unambiguously described as sources of effusive flows. The absence of obvious constructional volcanic features within the inlet basin argues against a local source for hypothesized volcanic flows, although it may be speculated that volcanic processes in the region could have been analogous to those which flooded lunar craters in the highlands and regions such as Mare Australe and Mare Smythii, where material is believed to have been extruded within each crater from separate feeders that were linked to a magma reservoir [*Schultz*, 1976; see also *Whitford-Stark*, 1982; *Head and Wilson*, 1992; *Yingst and Head*, 1997]. The absence of clear volcanic sources under such eruptive conditions would not be surprising; source regions for many lunar mare and highland basalts are apparently obscured by the lava flows themselves, making identification of sources tentative or

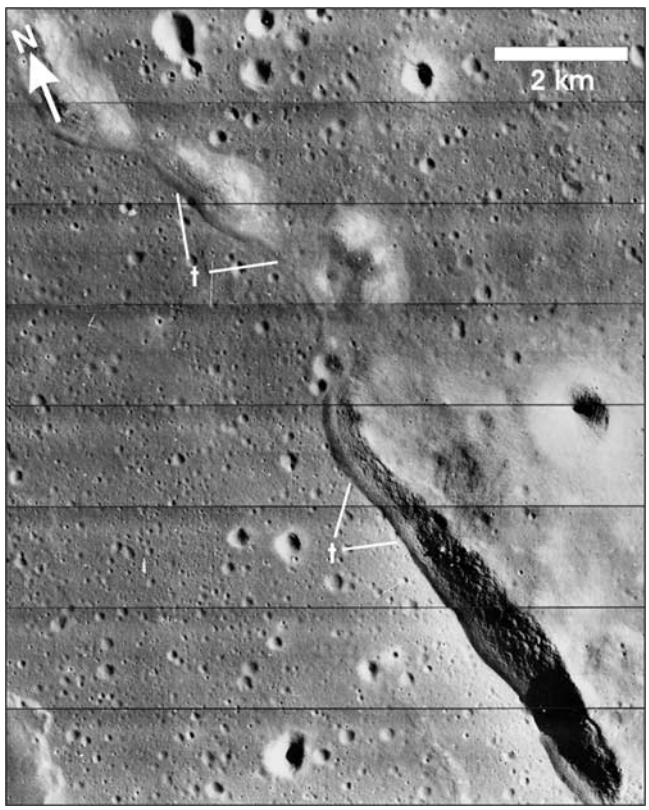


Figure 13. Terraces (t) located in the Flamsteed region of the Moon and believed to have formed through subsidence of volcanic deposits [*Schultz*, 1976]. Lunar Orbiter frame III-200-M.

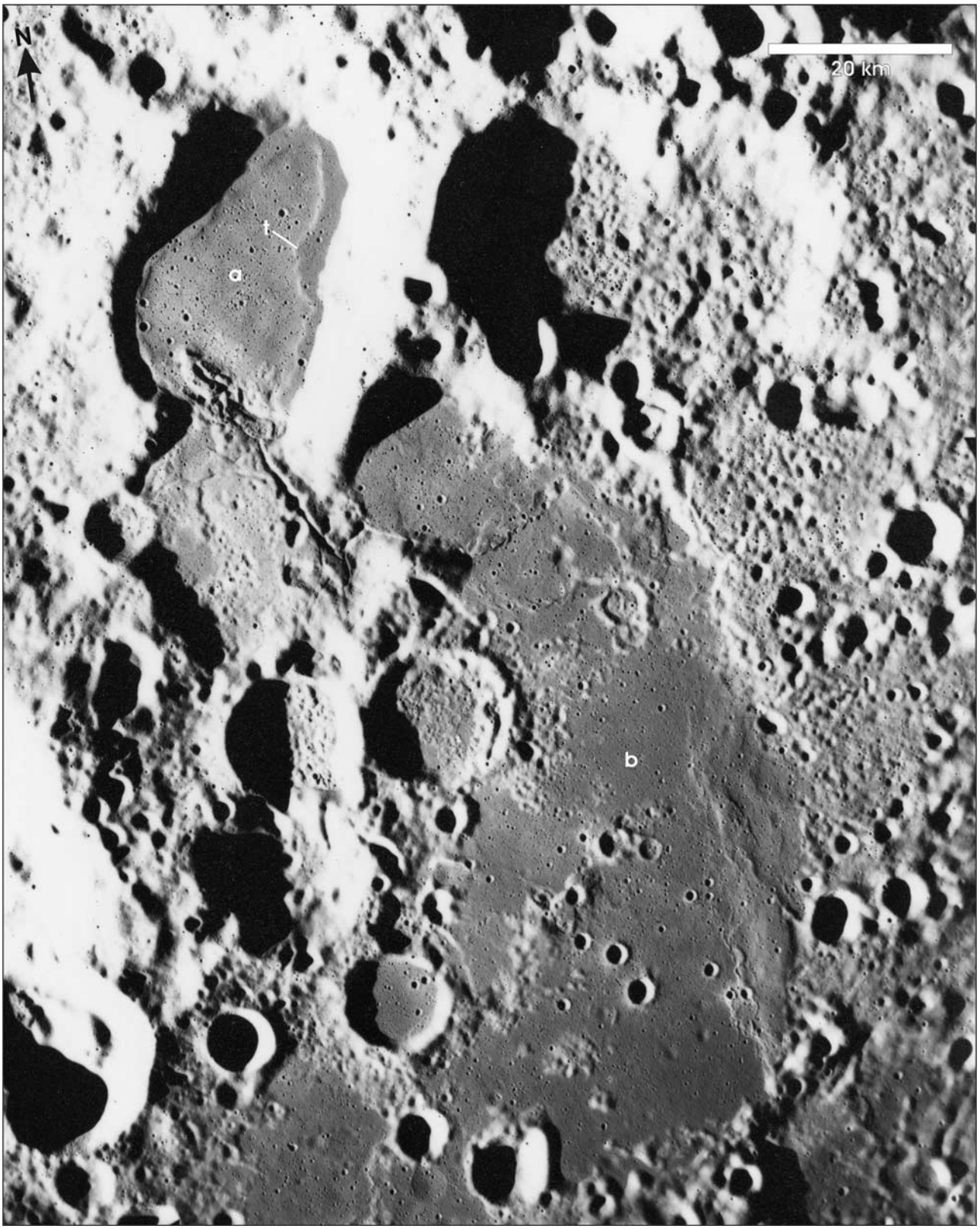


Figure 14. Lava terraces (t) in Bowditch (a), a 35 km long crater in the highlands of the lunar farside, northeast of Mare Australe. The terraces are believed to mark the high level of a lava lake that breached the crater rim, forming a large patch of mare-like material (Lacus Solitudinis; b) [*El-Baz*, 1972; *Young*, 1976]. Apollo 15 Metric Camera frame AS15-M3-2628 (see also Apollo 15 Panoramic Camera frame 9965).

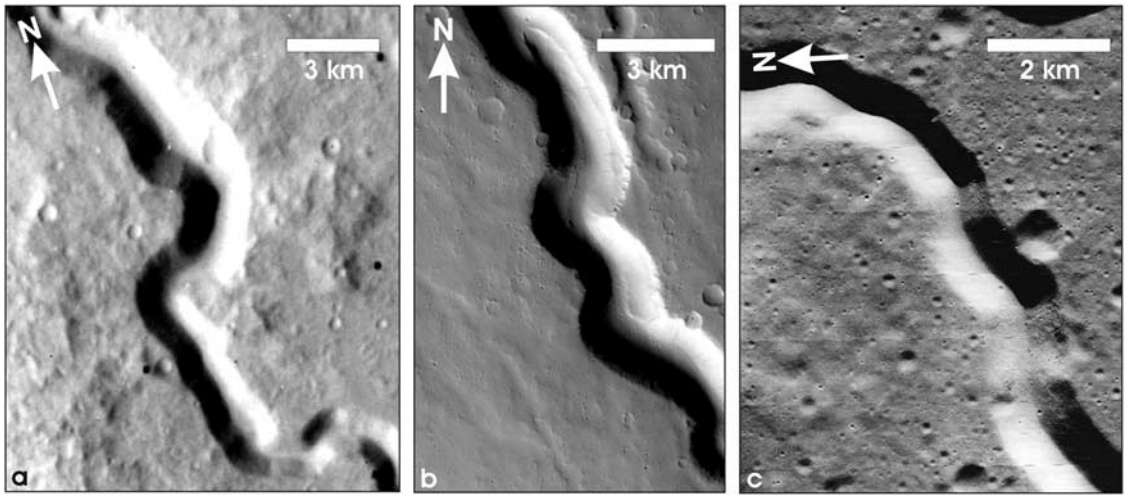


Figure 15. Comparison of (a) distal outlet channel of main terraced crater (Viking Orbiter frame 439S07; see also Figure 7) with lava conduits at (b) Ceranius Tholus, Mars (Themis frame V01002003), and (c) Hadley Rille (Lunar Orbiter frame V-105-H2).

impossible [e.g., *Whitford-Stark*, 1982; *Wilhelms*, 1987], and it is not unusual for sources of Martian lava flows to be similarly difficult to identify, even for flows that occur in otherwise obvious volcanic regions [e.g., *Yingst and Head*, 1997; *Baloga et al.*, 2003].

[25] Despite the lack of clear volcanic features within the inlet basin, the study region is located within a broad area that has clearly been subjected to highland volcanism. Extensive plains units characterized by flow lobes and wrinkle ridges, interpreted as highland flows of low-viscosity lava erupted from numerous sources at high rates, are found immediately south of the divide that defines the inlet basin [*Scott and Tanaka*, 1986]; there are several breaks in the divide that separated these flows from the inlet basin (e.g., at 173°52′W, 16°48′S), providing possible locations from which the inlet basin could have been repeatedly flooded by lava. Volcanic features located within several hundred kilometers of the study region (in adjacent regions of Memnonia, Aeolis, and Phaethontis quadrangles) include the fissure- and caldera-like depressions of *Scott and Tanaka* [1986], faults and graben of Memnonia Fossae [e.g., *Tanaka and Chapman*, 1990; *Wilson and Head*, 2002], and the lowland shield volcano, Apollinaris Patera.

5. Discussion

[26] The igneous hypothesis described in section 4 would have essentially involved an interlinked system of surface lava conduits and overflowing basins, with candidate lava sources including those of the extensive region of highland flows located immediately to the south of the study region. Under this hypothesis, most or all craters and other topographic depressions in the study region would have been passive repositories for multiple layered volcanic deposits, and, at times, lava lakes; not being sustained by mass exchange with subsurface magma chambers, individual lava lakes would likely have been very short-lived [see, e.g., *Worster et al.*, 1993; *Burgi et al.*, 2002]. Flow of volcanic materials through the inlet basin, central crater, and outlet channel would have accumulated on the plains north of the dichotomy boundary. The initial northward breach of the rim of the central crater would have stranded the terrace of the main crater, marking the elevation of highest fill in that crater.

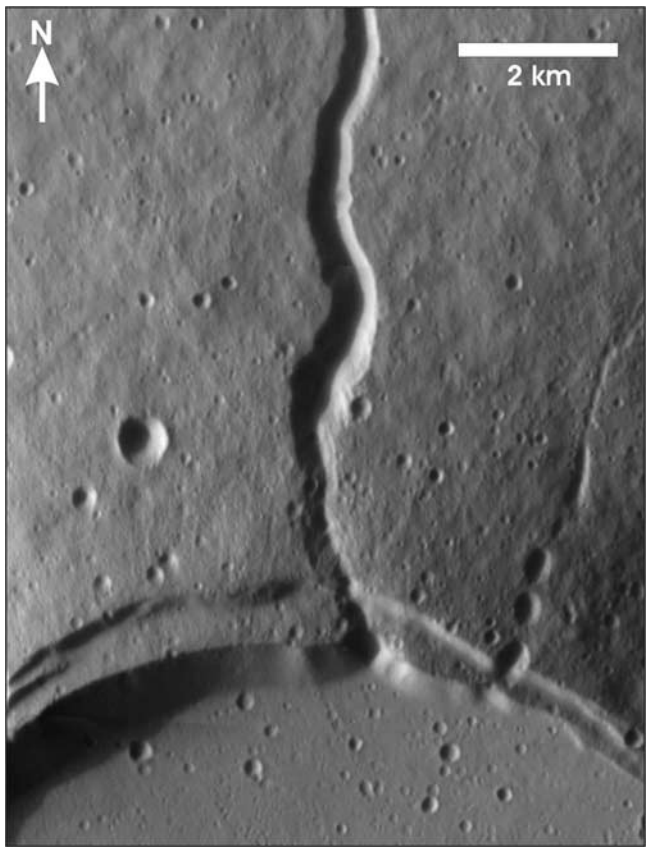


Figure 16. Breached rim of the terraced caldera of Elysium Mons, Mars (Themis frame V05450016).

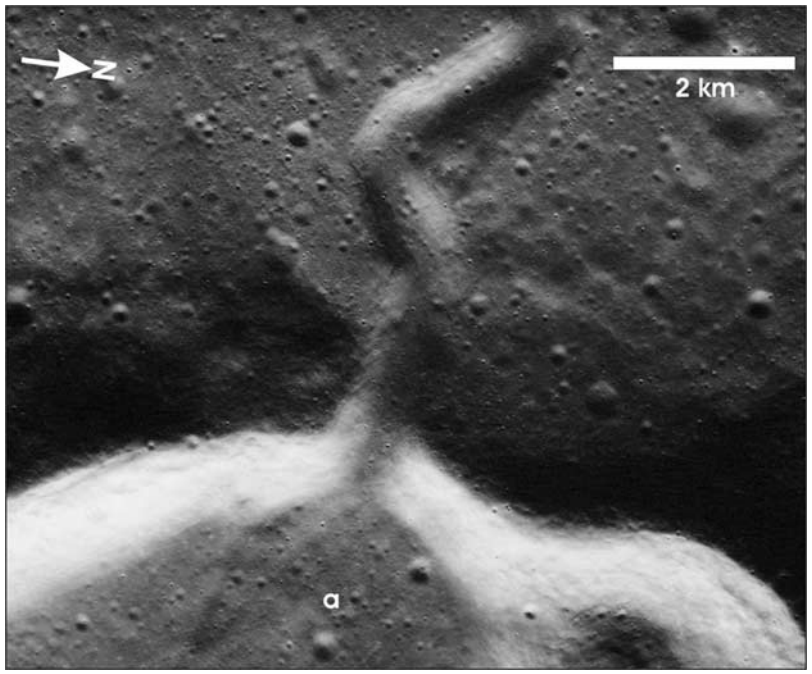


Figure 17. Breach of lunar crater Krieger (a) by a rille (Apollo 15 Panoramic frame 0327).

[27] If formation of channels, terraces, and associated fill accumulations in western Memnonia indeed occurred through igneous rather than fluvial and lacustrine processes, there is the potential for far-reaching consequences regarding (1) our understanding of the formation mechanisms of similar Martian features located in other regions, and of the possible interplay between volcanic and fluvial events in Martian history; (2) our understanding of the volatile and climate history of Mars; and (3) the evaluation and selection of sites relevant to the search for evidence of past life on Mars. While the findings of this research can be most directly applied to understanding north-sloping channel and valley systems to the west and east of the studied system, it is worthy of note that there are numerous Martian systems morphologically analogous to those examined above that are located well outside of Memnonia and adjacent Aeolis. For example, a system located on the southwest flanks of Hadriaca Patera is characterized by a broad, flat-floored inlet channel (which extends upslope in the direction of the Hadriaca caldera), a central terraced crater, and a simple sinuous outlet channel (which extends downslope in the direction of Hellas Planitia) (Figure 20) [see also *Cabrol and Grin*, 2001]. The strong morphological similarities between the two systems implies that both may have formed through igneous processes similar to those discussed above, and further suggests that sites previously interpreted as the ancient locations of Martian lakes systems [e.g., *Forsythe and Blackwelder*, 1998; *Cabrol and Grin*, 2001], as well as other related features previously interpreted as having formed through fluvial processes, should be reevaluated in light of new data and perspectives.

[28] An extrusive igneous alternative to ancient Martian lakes brings into question the paleo-environmental conclusions derived from lacustrine interpretations, and furthermore questions the choice of previously proposed paleolake

Figure 18. Valles Alpis (Alpine Valley) region of the Moon. Features are labeled as follows: (a) Valles Alpis and inner rille and (b) Mare Imbrium. Lunar Orbiter frame V-102-M (oblique).

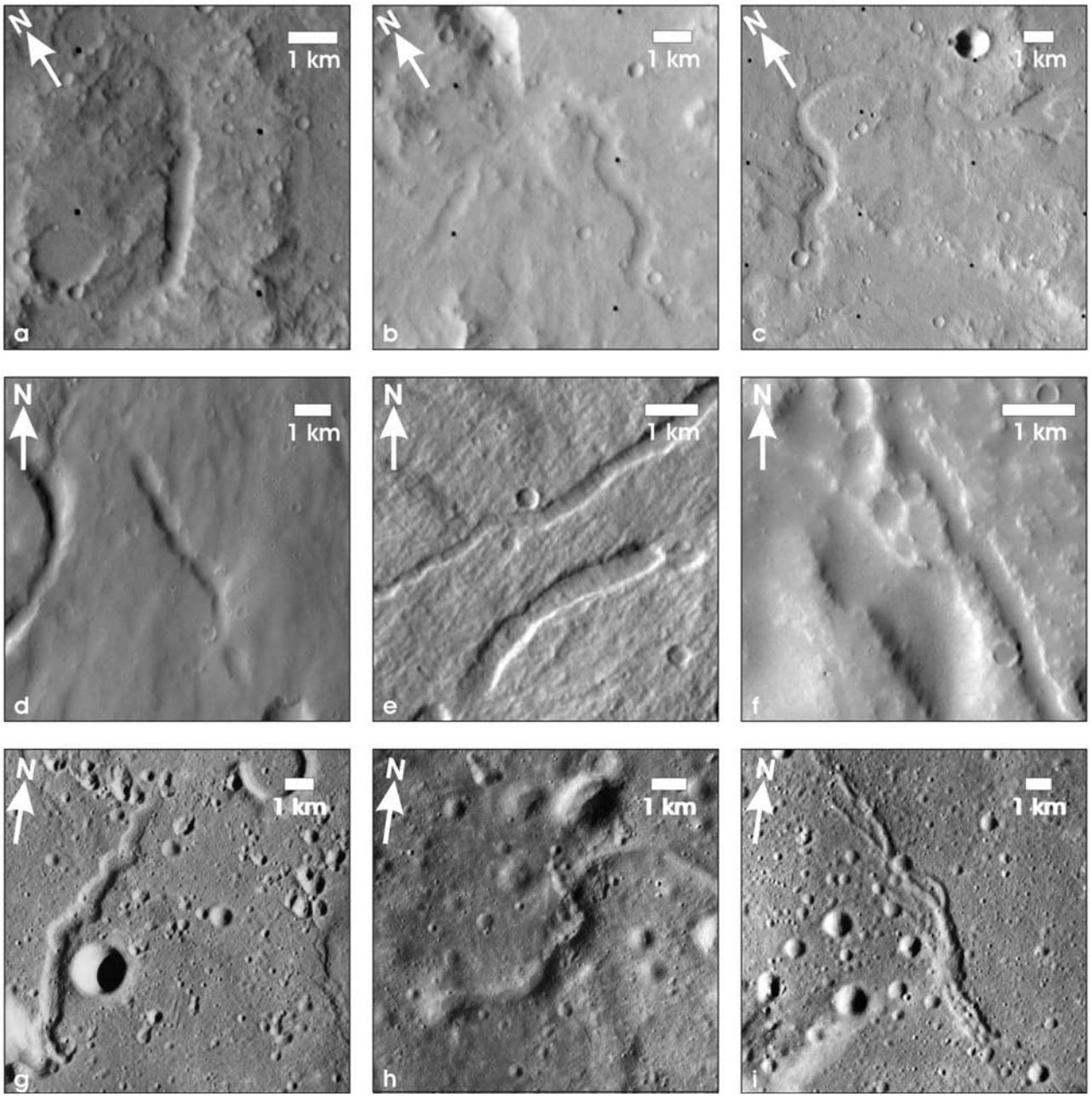


Figure 19. Comparison between the morphology of channels seen in the inlet basin of the Memnonia study region (Viking Orbiter frames (a) 437S14, (b) 438S13, and (c) 437S17) and those on the flanks of Martian volcanoes ((d) Ceranius Tholus, (e) Hecates Tholus, and (f) Ceranius Tholus; Themis frames V02787010, V03128003, V04310006) and volcanic rille features in the Aristarchus Plateau region of the Moon (Apollo 15 Panoramic frames (g) 0326, (h) 0326, and (i) 0328). The forms of the Memnonia channels, which are typically low-order channels characterized by coarsely uniform widths and abrupt heads and terminations, are not distinct from the Martian and lunar volcanic channels.

sites as prime locations for the search for evidence of ancient Martian life [e.g., *Ori et al.*, 2000; *Cabrol and Grin*, 2001]. While it is conceivable that an earlier episode of fluvial sculpture and denudation may have produced the gross topography seen in regions such as western Memnonia, the interpretations presented here regarding later episodes of infilling by lava suggest that at least the uppermost materials of the craters and plains (those that would be sampled by a Mars lander) in such regions may be volcanic in origin rather than sedimentary.

6. Conclusions

[29] A reevaluation of a candidate crater-lake system in western Memnonia suggests that its terraces and channels may not have formed through lacustrine and fluvial pro-

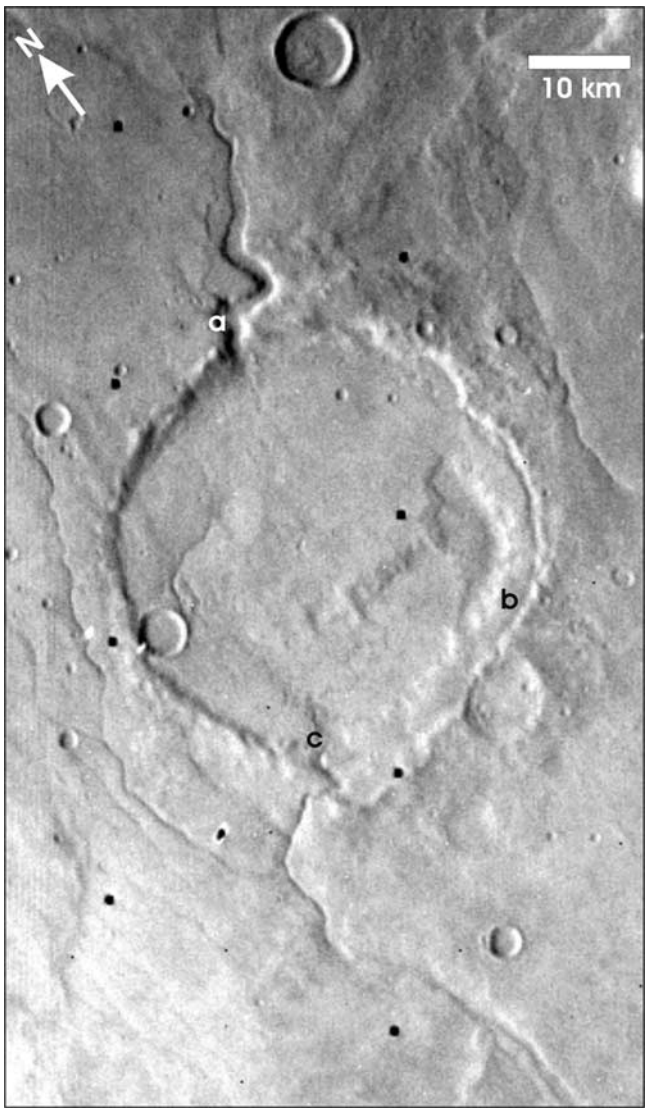


Figure 20. Channel and terrace system on the southwest flank of Hadriaca Patera. Features are labeled as follows: (a) inlet, (b) terrace, and (c) outlet. Compare with Figure 2. Viking Orbiter frame 363S49.

cesses as previously proposed. The limited fetch of the central crater of this system would have likely been insufficient to allow wave action to form its prominent terraces, and the morphological characteristics of inlet and outlet channels to this crater appear inconsistent with formation in association with a crater lake. While a variety of processes must have been responsible for terrain evolution in the study region, relatively recent features such as interior crater fill, terraces, and channels have characteristics and interrelations that are most consistent with formation through igneous processes involving the flow and ponding of lava. Accumulations of material in craters and other topographic lows throughout much of the study region have properties consistent with those of volcanic deposits, and terraces located in a number of craters are interpreted to have formed by drainage or subsidence of these materials. Channels previously identified as fluvial inlets and outlets to crater lakes are interpreted instead as volcanic rilles that carried overflow of effusive igneous products between zones of accumulation at local topographic lows. A regional volcanic setting for the study region lends support to the igneous hypothesis for formation of channel and terrace features. These findings suggest that similar systems located in other regions of Mars should be reexamined regarding their possible igneous origins.

[30] **Acknowledgments.** We thank Gian Ori and Joseph Boyce for their helpful reviews. This work was supported by grant NAG5-13300 from the NASA Mars Data Analysis Program.

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