ARTICLE IN PRESS
GEOMOR-03615; No of Pages 25

Geomorphology xxx (2011) xxx–xxx

Contents lists available at ScienceDirect

Geomorphology

journal homepage: www.elsevier.com/locate/geomorph

Review

A volcanic origin for the outflow channels of Mars: Key evidence and major implications

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A R T I C L E I N F O

Article history:
Received 3 February 2011
Received in revised form 27 May 2011
Accepted 27 May 2011
Available online xxxx

Keywords:
Mars
Channel
Volcanism
Venus
Moon

A B S T R A C T

The outflow channels of Mars are widely believed to have formed through aqueous outbursts from aquifers, implying the past existence of large near-surface water reservoirs and the past operation of a vigorous hydrological cycle. However, accepted mechanisms of channel development suffer from numerous weaknesses, including (1) reliance upon implausible hydrological assumptions; (2) requirement of water abundances that are orders of magnitude greater than can be justified by geochemical considerations; (3) reliance upon long-term conditions that may be incongruous with the preserved mineralogical record; (4) limited correspondence between expected and observed channel properties; and (5) reliance upon exotic aqueous processes for which there are no known solar system analogs. In contrast, channel characteristics are consistent with volcanic origins involving low viscosity lava flows and associated processes of thermal or mechanical erosion. The volcanic hypothesis is founded upon the existence of analog landforms on multiple bodies of the inner solar system and on corresponding analog processes that are variations on familiar terrestrial volcanic themes. Volcanic channel origins are compatible with available mineralogical and geochemical data and are consistent with the nature of preserved channel landforms. The volcanic hypothesis fits within a wider geological framework that economically accounts for the existence and nature of large outflow systems located on the Moon, Venus, and Mars. A volcanic origin reduces the probability that extensive aqueous environments existed during the Hesperian and Amazonian along the outflow channels and in associated terminal basins, and it narrows the possible range of Martian environments once hospitable to life.

Contents

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doi:10.1016/j.geomorph.2011.05.022

Please cite this article as: Leverington, D.W., A volcanic origin for the outflow channels of Mars: Key evidence and major implications, Geomorphology (2011), doi:10.1016/j.geomorph.2011.05.022

1. Introduction

Among the most extraordinary landforms on Mars are large channel systems formed by voluminous fluid flows from the subsurface. These 'outflow channels' generally head abruptly at disturbed terrain, ridged plains, or structural features and extend downslope for distances of up to several thousand kilometers (McCauley et al., 1972; Masurksy, 1973; Milton, 1973; Carr, 1974; Carr et al., 1976; Masursky et al., 1977; Mars Channel Working Group, 1983) (Fig. 1). Typical attributes of outflow channels include low sinuosity, high width-to-depth ratios, the presence of streamlined erosional residuals and complex anastamosing reaches, and the presence of longitudinal grooves or ridges on exposed channel floors (e.g., Milton, 1973; Carr, 1974; Carr et al., 1976; Masursky et al., 1977; Baker, 1982; Mars Channel Working Group, 1983; Carr, 1996). The largest Martian outflow channels head in the southern highlands and extend across the topographic dichotomy boundary to the plains of the northern lowlands (e.g., Masursky et al., 1977). A minority of highland outflow systems terminate in southern basins (e.g., Squyres et al., 1987; Crown et al., 1992; Moore and Wilhelms, 2001), and numerous relatively small systems with the characteristics of larger outflow channels are found on the flanks or peripheries of large volcanoes of Tharsis and Elysium (e.g., Masursky et al., 1977; Mouginis-Mark et al., 1984; Gulick and Baker, 1990: Tanaka et al., 1992; Wilson and Mouginis-Mark, 2003). Formation of the Martian outflow channels appears to have peaked during the Late Hesperian but continued well into the Amazonian (Tanaka, 1986; Baker et al., 1991, 1992a; Head et al., 2001; Burr et al., 2002a).

Numerous possible mechanisms for formation of the Martian outflow channels have been proposed, including the flow of liquid water (e.g., Milton, 1973; Carr, 1974; Masursky et al., 1977), the flow of lava (e.g., Carr, 1974; Schonfeld, 1976, 1977a,b; Cutts et al., 1978; Schonfeld, 1979a,b), aeolian processes (Cutts and Blasius, 1981), the flow of water ice (e.g., Lucchitta et al., 1981; Lucchitta, 1982), the flow of carbon dioxide (e.g., Sagan et al., 1973; Hoffman, 2000), emplacement of debris flows (e.g., McCauley et al., 1972; Nummedal and Prior, 1981; Tanaka, 1999), and crustal extension (Schumm, 1974). In recent decades, the Mars community has mainly favored aqueous interpretations involving catastrophic floods from large subsurface reservoirs (Baker and Milton, 1974; Carr, 1974, 1979; Mars Channel Working Group, 1983; Clifford, 1993; Carr, 2000; Clifford and Parker, 2001; Coleman, 2003; Carr, 2007; Keszthelyi et al., 2007; Leask et al., 2007; Montgomery et al., 2009; Warner et al., 2009; Wilson et al., 2009). Aqueous interpretations of the Martian outflow channels have led to inferences of the past existence of large bodies of standing water at channel mouths (e.g., Baker et al., 1991; Parker et al., 1993; Kargel et al., 1995; Moore et al., ; Gulick et al., 1997; Clifford and Parker, 1999; Baker, 2001; Clifford and Parker, 2001; Ivanov and Head, 2001; Fairén et al., 2003; Baker, 2006; Harrison and Chapman, 2008; Dohm et al., 2009), inferences of major past changes in global climate (Sagan et al., 1973; Sharp and Malin, 1975; Baker, 1979; Toon et al., 1980; Baker et al., 1991; Kasting, 1991; Schaefer,

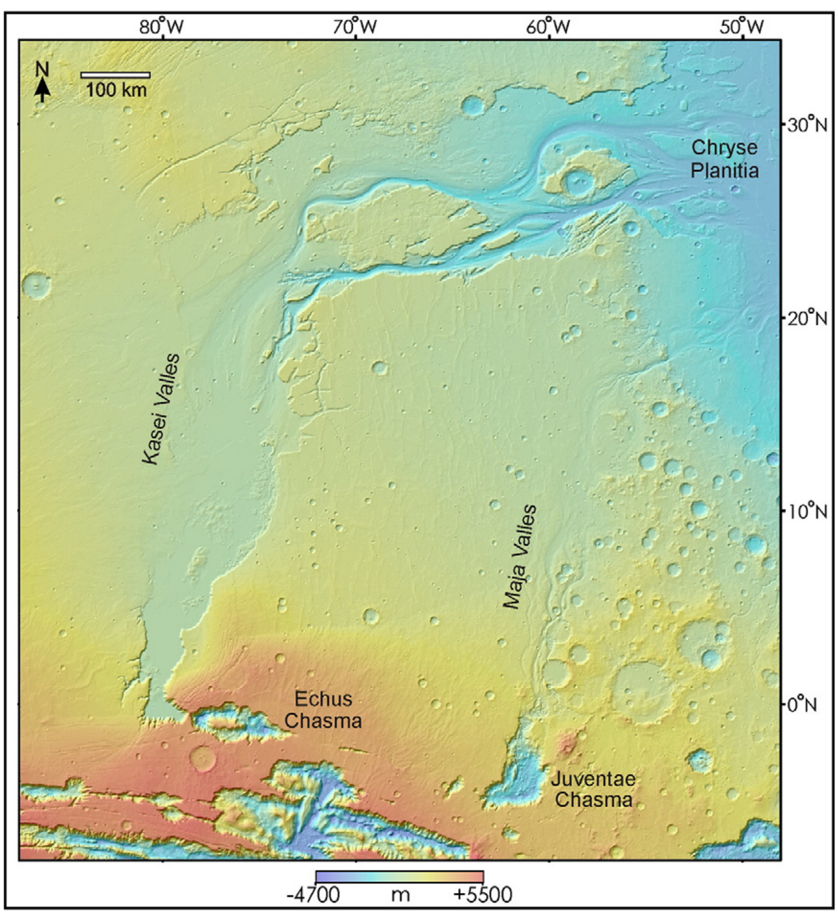

Fig. 1. Among the most prominent outflow systems on Mars are Kasei Valles and Maja Valles. These systems are located north of Valles Marineris in the southwestern part of the circum-Chryse region. Kasei Valles commences at a topographic depression located immediately west of Echus Chasma and extends more than 3000 km downslope, ultimately fading into the ridged plains of Chryse Planitia (e.g., Robinson and Tanaka, 1990; Chapman et al., 2010). The Maja Valles system heads at Juventae Chasma and similarly extends northward into Chryse Planitia (e.g., Baker and Kochel, 1979; Chapman et al., 2003). Mars Orbiter Laser Altimeter (MOLA) elevation data superimposed on shaded relief (topography after Smith et al., 2003).

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1993; Forget and Pierrehumbert, 1997; Carr, 2000; Baker, 2001), and suggestions that early volatile abundances on Mars may have been much higher than implied by geochemical considerations alone (e.g., Carr, 1986, 1987; Clifford, 1993; Carr, 1996, 2000; Clifford and Parker, 2001; McSween et al., 2001). The outflow channels and associated landforms are considered to be among the best available evidence for the past existence of an extensive groundwater system and large surface water bodies on Mars (e.g., Head et al., 2001; Masson et al., 2001).

The results of recent work suggest that aqueous hypotheses for formation of the outflow channels of Mars suffer from numerous weaknesses and that volcanic mechanisms of channel development are worthy of renewed consideration (Leverington, 2004, 2007, 2009). This paper builds on previous efforts by summarizing the most problematic characteristics of widely held aqueous hypotheses, with special emphasis placed on issues including (i) reliance on implausible hydrological assumptions; (ii) inconsistencies between required near-surface water abundances and those suggested by geochemical considerations; (iii) possible incompatibility between surface mineralogy and the required long-term existence of saturated near-surface water reservoirs; (iv) absence of clear fluvial deposits along outflow channels and at terminal basins; and (v) reliance upon the operation of exotic aqueous processes for which there are no known solar system analogs. This paper also outlines the volcanic hypothesis and uses recent advances in our understanding of Martian landforms and surface characteristics to further examine its relative capacity for explaining the existence and nature of the Martian outflow channels. The origin of the Martian outflow channels is of central importance to our understanding of the abundance of near-surface water on Mars, the basic geology and climate history of Mars, the existence and nature of Martian environments once hospitable to life, and possible commonalities among the processes that have shaped the surfaces of the inner solar system. Recognizing the broader significance of the Martian outflow channels, the possible implications of hypothesized volcanic channel origins are also explored in this paper.

2. The aqueous paradigm for formation of the Martian outflow channels

The terrestrial landscapes that most closely resemble the Martian outflow channels are those that have been subjected to catastrophic aqueous floods (e.g., Baker, 1978a,b; Rudoy, 2002; Fisher, 2004; Baker, 2009a,b). The streamlined erosional landforms of the channeled scabland of Washington (Bretz, 1923; Bretz et al., 1956) are particularly reminiscent of the teardrop islands and other scoured features of the Martian outflow systems. Morphological similarities between these terrestrial and Martian channel systems were recognized early in the Mariner-9 and Viking reconnaissance of Mars and quickly led to interpretations of common aqueous origins (e.g., Milton, 1973; Baker and Milton, 1974; Carr, 1974; Masursky et al., 1977; Baker and Kochel, 1979; Carr, 1979). Though alternative channel-forming processes have received consideration in subsequent decades (e.g., Hoffman, 2000; Tanaka et al., 2001, 2002), widespread acceptance of aqueous interpretations of the Martian outflow channels has been maintained to the present (e.g., Baker et al., 1991; Zimbelman et al., 1992; Carr, 1996; Komatsu and Baker, 1997; Baker, 2001; Clifford and Parker, 2001; Wilson et al., 2009; Burr, 2010). Modern hypotheses of aqueous channel development continue to predominantly feature incision by liquid water (e.g., Burr, 2003; Andrews-Hanna and Phillips, 2007; Keszthelyi et al., 2007; Harrison and Grimm, 2008; Montgomery et al., 2009; Warner et al., 2009), though the possibility of associated erosional or depositional contributions by mudflows or debris flows (e.g., Tanaka, 1999; Wilson and Mouginis-Mark, 2003; Williams and Malin, 2004; Rodriguez et al., 2005) or by the flow of glaciers (e.g., Lucchitta, 2001; Head et al., 2004; Pacifici et al., 2009; Chapman et al., 2010) is also incorporated into many models.

Minimum water volumes required for formation of the Martian outflow channels have been estimated by assuming volumetric sediment-to-water ratios of 40:60 (e.g., Komar, 1979,1980; Carr, 1987; Tanaka and Chapman, 1990; Rotto and Tanaka, 1992; Williams et al., 2000; Leask et al., 2006, 2007). For example, the estimated ~4–5×10^6 km^3 of material removed during outflow channel development at the circum-Chryse channels has been used to suggest that a minimum water volume of ~6–7.5×10^6 km^3 must have been involved in channel incision, equivalent to a global ocean ~40–50 m deep (Carr, 1987, 1996, pp. 63 and 165; Baker et al., 1992a, p. 518). Taking into account that it would be highly unlikely for water reservoirs to have existed only at the heads of channels in the circum-Chryse region, this volume has been previously extrapolated across the Martian surface to infer an equivalent water volume of a global ocean ~300–500 m deep (Carr, 1987, 1996, p. 165; Baker et al., 1992a, p. 518; Baker, 2006, p. 140). Because an average sediment load of 40% is unrealistically high, such minimum estimates must be treated with caution (e.g., Carr, 1996; Clifford and Parker, 2001; Ghatan et al., 2005; Kleinhans, 2005; Andrews-Hanna and Phillips, 2007); sediment-water ratios typical of terrestrial streams yield water volumes up to two or three orders of magnitude greater than those estimated under assumptions of high sediment loads (Komar, 1980; Harrison and Grimm, 2004; Ghatan et al., 2005; Kleinhans, 2005; Andrews-Hanna and Phillips, 2007). Some estimates for development of the Martian outflow channels and associated landforms suggest minimum volumes equivalent to a global water body well in excess of 1 km deep (Carr, 1996, p. 166), particularly if features such as thick cryospheric seals are hypothesized and if efficient recycling of water is not assumed (see Section 3 below).

Flow rates associated with hypothesized aqueous floods have been difficult to confidently estimate because key parameters such as flood depth are poorly constrained. Nevertheless, on the basis of expected flow conditions and empirical relations such as the scaled Manning equation (Komar, 1979; Wilson et al., 2004), discharge rates are estimated to have reached ~10^6 to 10^{10} m^3/s (e.g., Carr, 1979; Komar, 1979; Baker, 1982; Robinson and Tanaka, 1990; Baker et al., 1991; Carr, 1996; Komatsu and Baker, 1997; Dohm et al., 2000; Burr, 2003; Head et al., 2003; Andrews-Hanna and Phillips, 2007; Harrison and Grimm, 2008; Mangold et al., 2008; Wilson et al., 2009). The highest of these estimates are orders of magnitude larger than those of known terrestrial floods and are instead comparable to the rates of ocean currents such as the Gulf Stream (Baker, 2001, p. 230, 2009c, p. 25). The corresponding flow velocities of Martian outflow events are predominantly estimated to have been ~4–60 m/s for assumed water depths of ~10–500 m (e.g., Baker et al., 1991; Carr, 1996; Head et al., 2003; Leask et al., 2006; Wilson et al., 2009), but maximum floodwater velocities during the largest events are estimated to have reached or exceeded 100 m/s (e.g., Komatsu and Baker, 1997).

Formation of Martian outflow channels by the sudden release of ponded surface waters has previously been hypothesized for numerous individual systems (e.g., McCauley, 1978; Robinson and Tanaka, 1990; McKenzie and Nimmo, 1999; Chapman and Tanaka, 2002; Harrison and Grimm, 2004, 2008; Harrison and Chapman, 2008). However, regardless of the exact nature of water release, the relatively small volumes of basins at the heads of many systems strongly suggest that floodwaters would have necessarily been derived from much larger subsurface reservoirs (e.g., Carr, 1979, 1996; Clifford and Parker, 2001; Jakosky and Phillips, 2001; Burr et al., 2002a; Andrews-Hanna and Phillips, 2007); in this perspective, areas of disturbance or collapse at the heads of outflow channels mark the locations of outbursts rather than the full lateral dimensions of subsurface reservoirs (Clifford and Parker, 2001; Andrews-Hanna and Phillips, 2007). The high estimated discharge rates of flood events have suggested the need for highly porous and well-connected aquifer systems (Carr, 1979), with required porosities at 0–7 km depth of at least ~5–20% (e.g., Clifford, 1981; MacKinnon and Tanaka, 1989; Hanna and Phillips, 2005) and required permeabilities as great as ~10^{-6} and 10^{-7} m^2 (Head et al., 2003; Montgomery et al., 2009; Wilson et al., 2009). Permeabilities such as these are not typical of aquifers on the Earth, but are believed by some workers to have possibly been produced on Mars by widespread disturbance of the megaregolith by impacts during the heavy bombardment phase (e.g., Wilson et al., 2009). Martian porosities in excess of 40% have been hypothesized for depths of ~2–7 km in regions substantially underlain by fractured and vesicular basalt units (Dohm et al., 2001). Large increases in hydrologic connectivity are additionally hypothesized to have developed in the subsurface as a result of processes such as the dewatering of hydrous salts (Montgomery and Gillespie, 2005; Montgomery et al., 2009), gradual formation of caverns by hydrothermal processes (Rodriguez et al., 2003, 2005), or physical disruption of rock units during outburst events (Carr, 1996).

Though many uncertainties remain (Hanna and Phillips, 2005; Wang et al., 2006), a widely cited mechanism for development of Martian outflow events relies on the early pressurization of aquifers by the downward movement of freezing fronts (e.g., Carr, 1979, 2000; Clifford and Parker, 2001; Ghatan et al., 2005; Basilevsky et al., 2009). Cooling of the Martian megaregolith, possibly linked to early reductions in interior heat flow or changes in surface conditions (Carr, 1979, 1996), is hypothesized to have gradually formed a ~4-km-thick global cryospheric seal that pushed underlying pore waters against low porosity materials at depths of >6 km (Carr, 1979, 2000; Wilson et al., 2009). Periodic breaching of the ice-rich cryosphere is speculated to have been variously caused by igneous intrusions (e.g., Head et al., 2003; Wilson and Head, 2004), tectonic processes (e.g., Tanaka and Chapman, 1990; Tanaka and Clifford, 1993), impact events (Carr, 1979; Tanaka and Clifford, 1993; Carr, 1996), or the thickening of the cryosphere itself (Carr, 1979), ultimately leading to pressure-driven outbursts capable of supplying peak channel discharges up to or exceeding ~10^9 m^3/s (e.g., Carr, 1979; Robinson and Tanaka, 1990; Carr, 1996; Masson et al., 2001; Wilson et al., 2009).

Today, atmospheric temperatures and pressures at the surface of Mars vary from ~140 to 220 K and from ~1 to 14 mbar, respectively, with corresponding ground temperatures ranging from ~160 to 290 K (e.g., Fanale and Cannon, 1979; Carr, 1996; Jakosky and Phillips, 2001; Forget, 2007). Only trace amounts of water are present in the tenuous Martian atmosphere, with the column abundance of atmospheric water mainly ranging from ~1 to 100 precipitable microns (e.g., Jakosky and Farmer, 1982; Jakosky, 1985; Carr, 1996). Although exceptions may exist (Haberle et al., 2001; Rennó et al., 2009), the temperature and pressure conditions that prevail today at the Martian surface should render liquid water unstable at any latitude and should make exposed water ice unstable within ~30° of the equator (e.g., Ingersoll, 1970; Farmer and Doms, 1979; Clifford and Hillel, 1983; Carr, 1996; Jakosky and Phillips, 2001; Bryson et al., 2008). Modern environmental conditions thus inhibit the flow or accumulation of liquid water at the surface, but hypothesized aqueous outflow events are believed by some workers to have been sufficiently rapid and voluminous to have been possible even under conditions similar to those of today (e.g., Wallace and Sagan, 1979; Carr, 1983, 1996). The outflow record of Mars has nevertheless been cited as possible evidence for the past occurrence of warmer and wetter conditions, either because channel formation was triggered by periods of warming or because the channel outbursts themselves contributed to changes in climatic conditions (e.g., Sagan et al., 1973; Toon et al., 1980; Baker et al., 1991; Kasting, 1991; Forget and Pierrehumbert, 1997; Gulick et al., 1997; Baker, 2001; Head et al., 2001; Masson et al., 2001; Pacifici et al., 2009). Development of the outflow channels has been linked by some workers to the dynamics of a global hydrological cycle driven at least partly by internal heat sources. In this view, episodes of aqueous outflow activity lead to formation of short-lived oceans, the release of large stores of H_2O and CO_2 into the atmosphere, and geologically brief periods of greenhouse warming (Baker et al., 1991; Baker, 2001).

Regardless of the climatic conditions under which hypothesized outflow events took place, the enormous volumes required for incision of individual outflow channels are now generally believed to have necessitated the occurrence of multiple flood events (e.g., Baker et al., 1991; Manga, 2004; Andrews-Hanna and Phillips, 2007). For hypotheses involving pressurized aquifers, the termination of each flood event should have been followed by resealing of the cryosphere and recharge of the local aquifer (Carr, 1979; Clifford, 1993; Carr, 1996). Recent work has suggested that rapid decreases in aquifer pressurization during outflow events would have greatly limited the sizes of individual floods, possibly requiring the occurrence of dozens to thousands of relatively small flood events to complete development of individual outflow systems (Manga, 2004; Andrews-Hanna and Phillips, 2007; Harrison and Grimm, 2008). Recharge of southern hemisphere aquifers is envisioned by some workers to have taken place by the freezing of terminal water bodies produced by outflow events, transfer of a proportion of this frozen water to the south polar ice cap through sublimation and subsequent freezing at high latitudes, and recharge of southern aquifers by infiltration and lateral migration of meltwaters produced through basal melting of the southern ice cap (Clifford, 1987,1993; Carr, 2000; Clifford and Parker, 2001). In this view, the processes involved in aquifer recharge are important components of a global hydrological cycle (e.g., Clifford, 1993; Clifford and Parker, 2001). Based on the possible past existence of glaciers near the summits of shield volcanoes, a similar but more localized set of processes has been hypothesized for the Tharsis region (e.g., Russell and Head, 2007).

3. Weaknesses in aqueous interpretations of the Martian outflow channels

Although aqueous interpretations of the Martian outflow channels have received widespread acceptance over recent decades, these perspectives suffer from important deficiencies. The most problematic issues include the following: (i) the hydrological assumptions that underlie aqueous interpretations are not realistic; (ii) correspondence between the outflow channels and terrestrial flood analogs is limited; (iii) the mineralogical characteristics of surface materials may not be consistent with the long-term persistence of planet-wide aquifers and water-saturated cryospheric seals; and (iv) hypothesized mechanisms of outflow channel formation imply Martian volatile contents that are far larger than those previously determined on the basis of independent geochemical considerations.

3.1. Reliance on implausible hydrological assumptions

The minimum total volume of water hypothesized to have been involved in development of the Martian outflow channels is estimated to have been several tens of millions of cubic kilometers, a volume equivalent to a global ocean ~300–500 m deep (Carr, 1987, 1996, p. 165; Baker et al., 1992a, p. 518; Baker, 2006, p. 140). These floodwaters are generally believed to have incised outflow systems during catastrophic outbursts involving enormous subsurface reservoirs (e.g., Carr, 1979,1996; Clifford and Parker, 2001; Jakosky and Phillips, 2001; Burr et al., 2002a; Andrews-Hanna and Phillips, 2007). Unfortunately, the viability of aqueous hypotheses is weakened by the hydrological implications of hypothesized channel-formation mechanisms. Specifically, (i) permeabilities required of hypotheses involving outbursts from aquifers are orders of magnitude larger than can be reasonably assumed for the Martian megaregolith; and (ii) elevations of the heads of the outflow channels do not coincide with those

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expected for aqueous floods driven or otherwise influenced by hydrostatic pressures in extensive and highly permeable aquifers.

3.1.1. Megaregolith permeability

Mean annual discharge at the mouth of the Amazon River is ~200,000 m^3/s, a volume equivalent to ~10% of the river input to terrestrial oceans (Richey et al., 1989). Peak discharge rates of ~10^6–10^7 m^3/s, roughly equivalent to ~5 to 50 Amazon Rivers flowing simultaneously, have been predicted for development of numerous relatively small outflow systems on Mars, including Athabasca Valles and Mangala Valles (e.g., Komar, 1979; Burr et al., 2002a; Head et al., 2003; Leask et al., 2006, 2007; Mangold et al., 2008). Still higher peak discharge rates of ~10^8–10^{10} m^3/s, equivalent to ~500 to 50,000 Amazon Rivers flowing simultaneously, have been predicted for large outflow channels such as Kasei Valles and Ares Vallis (Robinson and Tanaka, 1990; Komatsu and Baker, 1997; Dohm et al., 2000; Baker, 2001; Masson et al., 2001; Harrison and Grimm, 2008) (see summary in Wilson et al., 2009). The limited sizes of basins at the heads of many outflow systems and the abundance of geomorphological evidence indicating flow from the subsurface have favored channel interpretations involving voluminous outbursts from deep aqueous reservoirs (e.g., Carr, 1979, 1996; Clifford and Parker, 2001; Jakosky and Phillips, 2001; Burr et al., 2002a; Andrews-Hanna and Phillips, 2007). Such outbursts imply extraordinarily high permeabilities in the upper ~7 km of the Martian megaregolith (e.g., Carr, 1996; Head et al., 2003; Wilson et al., 2009).

Quantitative models of Martian outburst events have utilized a wide range of estimates for flow rates, porosities, and permeabilities. Typical parameterizations include maximum outburst rates of ~10^6 m^3/s (Komar, 1979; Burr et al., 2002a; Mangold et al., 2008), porosities of ~5–20% at depths of 0–7 km (e.g., Clifford, 1981; MacKinnon and Tanaka, 1989; Hanna and Phillips, 2005) and high but plausible regional permeabilities of ~10^{-15} to 10^{-9} m^2 (e.g., Andrews-Hanna and Phillips, 2007; Harrison and Grimm, 2008). However, recent work has suggested that the permeabilities required to sustain estimated peak discharge rates should instead approach much higher values on the order of ~10^{-7} to 10^{-6} m^2 (Head et al., 2003; Wilson et al., 2009). Though permeabilities of ~10^{-7} m^2 correspond to those expected of highly porous gravels and are therefore strictly feasible, they are also ~7 orders of magnitude larger than are typical at the regional scale for terrestrial aquifers (e.g., Head et al., 2003; Wilson et al., 2009).

Terrestrial flood basalts such as those of the Columbia River plateau and the Deccan plateau are not homogeneous with regard to permeability, with relatively impermeable materials typically forming the bulk of thick and massive units and with relatively permeable materials forming the upper vesicular parts of volcanic flows; the presence of subvertical and horizontal fractures in these materials is an additional influence on permeability (Shelton, 1982; Fisher, 1998; Kulkarni et al., 2000; McGrail et al., 2006). The permeability of terrestrial oceanic crust similarly varies between units comprised of pillow lavas, hyaloclastites, breccias, intrusions, and zones of hydrothermal mineralization (Fisher, 1998). Though the permeabilities of young porous basalts in the near surface can approach or even exceed values of ~10^{-11} to 10^{-10} m^2 (Fontaine et al., 2002), characteristic permeabilities of a terrestrial basaltic basement range from ~10^{-21} to 10^{-15} m^2 (Fisher, 1998). A wider range of permeabilities is possible at regional scales that resolve fluid conduits (e.g., fractures) that are otherwise not represented at the core scale (Garven, 1995; Guéguen et al., 1996; Fisher, 1998; Yang et al., 1998). Core-scale measurements of terrestrial oceanic crust commonly give permeability values of 10^{-22} to 10^{-17} m^2, in situ measurements of oceanic crust commonly give permeabilities of 10^{-18} to 10^{-13} m^2, and numerical models of heat and fluid flow along the crests of ocean ridges suggest near-surface permeabilities of 10^{-16} to 10^{-9} m^2 (Fisher, 1998).

The largest hypothesized outbursts on Mars involve discharges comparable to those of terrestrial ocean currents (Baker, 2001, p. 230, 2009c, p. 25), but on the basis of data such as those summarized above, it is not clear that implied aquifer permeabilities of ~10^{-7} to 10^{-6} m^2 (Head et al., 2003; Wilson et al., 2009) can be reasonably expected for the upper ~7 km of the Martian megaregolith (much smaller values of ~10^{-9} to 10^{-22} m^2 are instead expected). Brecciation and faulting of near-surface materials by impact events has been suggested as a possible mechanism for development of high permeabilities in Martian crust (Clifford, 1981; Clifford and Hillel, 1983; MacKinnon and Tanaka, 1989; Clifford, 1993; Dohm et al., 2001; Hanna and Phillips, 2005; Wilson et al., 2009), but the range of permeabilities typical of terrestrial fault zones (~10^{-22} to 10^{-12} m^2; Smith et al., 1990) suggests that development of fractures may be insufficient to permit requisite outburst rates. In addition, although development of fractures can facilitate water flow (e.g., Kerrich, 1986; Davison and Kozak, 1988; Forster and Evans, 1991), permeability along fractures can also in some cases be reduced through processes such as cataclasis (e.g., Pittman, 1981) and mineralization (e.g., Lowell, 1991; Shaw, 1994). Development of high permeabilities by severe disruption of rock units during outburst events has been proposed (Carr, 1996), but this process cannot account for the high permeabilities that are nevertheless required of aquifers located beyond the spatial limits of disrupted materials (Andrews-Hanna and Phillips, 2007). Dramatic increases in permeability could conceivably have developed as a result of the dewatering of hydrous salts (Montgomery and Gillespie, 2005; Montgomery et al., 2009) or the melting of large volumes of relict or segregated ice (Rodriguez et al., 2003, 2005), but clear evidence for the activity of these processes in outflow channel formation is not presently available.

Inconsistencies between required and expected megaregolith permeabilities suggest that aqueous outbursts are unlikely to have occurred as widely hypothesized. Might aqueous outbursts instead have involved much lower peak discharges? This interpretation would be consistent with recent work suggesting that the volumes and durations of individual aqueous outbursts must have been limited by rapid local decreases in aquifer pressurization following outburst initiation (Manga, 2004; Andrews-Hanna and Phillips, 2007; Harrison and Grimm, 2008), though it would not be congruous with past inferences of flood depths and, depending on flood volumes, could deprive outflow events of the required capacity for catastrophic incision of channels with lengths of up to several thousand kilometers. Could aqueous outbursts have involved sudden releases from pooled surface reservoirs accumulated through much slower rates of outflow from the subsurface? Such a scenario has been proposed for systems including those linked to the Valles Marineris canyon system (e.g., Lucchitta and Ferguson, 1983), but cannot apply to the many outflow channels that lack large head basins (e.g., Sharp and Malin, 1975; Blasius et al., 1978; Carr, 1979, 1996; Clifford and Parker, 2001; Jakosky and Phillips, 2001; Burr et al., 2002a; Andrews-Hanna and Phillips, 2007). Might development of the outflow channels have instead been related to sudden releases from surface reservoirs unrelated to aquifer outbursts, such as pooled glacial meltwaters? Though development of Martian channels by rapid melting of glaciers has been proposed for some systems (e.g., Fassett and Head, 2006; Coleman et al., 2007; Fassett and Head, 2007), alternative non-glacial scenarios also appear viable for these systems (e.g., Leverington, 2009). More importantly, glacial mechanisms are unlikely to be usefully applicable to the large majority of outflow systems that clearly head at zones of fluid outflow from the subsurface (e.g., Carr, 1974, 1979; Mars Channel Working Group, 1983).

3.1.2. Elevations of channel heads

Development of the Martian outflow channels is generally believed to have involved aquifer outbursts caused by disruptions of cryospheric seals by volcanism, tectonism, and/or impacts (e.g., Tanaka and Chapman, 1990; Tanaka and Clifford, 1993; Carr, 1996; Head et al., 2003; Wilson and Head, 2004). The widespread

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distribution of past volcanic activity on Mars is demonstrated by the pervasiveness of preserved volcanic features, including the low elevation flood basalts of Elysium Planitia and Amazonis Planitia (e.g., Keszthelyi et al., 2000), the ridged volcanic plains of the southern highlands (e.g., Frey et al., 1991), the high-elevation volcanic plains of Daedalia Planum and Solis Planum (e.g., Scott and Tanaka, 1980; Caprarelli and Leitsch, 2009), and the many volcanic shields that are distributed across a wide range of Martian elevations (e.g., Mouginis-Mark et al., 1992). The geographic distribution of extensional and compressional tectonic features is similarly wide on Mars, with some of the most prominent features comprising a hemispheric system of graben centered on the Tharsis region (Plescia and Saunders, 1982; Anderson et al., 2001), and with other large sets of structural features distributed across regions such as Elysium and Utopia Planitia (e.g., Hall et al., 1986). Though the largest well-exposed impact features on Mars are associated with the southern highlands, the geological record clearly indicates the past occurrence of impact events across the planet's surface over broad timeframes (e.g., Barlow and Bradley, 1990; Frey et al., 2002; Neukum et al., 2009).

Despite clear evidence on Mars for widespread past disturbance of the near-surface by volcanic, tectonic, and impact processes, the heads of the outflow channels are predominantly found in the southern highlands or on the flanks of large volcanic shields (e.g., Carr, 1974; Mars Channel Working Group, 1983; Gulick and Baker, 1990; Fassett and Head, 2007). Thus, although gradients in hydraulic head should arguably have favored aquifer outbursts at low elevations (Carr, 1979; Clifford and Parker, 2001; Carr, 2002; Hanna and Phillips, 2005), outflows instead took place primarily at high elevations where hydraulic heads should have approached minimums. Indeed, the highest outflow systems on Mars head at elevations that cannot be reconciled with widely cited mechanisms of aquifer recharge involving infiltration beneath the south polar ice cap (e.g., Carr, 2002; Coleman et al., 2007), implying the action here of alternative but no less exotic processes such as those related to the catastrophic melting of ice sheets (Coleman et al., 2007; Leverington, 2009). Located on the flanks of large shields and high in the southern uplands, the heads of the outflow channels of Mars are generally positioned well above the lowlands that otherwise should have been favored as the outburst zones of large and highly permeable aquifers (Carr, 1979; Clifford and Parker, 2001; Carr, 2002; Hanna and Phillips, 2005), suggesting that aqueous mechanisms of channel formation did not operate as a result of outbursts from extensive cryosphere-confined aquifers.

3.2. Limited correspondence between Martian outflow channels and terrestrial analogs

The extent of correspondence between terrestrial diluvial landscapes and those of Martian outflow systems is more limited than is generally perceived. Clear examples of fluvial deposits have not yet been found along the outflow channels of Mars, and associated terminal basins similarly lack obvious examples of delta and shoreline deposits. More fundamentally, the root processes that drove development of hypothesized terrestrial analogs are unlikely to correspond to those that triggered formation of the Martian systems.

3.2.1. Absence of clear examples of fluvial or shoreline deposits

Aqueous interpretations of the Martian outflow channels are partly based on recognized similarities between the characteristics of these systems and those of terrestrial diluvial landscapes. However, direct correspondence between Martian outflow systems and proposed terrestrial analogs is notably limited. For example, although depositional fluvial features such as longitudinal bars, eddy bars, and giant current ripples are widespread in the Channeled Scabland (Bretz et al., 1956; Baker, 1978b, 2009a), no obvious examples of such landforms are known at the Martian outflow channels. Correspondence between the terrestrial and Martian landscapes is instead mostly or completely confined to erosional features such as quadrilateral residuals, hanging valleys, and cataracts (Baker and Milton, 1974; Baker, 1978b; Ghatan et al., 2005; Leverington, 2007). Accumulation of sedimentary flood deposits in the Channeled Scabland took place in a wide range of environments such as those located along the floors of major and minor channel reaches, in the lees of flow obstacles, and in zones that variously experienced slackwater and high turbulence conditions (e.g., Bretz et al., 1956; Benito and O'Connor, 2003). Given the millions of cubic kilometers of material removed during outflow channel development on Mars (Carr, 1987, 1996) and given hypothesized sediment loads of up to 40% (e.g., Komar, 1979, 1980; Carr, 1987; Tanaka and Chapman, 1990; Rotto and Tanaka, 1992; Williams et al., 2000; Leask et al., 2006, 2007), the absence of prominent fluvial deposits analogous to those of the Channeled Scabland is conspicuous. Although candidate fluvial deposits continue to be routinely identified at Martian outflow channels (e.g., possible diluvial boulder trains, giant eddy bars, pendant bars, and giant transverse dunes; Burr et al., 2002b; Burr and Parker, 2006; Carling et al., 2009; Pacifici et al., 2009), unambiguous examples of fluvial deposits remain frustratingly difficult to identify (e.g., Greeley et al., 1977; Baker and Kochel, 1979; Mars Channel Working Group, 1983; Tanaka, 1997; Wilson and Mouginis-Mark, 2003; Ghatan et al., 2005; Burr and Parker, 2006; Leverington, 2007; Carling et al., 2009). Notably, though many uncertainties remain, numerous channel features once believed to be excellent candidates for depositional fluvial units (e.g., landforms at Mangala Valles, Athabasca Valles, and Kasei Valles; Craddock and Greeley, 1994; Burr et al., 2002b; Williams and Malin, 2004; Burr, 2005) have recently been interpreted by some workers instead as units likely formed by scour or by processes related to volcanism (e.g., Ghatan et al., 2005; Leverington, 2007, 2009; Mangold et al., 2009; Jaeger et al., 2010). A major group of diluvial landforms, prominent sedimentary units of fluvial origin, might not exist along the outflow channels of Mars.

The terminal basins of Martian outflow channels similarly seem to lack clear examples of aqueous sedimentary deposits. Large bodies of water are hypothesized to have pooled at the surface of Mars as a consequence of aqueous development of the outflow channels (e.g., Baker et al., 1991; Scott et al., 1992; Parker et al., 1993; Moore et al., 1995; Clifford and Parker, 1999; Baker, 2001; Clifford and Parker, 2001; Ivanov and Head, 2001; Masson et al., 2001; Harrison and Chapman, 2008; Dohm et al., 2009), but strong supporting evidence for the past existence of these water bodies has not yet been identified. For example, the distal parts of the Martian outflow channels lack deltas, and instead typically fade into the extensive plains that characterize most terminal basins (e.g., Greeley et al., 1977; Schonfeld, 1977a; Cutts et al., 1978; Schonfeld, 1979a; Baker et al., 1992a; De Hon, 1992; Carr, 1996; Burr et al., 2002b). Compelling examples of strandline features are similarly lacking for these terminal basins. For example, hypothesized shoreline zones (e.g., Edgett and Parker, 1997; Clifford and Parker, 2001; Webb, 2004) appear to lack convincing examples of sedimentary features such as beaches, barrier ridges, and spits (Malin and Edgett, 1999; Ghatan and Zimbelman, 2006). Some of the most extensive shoreline candidates (e.g., Parker et al., 1989, 1993; Edgett and Parker, 1997) are also problematically characterized by variations in elevation of up to thousands of meters (Head et al., 1998, 1999), though some of this variation could conceivably have been produced through later deformation through processes such as true polar wander (Perron et al., 2007).

Might the nature of hypothesized aqueous outflow environments have been incompatible with development of prominent deltas and shoreline features? The apparent absence of such landforms has been attributed to factors including the rapidity with which hypothesized flood events took place (e.g., Irwin et al., 2004), the influence of low channel slopes on sediment deposition (Ivanov and Head, 2001), and the susceptibility of sedimentary landforms to weathering and erosion (e.g., Ghatan and Zimbelman, 2006). Deposition of terminal sedimentary units by turbidity currents, possibly related to hyperpycnal flow (i.e., deep flow of relatively dense fluids into a standing water body) associated with emplacement of large debris flows, is hypothesized as a means by which development of deltas might have been inhibited in favor of wider sediment distribution (Tanaka, 1999; Komatsu and Ori, 2000; Ivanov and Head, 2001). Might units such as the Vastitas Borealis Formation, a deposit that mantles parts of the northern plains and is interpreted to have formed under marine conditions (e.g., Fuller and Head, 2002; Kreslavsky and Head, 2002; Baker, 2006), have accumulated through the action of such mechanisms? Emplacement of sedimentary units over wide areas is considered plausible by some workers if, for example, fine sediments were hyperconcentrated in outflow waters (e.g., Kleinhans, 2005). It is not clear, however, that marine interpretations of units such as the Vastitas Borealis Formation are consistent with the ubiquitous association of these units with meter-scale boulders (McEwen et al., 2007).

The mouths of several major outflow systems were directly investigated by the Viking 1, Mars Pathfinder, and Spirit spacecraft. The Viking 1 landing site is located on the basaltic and wrinkle-ridged plains of southwestern Chryse Planitia, near the mouths of Kasei Valles and Maja Valles (Carr et al., 1976; Greeley et al., 1977). This site is characterized by aeolian drifts (Mutch et al., 1976) and rocks with a size-frequency distribution similar to that of the Surveyor 7 landing site on the Moon (Moore et al., 1977). Large angular rocks present at this site are interpreted as probable impact debris (Greeley et al., 1977). Also located in Chryse Planitia, the Mars Pathfinder landing site is situated in the southern part of the basin near the confluence of the mouths of Ares Vallis and Tiu Vallis (Nelson and Greeley, 1999). The Pathfinder landing site broadly resembles the Viking 1 landing site (Rover Team, 1997) and is similarly characterized by abundant aeolian features (Greeley et al., 1999) and rocks of a wide range of sizes (Golombek et al., 1999a). Most rock spectra collected at the Pathfinder site are consistent with mafic, and possibly intermediate, rock compositions (McSween et al., 1999). The subrounded, perched, or imbricated character of some rocks at the Pathfinder site, and the possible presence of conglomerates, has been interpreted as consistent with emplacement of site materials by catastrophic floods (Rover Team, 1997; Golombek et al., 1999ab; Ward et al., 1999). However, as with the Viking 1 site, fluvial interpretations are not considered conclusive (e.g., Chapman and Kargel, 1999; Tanaka, 1999; Ward et al., 1999). The Spirit landing site is located at the mouth of Ma'adim Vallis and is also characterized by an abundance of aeolian features and rocky materials, with plains rocks mainly comprised of primitive olivine-bearing basalts (Arvidson et al., 2006). Though associated Columbia Hills materials have been aqueously altered and are relatively enriched in phases such as goethite, hematite, and nanophase iron oxides (Arvidson et al., 2006), the preserved mineralogy of surface materials in the region clearly indicates that physical weathering processes have generally dominated over chemical processes (Morris et al., 2004). Importantly, none of the characteristics of the Viking 1, Mars Pathfinder, and Spirit landing sites require development through catastrophic aqueous flooding (e.g., Arvidson et al., 1989; Chapman and Kargel, 1999; Ward et al., 1999). Indeed, the basic attributes of all three sites are consistent with those expected of volcanic plains subjected to limited chemical alteration and substantial disruption by impacts (Fig. 2).

3.2.2. Lack of correspondence in root processes of system development

Landforms of terrestrial diluvial systems such as those of the Channeled Scabland and Mongolian Plateau have previously been appropriately cited as candidate analogs for landforms of the Martian outflow channels (e.g., Baker and Milton, 1974; Sharp and Malin, 1975; Komatsu et al., 2004; Andrews-Hanna and Phillips, 2007; Baker, 2009a,b). However, the features of interest of these terrestrial systems were formed by releases from dammed lakes rather than by catastrophic outbursts from aquifers (e.g., Baker and Bunker, 1985; Baker et al., 1993; Komatsu et al., 2009). This is a critical distinction that, for most Mars hypotheses, necessarily limits the analog value of terrestrial processes to those that operated downstream of source regions (Leverington, 2007, 2009). Though few workers would contest this basic observation, its significance in constraining the explanatory power of terrestrial flood analogs is worth emphasizing. Fundamentally, the triggers that forced development of proposed terrestrial analogs have limited potential to serve as analog processes for those hypothesized to have driven catastrophic outbursts from large Martian aquifers.

Are there viable solar system analogs for the aqueous processes widely hypothesized to have led to development of the Martian outflow channels? Among the best available candidates may be those linked to development of large terrestrial springs and mud volcanoes. For example, aquifer circulation driven by hydraulic head and influenced by inclined strata has led to development of the Dalhousie Mound Spring complex of Australia, one of the largest groundwater discharge landforms on Earth (Clarke and Stoker, 2003; Clarke et al., 2007). Terrestrial mud volcanoes, in contrast, commonly form in association with mud diapirs in regions of relatively high petroleum potential, with expulsion of mud and hydrocarbons typically forming crater-like features, hummocky mud flows, and small cones (e.g., Milkov, 2000). Terrestrial mud volcanoes mainly form in marine environments, but prominent among subaerial examples is a feature that formed in Java as a result of drilling-induced processes related to petroleum exploration operations (Mazzini et al., 2007; Davies et al., 2008). Unfortunately, the availability of terrestrial analogs such as these has not led to major advances in our understanding of outflow systems on Mars, in part because associated processes are inconsistent with the volumes and rates of flow inferred to have been required in development of the Martian systems. The processes that drive groundwater flow at terrestrial features such as springs and mud volcanoes are subject to the same permeability-related limitations that affect all other candidate processes involving the movement of fluids through porous media, and it is therefore not surprising that the average discharge rates produced at associated landforms are commonly no greater than several cubic meters per second (e.g., Clarke and Stoker, 2003; Mazzini et al., 2007).

3.3. Inconsistency between Mars surface mineralogy and proposed aqueous outflow models

Though notable exceptions exist (Fishbaugh et al., 2007; Milliken et al., 2008; Mangold et al., 2010), hydrated minerals exposed at the surface of Mars (including varieties of phyllosilicates and sulfates) are predominantly associated with materials altered during the Noachian or earlier parts of the Hesperian, suggesting that widespread aqueous conditions were mostly confined to the earliest stages of Mars history (Bibring et al., 2005, 2006). The persistence of olivine-rich units (~20–35% olivine) exposed during the Noachian — and the expected susceptibility of olivine to aqueous alteration at timescales of no more than tens of thousands of years (Stopar et al., 2006) — confirms that even early hydrous conditions on Mars should have been of relatively limited spatial and temporal extent (Hoefen et al., 2003; Rogers et al., 2005; Bibring et al., 2006; Koeppen and Hamilton, 2008). Apparent preservation of exposed carbonate units of Noachian age (Ehlmann et al., 2008) suggests that later acidic conditions of the early Hesperian were not globally pervasive (Murchie et al., 2009). More significantly, the high solubility of sulfate minerals on Mars is consistent with the maintenance of relatively dry conditions since formation (King et al., 2004; King and McLennan, 2010), and the incomplete diagenesis of

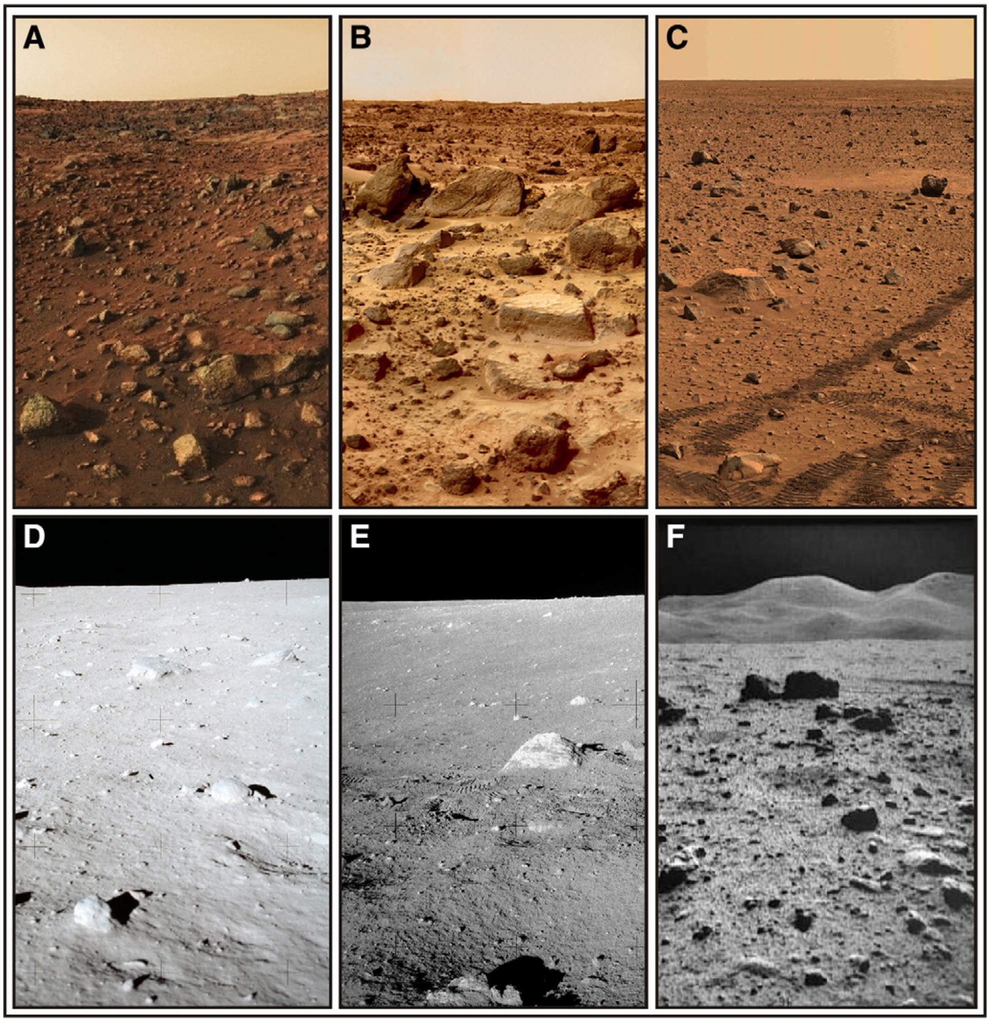

Fig. 2. The mouths of outflow systems on Mars have been visited by three spacecraft: Viking 1 landed at a site in Chryse Planitia near the mouths of Kasei Valles and Maja Valles (A), Mars Pathfinder landed at a site in Chryse Planitia near the mouths of Ares Vallis and Tiu Vallis (B), and the Spirit rover landed at Gusev crater near the mouth of Ma'adim Vallis (C). Though widely hypothesized as the sites of catastrophic outflow events (e.g., Rover Team, 1997), the basic attributes of all three sites are arguably consistent with those expected of volcanic plains subjected to limited chemical alteration and substantial disruption by impacts (e.g., Arvidson et al., 1989). The disrupted nature of the surface units of these sites is analogous to that of lunar volcanic plains visited during the Apollo program: Apollo 11 at Mare Tranquillitatis (D), Apollo 12 at Mare Cognitum (Oceanus Procellarum) (E), and Apollo 15 at Palus Putredinus (Mare Imbrium) (F). Apollo image numbers: AS11-40-5940; AS12-48-7047; AS15-82-11175, respectively.

early sulfate-bearing units at Meridiani Planum (McLennan et al., 2005; Squyres and Knoll, 2005) is considered incompatible with the existence of wet conditions at this site beyond the time of original alteration (Fairén et al., 2009; Tosca and Knoll, 2009). The long-term dry conditions implied by Martian bedrock units are congruous with the high concentrations of olivine measured in ancient aeolian sediments, materials that should be especially susceptible to aqueous weathering (Goetz et al., 2005; Yen et al., 2005; Morris et al., 2006; Mustard et al., 2007). The global mineralogical record suggests that the role of liquid water in Martian weathering processes has been especially limited over the past ~3.5 billion years (Hoefen et al., 2003; Bibring et al., 2006; Koeppen and Hamilton, 2008).

Development of the Martian outflow channels mainly took place during the Late Hesperian and Amazonian (Tanaka, 1986; Baker et al., 1991, 1992a; Head et al., 2001), postdating the early period of widespread aqueous alteration on Mars. Beyond scattered associations with ancient altered materials (e.g., Michalski and Noe Dobrea, 2007), a clear and consistent spatial correlation between Martian outflow channels and exposures of hydrated minerals has not been recognized (Bibring et al., 2006; Mangold et al., 2007, 2008). Instead, perhaps the best aqueous case that can be made from available mineralogical data is that some Hesperian and Amazonian outflow channel systems cut across or head at much older highland units that show evidence for aqueous alteration in the Noachian and earliest Hesperian. Importantly, the olivine-rich units that are distributed across the planet's southern hemisphere are prominently exposed at large outflow channels such as those of the circum-Chryse region (Rogers et al., 2005; Koeppen and Hamilton, 2008) (Fig. 3).

The absence of a notable record of hydration along the Martian outflow channels and the association of large channel systems with ancient bedrock exposures composed of ~20 to 35% olivine and showing little or no evidence for aqueous alteration, together appear to argue against the long-term persistence of near-surface water along the outflow channels of Mars. Thus, although aqueous outflow hypotheses cannot be conclusively ruled out on the basis of mineralogical considerations alone, models involving periodic recurrence of relatively warm and wet conditions sustained over timeframes of ~10^3 to 10^5 years (e.g., Baker et al., 1991; Baker, 2001) do not appear to be supported by the nature of near-surface mineralogy as presently understood. In addition, although models involving discrete aqueous outburst events with timeframes of weeks to years and occurrence under otherwise cold and dry conditions (e.g., Zimbelman et al., 1992; De Hon and Pani, 1993) might be expected to not require mineralogical alteration along the Martian outflow

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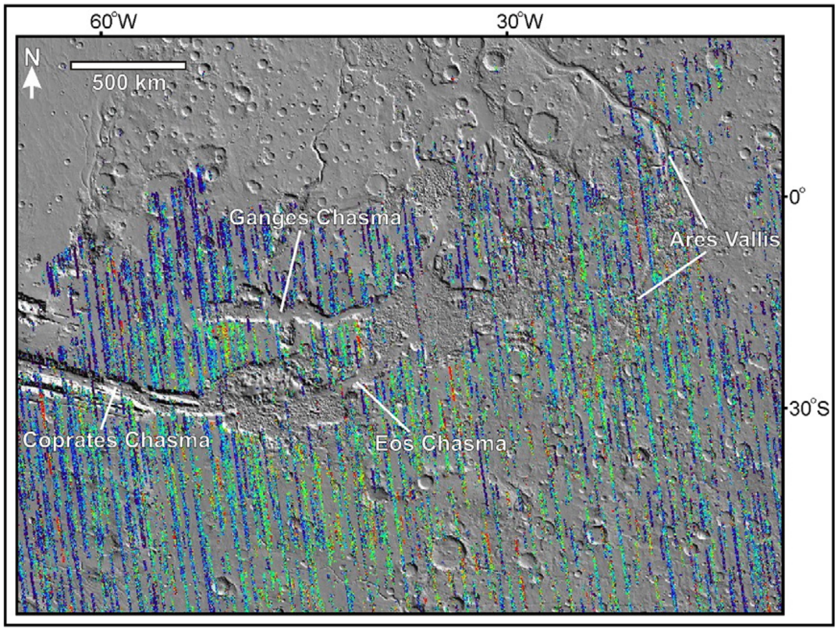

Fig. 3. Olivine-rich units that are distributed across the southern hemisphere of Mars are prominently exposed at, and in the vicinity of, large outflow channels such as Ares Vallis (Rogers et al., 2005; Koeppen and Hamilton, 2008). In this map of the eastern Valles Marineris region, total olivine abundance for well-exposed sites is plotted on a map of shaded relief. Estimated total olivine abundances range from low values of ~1 to 7% (blue) to moderate values of ~10 to 20% (green to yellow) to high values of ~30% (red). Abundance data were plotted using the Java Mission-planning and Analysis for Remote Sensing (JMARS) system and were originally generated from Thermal Emission Spectrometer (TES) data by Koeppen and Hamilton (2008). Data are not available for sites with albedo values >0.2 (i.e., sites with substantial dust cover). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

channels, many such models nevertheless rely on the long-term presence of large Martian aquifers and water-saturated cryospheric seals, both of which might be expected to have caused widespread aqueous alteration of near-surface materials later exposed through channel incision. In addition, evaporation or sublimation of hypothesized floodwaters might, for even the most rapid of aqueous flood events, be expected to have led to emplacement of detectable evaporite or other mineralogical mantles along outflow channels and at terminal basins.

3.4. Contradictions in water inventories

From a theoretical standpoint, virtually all of the original water content of Mars should have reacted with iron and other metallic phases to produce oxides and large quantities of hydrogen gas (e.g., Dreibus and Wänke, 1985, 1987). Today, the bulk water content of the planet should consist of the vestiges of this early water, along with water contributed through processes such as the surface impact of volatile-rich bodies (e.g., Anders and Owen, 1977). The largest confirmed modern reservoirs of water on Mars are the perennial ice caps and the near-surface stores that are distributed across middle to high latitudes (e.g., Kiefer et al., 1976; Clifford and Hillel, 1983; Smith et al., 1999; Seibert and Kargel, 2001; Boynton et al., 2002; Feldman et al., 2003; Mangold, 2005; Smith et al., 2009; Dundas and Byrne, 2010; Zanetti et al., 2010). Although the minimum total volume of these reservoirs is estimated to be ~3×10^6 km^3 (Christensen, 2006), the modern Martian surface environment is extraordinarily dry. Liquid water is not stable anywhere at the planet's surface, and exposed water ice is unstable at middle to low latitudes (e.g., Ingersoll, 1970; Farmer and Doms, 1979; Clifford and Hillel, 1983; Carr, 1996; Jakosky and Phillips, 2001).

The shergottite, nakhlite, and chassignite (SNC) meteorites are basalts or basaltic cumulates that are believed to have originated as subsurface components of the Martian crust and mantle, and therefore represent direct sources of information regarding the geochemistry of Mars (e.g., McSween, 1994; Treiman et al., 2000; Halliday et al., 2001). The shergottites are notable among the SNC meteorites for possessing reflectance properties similar to those that characterize much of the Martian surface (e.g., McSween, 1994; Mustard et al., 1997, 2005). As a group, the SNC meteorites are generally very dry; and although limited aqueous alteration is implied by the presence of minerals such as amphibole and gypsum (e.g., Bunch and Reid, 1975; Gooding et al., 1990; Karlsson et al., 1992; McSween and Harvey, 1993; McSween, 1994; Treiman et al., 2000; Bridges et al., 2001), many of the shergottites show evidence for having formed in essentially water-free environments (e.g., Carr, 1986; Wanke and Dreibus, 1994; Norman, 1999; Kiefer, 2003). Typical bulk water contents of the SNC meteorites are only ~130 to 350 ppm, whereas terrestrial oceanic basalts commonly fall within the range of 1500–6000 ppm (Carr and Wänke, 1992; McSween and Harvey, 1993; Carr, 1996).

If the magmas that formed the SNC meteorites did not lose substantial amounts of water during ascent toward the surface (McSween et al., 2001), their geochemical characteristics suggest a low abundance of water on Mars relative to the Earth (McSween, 1994; Wanke and Dreibus, 1994). On the basis of the compositions of the SNC meteorites and the modern Martian atmosphere, the early water content of Mars is estimated to have been equivalent to a global layer of ~6 to 200 m thickness, only a proportion of which would have been outgassed to the near-surface environment (Anders and Owen, 1977; Pollack and Black, 1979; Squyres, 1984; Dreibus and Wänke, 1985; Carr, 1987; Dreibus and Wänke, 1987; Owen et al., 1988; McSween and Harvey, 1993; McSween, 1994). The model of Scambos and Jakosky (1990) suggests a much wider range of possible Martian bulk water abundances (~ 90 to 3000 m), but restricts the amounts of outgassed water to an equivalent layer of only tens of meters thickness. The model of Lunine et al. (2003) suggests a similarly wide range in possible bulk water content (~600 to 2700 m), but proposes that a much higher proportion of water was outgassed (~50%). Recognized modern water reservoirs of Mars (Smith et al., 1999; Masson et al., 2001; Christensen, 2006) are approximately equivalent to a global layer with a thickness of several tens of meters,

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and thus have a total volume that is consistent with many geochemical predictions of near-surface water abundance. However, direct comparisons between early bulk abundance and modern near-surface abundance are complicated by factors such as the expected partial loss of near-surface water by atmospheric escape (e.g., Shizgal and Blackmore, 1986; Jakosky et al., 1994; Leshin, 2000; Lammer et al., 2003).

The possible incompatibility between the low water abundances inferred from most geochemical models and the high abundances required of aqueous channel interpretations is a long-standing issue in the study of Mars (Carr and Wänke, 1992; Wänke and Dreibus, 1994; Carr, 1996; Beaty et al., 2005). The minimum volume of water necessary for development of the Martian outflow channels, determined under assumptions of 40% sediment content, is equivalent to a global water layer of ~300–500 m thickness (Carr, 1987, 1996, p. 165; Baker et al., 1992a, p. 518; Baker, 2006, p. 140). Sediment loads more typical of terrestrial fluvial systems suggest water volumes up to two to three orders of magnitude greater than these estimates (e.g., Komar, 1980; Harrison and Grimm, 2004; Ghatan et al., 2005; Kleinhans, 2005; Andrews-Hanna and Phillips, 2007), implying equivalent water layers of at least tens of kilometers thickness if water recycling is unrealistically assumed to have not occurred. Because water in the regions of hypothesized aqueous outbursts cannot be assumed to have been recycled with full efficiency, aqueous interpretations of the outflow channels must also account for the existence of significant groundwater volumes never expelled to the surface and therefore never directly involved in outflow channel incision (e.g., Scott et al., 1991; Baker et al., 1992a). Thus, even under the reasonable assumption of the past operation of an active hydrological cycle capable of repeated recharge of hypothesized reservoirs, outflow processes appear to require volumes equivalent to a global layer of at least several kilometers thickness (even assuming high sediment loads of ~5 to 10%). Beyond this volume is the water required to maintain a cryospheric cap of global extent and ~4 km thickness (e.g., Wilson et al., 2009), equivalent to a global water layer of up to ~250 to 1000 m thickness if commonly cited megaregolith porosities of 10–40% are assumed and if pores are largely (~60%) filled with water in the solid or liquid phases. Consideration of the various reservoirs required for hypothesized aqueous development of the outflow channels therefore suggests a total water volume potentially far in excess of the 300–500 m commonly cited, and conceivably much larger than that of the modern near-surface reservoirs of the Earth (equivalent to a global layer of ~2700 m thickness; Carr and Wänke, 1992; Beaty et al., 2005).

Can geochemical estimates of Martian water abundances be reconciled with the near-surface volumes implied by aqueous interpretations of the Martian outflow channels? The properties of the SNC meteorites and modern Martian atmosphere do not impose firm constraints upon the early volumes of near-surface water reservoirs (McSween and Harvey, 1993), and numerous possibilities correspondingly exist for resolution. For example, the relatively low internal pressures of Mars should cause magmas to be especially susceptible to devolatization, and the amount of water outgassed by the planet may therefore have been larger than is otherwise implied by the SNC meteorites (McSween and Harvey, 1993; McSween et al., 2001). Also, a larger volume of water could have been degassed to the Martian surface if the reaction between water and iron during core formation was less complete than is widely assumed (McSween and Harvey, 1993). Moreover, a late influx of water to the planet's surface by impacts of volatile-rich bodies could have supplemented surface reservoirs without affecting the compositions of mantle-derived materials (Anders and Owen, 1977; Chyba, 1987; Carr and Wänke, 1992). Notably, despite the validity of these considerations, the low estimates of early near-surface water abundance derived from geochemical measurements are consistent with both the low volumes of modern water reservoirs and available mineralogical data that suggest the occurrence of remarkably dry conditions over most of the planet's history (e.g., Hoefen et al., 2003; Bibring et al., 2005, 2006; Forget, 2007; Koeppen and Hamilton, 2008).

4. The volcanic hypothesis

Although aqueous diluvial hypotheses for outflow channel formation on Mars have consistently received widespread acceptance over the past three decades (e.g., McCauley et al., 1972; Milton, 1973; Baker and Milton, 1974; Carr, 1974; Masursky et al., 1977; Baker and Kochel, 1979; Carr, 1979; Mars Channel Working Group, 1983; Baker et al., 1991; Zimbelman et al., 1992; Carr, 1996; Komatsu and Baker, 1997; Baker, 2001; Clifford and Parker, 2001; Baker, 2004, 2006; Wilson et al., 2009; Burr, 2010), several other candidate mechanisms have been considered. Incision by CO_2-supported density flows (Hoffman, 2000) was assessed as a process of channel incision but was ultimately rejected by most workers as unrealistic and incapable of fully accounting for channel properties (e.g., Coleman, 2003; Baker, 2004, 2006). Glacial and debris flow mechanisms (e.g., Lucchitta et al., 1981; Nummedal and Prior, 1981; Lucchitta, 1982) continue to be proposed, though they are most commonly hypothesized as processes that operated in conjunction with aqueous floods (e.g., Baker, 2001; Costard and Baker, 2001; Wilson and Mouginis-Mark, 2003; Head et al., 2004; Williams and Malin, 2004; Rodriguez et al., 2005; Pacifici et al., 2009; Chapman et al., 2010).

Volcanic hypotheses for outflow channel development were considered in the earliest stages of the spacecraft-based exploration of Mars (Carr, 1974; Schonfeld, 1976, 1977a,b; Cutts et al., 1978; Schonfeld, 1979a,b), but failed to receive broad acceptance (e.g., Baker, 1978a; Mars Channel Working Group, 1983; Baker et al., 1992a). The case for volcanic mechanisms was undermined in early investigations by a general perception that water has a unique capacity to form conduits possessing the various attributes of the Martian channels, including both the outflow channels and the Noachian valley networks (e.g., Milton, 1973; Sagan et al., 1973; Sharp and Malin, 1975). In particular, the flow of lava was rejected as a likely means by which inner channels, very wide channels, and anastamosing reaches could have developed at outflow systems (e.g., Carr, 1974; Baker, 1978a,b,c, 1982; Baker et al., 1992a). Volcanic mechanisms for outflow channel development were also considered inconsistent with the absence of distributary channels in the distal reaches of outflow systems and with the presumed absence of extensive lava deposits at terminal basins (e.g., Sharp and Malin, 1975; Baker, 1982; Baker et al., 1992a). The presence of apparent scabland analogs such as longitudinal grooves and scour marks was considered especially suggestive of the past action of aqueous scour at the outflow channels (Baker, 1978a; Baker et al., 1992a), strengthening support for aqueous hypotheses. Difficulties in identification of viable aqueous mechanisms for development of hypothesized floods were acknowledged but were minimized by noting that similar difficulties had plagued early investigations of the Channeled Scabland (Baker, 1978c, 1979, 1982), ultimately resulting in broad acceptance of aqueous hypotheses as the most realistic of available candidate mechanisms.

Recent work has revived consideration of the volcanic hypothesis for development of the Martian outflow channels (Leverington, 2004, 2007, 2009; Jaeger et al., 2010). Although aqueous hypotheses for outflow channel formation are widely accepted today, the volcanic hypothesis is arguably most consistent with the landforms of Martian systems and with the outflow processes most likely to have operated on Mars; specifically (i) although analog processes are not known for development of large channel systems by aqueous outbursts from aquifers, analogs exist on multiple solar system bodies for volcanic development of large outflow systems through effusion of low viscosity magma from the subsurface; and (ii) although the Martian outflow channels lack obvious examples of aqueous sedimentary deposits along channels and at terminal basins, these channels and

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basins possess distinct volcanotectonic affinities that are comparable with those of volcanic analogs and are associated with landforms of likely volcanic origin from channel heads to terminal basins.

4.1. Candidate analog landforms and processes

Volcanic processes operating in modern terrestrial environments predominantly develop landforms in a constructive manner. However, the flow of lava has a capacity for thermal or mechanical erosion if suitable eruptive conditions exist (e.g., Hulme, 1973; Peterson and Swanson, 1974; Hulme and Fielder, 1977; Hulme, 1982; Baird and Clark, 1984; Huppert et al., 1984; Bussey et al., 1995; Greeley et al., 1998; Kauahikaua et al., 1998; Williams et al., 1998; Griffiths, 2000; Fagents and Greeley, 2001; Kerr, 2001; Williams et al., 2001; Wilson and Mouginis-Mark, 2001; Schenk and Williams, 2004; Williams et al., 2004, 2005; Ferlito and Siewert, 2006; Kervyn et al., 2008; Siewert and Ferlito, 2008; Kerr, 2009; Williams et al., 2011). Parameters that strongly influence the capacity of flows to erode substrates include lava properties (composition, temperature, viscosity, yield strength, volatile content, and crystal content), magnitude of flow turbulence, rate of effusion, total flow volume, presence and nature of insulating flow cover, flow gradient, and nature of substrate (e.g., see summaries in Hulme, 1973; Huppert et al., 1984; Williams et al., 1998; Leverington, 2007; Jaeger et al., 2010; Hurwitz et al., 2010). In general, the capacity for incision increases with higher lava temperatures, lower lava viscocities, greater levels of turbulence, and greater rates and volumes of effusion (e.g., Hulme, 1973; Pinkerton et al., 1975; Hulme and Fielder, 1977; Hulme, 1982; Bussey et al., 1995; Williams et al., 1998). Mafic and ultramafic magma compositions are among those most commonly associated with such flow properties and conditions (e.g., Murase and McBirney, 1970; Shaw and Swanson, 1970; Hulme, 1973; Murase and McBirney, 1973; Hulme, 1974,1982; Huppert et al., 1984; Head and Wilson, 1992; Komatsu et al., 1992; Reidel, 1998; Williams et al., 1998; Griffiths, 2000; Keszthelyi et al., 2006).

Numerous examples exist of lunar and Venusian channels that appear to have developed as a result of voluminous effusion of low viscosity lavas (e.g., Oberbeck et al., 1969; Greeley, 1971a; Cruikshank and Wood, 1972; Schultz, 1976; Wilhelms, 1987; Baker et al., 1992b, 1997) (Figs. 4 and 5). In recognition of their outflow origins, these systems are referred to here as 'outflow channels.' Though alternative interpretations have previously been proposed for these channels

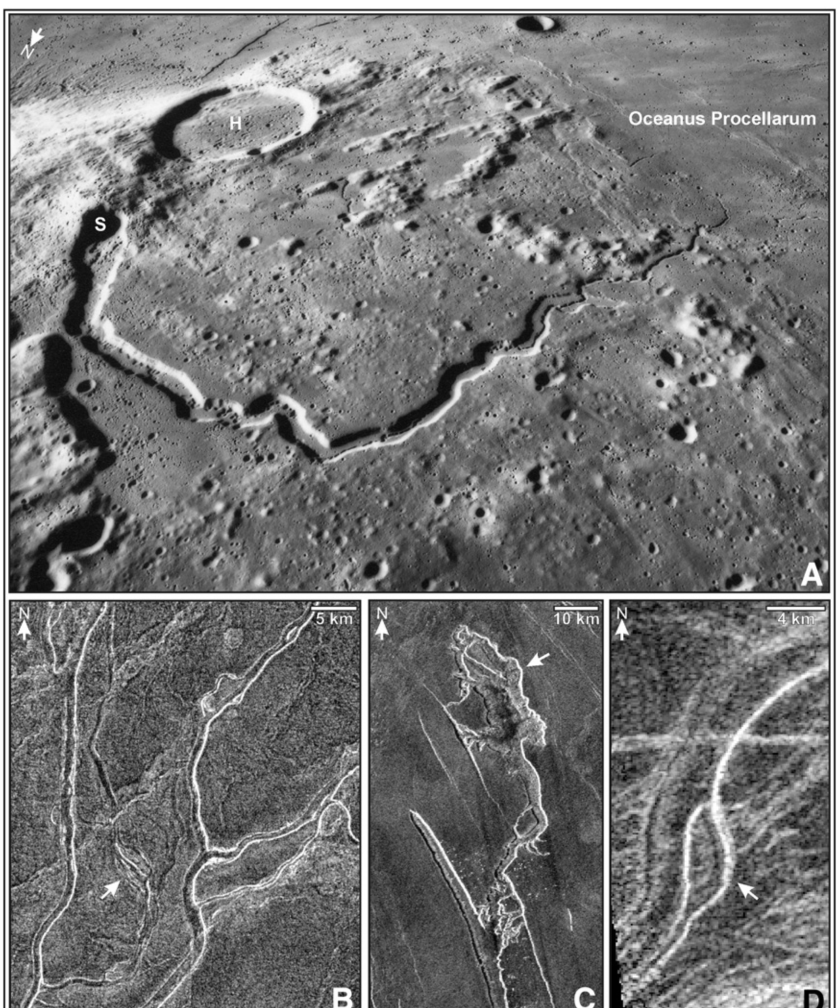

Fig. 4. Examples of lunar and Venusian channels believed to have formed by the flow of lava. (A) Oblique view of Vallis Schröteri, the largest (though not the longest) outflow system on the Moon; the head depression of this system (S) is located near crater Herodotus (H), which has a diameter of 35 km. (B) An example of a branching Venusian channel system (Komatsu et al., 1993); streamlined residuals are common here, as are small reaches of complex anastamosing character (arrow). (C) The head depression (arrow) of the 1200-km-long Venusian outflow channel Kallistos Vallis; the channel continues southward from this depression, ultimately forming complex anastamosing reaches with widths of ~30 km (Baker et al., 1992b); Fig. 5A. (D) Streamlined landform (arrow) associated with a Venusian channel system. Figure centers: (A) 25.5° N., 51.5° W.; (B) 5° N., 52.95° E.; (C) 48° S., 19.25° E.; and (D) 30° S., 105° E. Image numbers: (A) Apollo 15 Metric camera image AS15-M-2612; (B)–(D) Magellan full-resolution radar map (FMAP) left-look synthetic aperture radar mosaics; radar illumination is from the left.

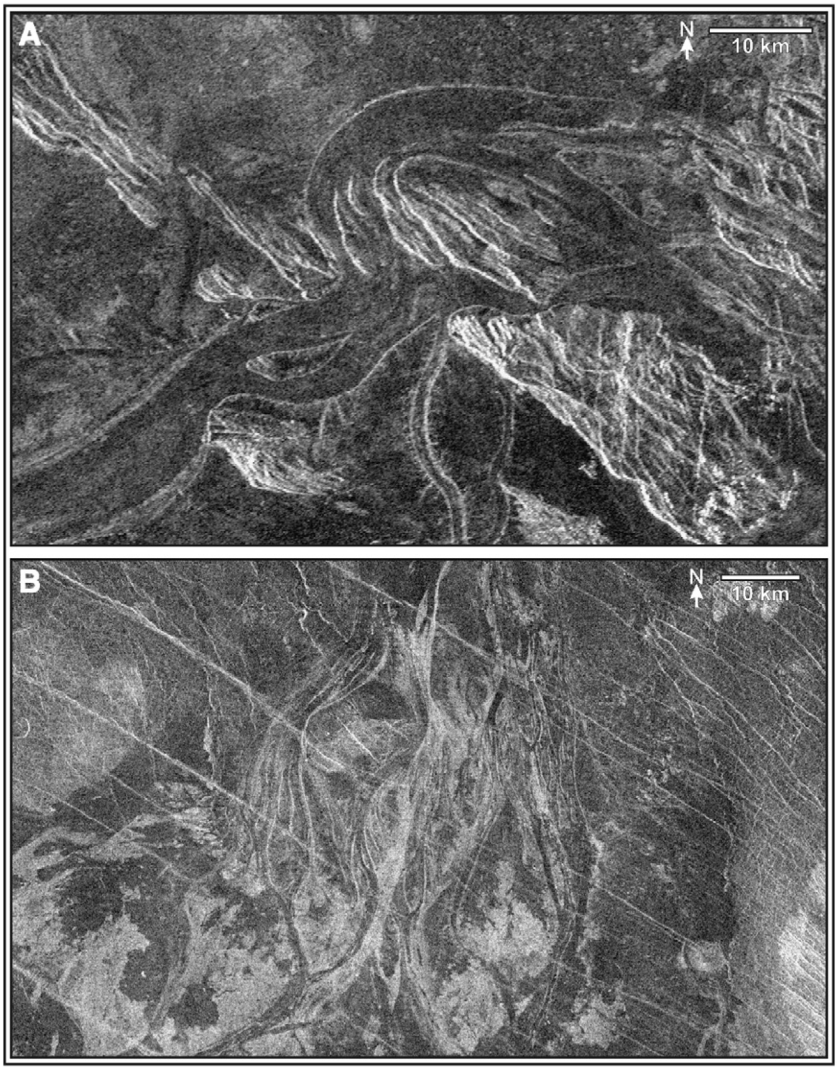

Fig. 5. The largest Venusian volcanic channels have properties comparable to those of the largest Martian systems. (A) The Kallistos Vallis system (Baker et al., 1992) is part of the Ammavaru volcanic complex and is over 1000 km long (after Leverington, 2009); flow along this reach was eastward. The system heads to the NW of this reach, in a zone of chaotic terrain (Fig. 4C). (B) A large anastamosing volcanic channel system (Komatsu et al., 1993; Leverington, 2007) is located NE of Ozza Mons, eastern Aphrodite Terra. Flow along this reach was northward across the central part of the depicted region. Figure centers: (A) 51.0° S., 22.9° E.; (B) 11.6° N., 211.6° E. Magellan FMAP left-look synthetic aperture radar mosaics; radar illumination is from the left.

(e.g., Lingenfelter et al., 1968; Jones and Pickering, 2003; Waltham et al., 2008), volcanic interpretations are widely favored on the basis of the availability of relevant terrestrial analogs (e.g., Greeley, 1971a,b, 1977), the strong association of Venusian and lunar systems with volcanic landscapes (e.g., Wilhelms, 1987; Baker et al., 1992b), the anhydrous nature of associated geological materials (e.g., Goles et al., 1970; Keil et al., 1970; Swann et al., 1972; Papike et al., 1991; Nimmo and McKenzie, 1998), and the long-term instability of water at the surfaces of Venus and the Moon (Wilhelms, 1987; Papike et al., 1991). The properties of Rima Hadley, a lunar channel visited by astronauts during the Apollo 15 mission, are consistent with its development as a conduit for lavas of mafic composition (Greeley, 1971a; Howard et al., 1972; Swann et al., 1972; Carr, 1974). Eruption rates in excess of $\sim 1 \times 10^4$ to 1×10^6 m^3/s are estimated to have been involved in the development of lunar channels and emplacement of mare flows (e.g., Schaber, 1973; Schaber et al., 1976), and corresponding eruption rates of up to $\sim 5 \times 10^7$ m^3/s have been estimated for development of the largest Venusian channels (Baker et al., 1997). Such rates are considerably larger than the tens of cubic meters per second typical of modern volcanic eruptions on Earth (e.g., Hon et al., 1994; Calvari et al., 2002), as well as the maximum effusion rates estimated for large recent terrestrial eruptions such as the Laki fissure eruption of 1783–1784 ($\sim 8.7 \times 10^3$ m^3/s; Thordarson and Self, 1993). Instead, estimated lunar and Venusian rates of flow are more comparable to, for example, the rates of $\sim 1 \times 10^4$ to 1×10^6 m^3/s estimated for emplacement of individual units of the Columbia River Basalt Group (Reidel, 1998) (see discussion in Leverington, 2007).

Many lunar systems are comprised of sinuous channels that head at outflow zones marked by topographic depressions or other features of disturbance and typically extend downslope to terminal basins mantled by ridged volcanic plains (e.g., Gornitz, 1973; Guest and Murray, 1976; Head, 1976; Schultz, 1976; Wilhelms, 1987). The head depressions of lunar channels are characterized by a wide range of possible forms, with the elongate morphologies of some depressions (e.g., at Rimae Prinz and Rima Hadley) suggesting tectonic control of underlying intrusions (e.g., Gornitz, 1973; Strain and El-Baz, 1977; Leverington, 2009). Channels of relatively limited dimensions (with widths of up to ~3 km and lengths of tens to hundreds of kilometers) are most common on the Moon and include systems that head in upland regions (e.g., Rima Hadley, Rimae Plato, and Rimae Maupertuis) and systems that are mostly or entirely confined to mare units (e.g., Rimae Herigonious and Rima Brayley) (e.g., Oberbeck et al.,

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1969; Greeley, 1971a,b; Howard et al., 1972; M'Gonigle and Schleicher, 1972; Gornitz, 1973; Young et al., 1973; Schultz, 1976; Greeley and Spudis, 1978; Head and Wilson, 1991, 1992; Leverington and Maxwell, 2004; Leverington, 2006). With channel widths of up to ~10 km, the largest lunar outflow systems are located in the Aristarchus region of the Moon, and include Vallis Schröteri, Rimae Aristarchus, and Rimae Prinz (e.g., Gornitz, 1973; Schultz, 1976; Strain and El-Baz, 1977; Zisk et al., 1977; Wilhelms, 1987; Campbell et al., 2008) (Fig. 4A). Most lunar systems are comprised of simple channels, but some possess attributes that on other solar system bodies might be considered diagnostic of the flow of water, such as inner channels, anastamosing reaches, streamlined erosional residuals, branching channel patterns, and reaches suggestive of lateral or vertical erosion (e.g., Howard et al., 1972; Gornitz, 1973; Carr, 1974; Schultz, 1976; Strain and El-Baz, 1977; Greeley and Spudis, 1978; Wilhelms, 1987; Leverington, 2004, 2006; Bleacher et al., 2010). The morphologies of lunar channels and numerous mare flows suggest that associated lavas were generally erupted with low yield strengths and at high flow rates (Hulme, 1973; Schaber, 1973; Hulme, 1974; Guest and Murray, 1976; Head, 1976; Schaber et al., 1976; Hulme and Fielder, 1977; Gifford and El-Baz, 1981; Hulme, 1982; Head and Wilson, 1991, 1992; Zimbelman, 1998), and such inferences are consistent with the high iron and titanium contents and low silica content typical of lunar magmas (Murase and McBirney, 1970, 1973). Importantly, the high rates of effusion implied by volcanic features on the Moon are in accord with our understanding of igneous plumbing systems and can be accommodated by the existence of deep magma reservoirs (Head and Wilson, 1991, 1992) and by volcanic vents only slightly wider than those typical of terrestrial vents (Wilson and Head, 1981). Unlike aqueous outflow mechanisms hypothesized to have operated on Mars, volcanic mechanisms need not require fluid movement through porous media and, for many types of igneous intrusions, are not subject to associated restrictions on rates of subsurface flow.

As with their lunar counterparts, Venusian outflow systems appear to have formed as a result of effusion of large volumes of low viscosity magma to the surface (e.g., Baker et al., 1992b; Head et al., 1992; Komatsu et al., 1993; Komatsu and Baker, 1994; Baker et al., 1997). Many Venusian channels are comparable in size and morphology to simple lunar systems (e.g., Komatsu and Baker, 1994), but some channels are notable for their greater dimensions and more complex reaches (Fig. 5). Some of the largest systems on Venus head at chaotic zones of collapse and disturbance; and several systems have reaches with complex anastamosing character, large streamlined erosional residuals, and widths of several tens of kilometers (Baker et al., 1992b; Komatsu et al., 1993). Some Venusian systems are associated with radar-bright levees and flow deposits that possess properties consistent with volcanic origins (e.g., Baker et al., 1997). Though some Venusian channels may have developed through constructive emplacement of lava flows, many show evidence for having incised into underlying units (e.g., Baker et al., 1992b; Komatsu and Baker, 1994). Exotic magma compositions have previously been hypothesized as necessary for channel development on Venus (Baker et al., 1992b; Komatsu et al., 1992, 1993; Kargel et al., 1994; Baker et al., 1997; Treiman, 2009), but the lunar record demonstrates that lavas of mafic composition can have the capacity for voluminous low viscosity flow, substantive vertical and lateral erosion, and extreme lengths for both open and channelized flows (Murase and McBirney, 1970; Hulme, 1973, 1974; Schaber et al., 1976; Wilhelms, 1987; Leverington and Maxwell, 2004; Leverington, 2006, 2009). Venusian channels have not yet been directly visited by spacecraft, but surface compositions measured at the Venera and Vega landing sites are consistent with those expected of dry mafic lavas (e.g., Surkov, 1983; Surkov et al., 1987; Kargel et al., 1993; Fegley et al., 1997; Nikolaeva and Ariskin, 1999). Eruption of low viscosity, lunar-type basalts on Venus would not be unexpected or inconsistent with available geochemical data (Gregg and Greeley, 1993; Gregg, 1996).

4.2. The volcanotectonic nature of the Martian outflow channels

The lunar and Venusian outflow systems appear to be excellent candidate analogs to the Martian outflow channels (Leverington, 2004, 2007, 2009). Direct correspondence exists between the main components of Martian outflow systems (heads, channels, and terminal basins) and those of lunar and Venusian systems, and an abundance of Martian channel landforms records the past eruption and flow of magma. For example, many Martian outflow channels clearly head at source features that mark the sites of voluminous effusion of lava from the subsurface, including chaotic or otherwise disturbed zones comparable to the heads of lunar and Venusian systems (e.g., Ravi Vallis and Maja Valles), ridged plains comparable to lunar plains emplaced by mare-style volcanism (e.g., Ares Vallis and Ma'adim Vallis), and structural features such as graben-like landforms (e.g., Mangala Valles and Athabasca Valles). Some of the clearest evidence for eruption of volcanic flows at the heads of Martian outflow channels takes the form of ridged or platy flows such as those erupted at Mangala Valles (Leverington, 2007), Athabasca Valles (Jaeger et al., 2007), and the largest outflow channel, Kasei Valles (Chapman et al., 2010) (Fig. 6). The reflectance spectra of well-exposed parts of these flows are consistent with mafic lava compositions (Mangold et al., 2009; Jaeger et al., 2010).

Martian outflow systems possess numerous features suggestive of incision and scour, including inner channels, anastamosing reaches, and streamlined erosional residuals (e.g., Baker and Milton, 1974; Carr et al., 1976; Baker, 1978b; Ghatan et al., 2005). Although identification of associated fluvial deposits of clear sedimentary character has proven difficult (e.g., Greeley et al., 1977; Baker and Kochel, 1979; Mars Channel Working Group, 1983; Tanaka, 1997; Wilson and Mouginis-Mark, 2003; Ghatan et al., 2005; Burr and Parker, 2006; Leverington, 2007; Carling et al., 2009), parts of the floors of all Martian outflow channels are mantled by units possessing attributes consistent with those expected of mare-style ridged volcanic plains, lobate-margined lava flows, platy-ridged lava flows, polygonal-textured lava pools, roofed volcanic rilles, and pitted and cuspate-margined lava flows emplaced through inflation (e.g., Plescia, 1990; Keszthelyi et al., 2000, 2004; Ghatan et al., 2005; Keszthelyi et al., 2006; Jaeger et al., 2007; Leverington, 2007; Keszthelyi et al., 2008; Leverington, 2009; Vaucher et al., 2009; Chapman et al., 2010; Jaeger et al., 2010) (Fig. 7). Overflow deposits with properties consistent with those expected of volcanic origins are present along the margins of numerous Martian outflow systems, including Allegheny Vallis, Athabasca Valles, and Ares Valles (e.g., Fig. 7C and D; Jaeger et al., 2007, 2010; Leverington, 2009). Importantly, new high resolution analyses of the lava flows that completely mantle the Athabasca Valles outflow system are strikingly consistent with channel development through incision by turbulent lava flows (Jaeger et al., 2010), and such inferences are consistent with independent predictions of generally low viscosities for Martian basaltic lavas (McGetchin and Smith, 1978; Greeley and Spudis, 1981; Williams et al., 2005). The longitudinal grooves and ridges found along the floors of many Martian outflow channels are considered by some workers to be excellent evidence for aqueous scour involving roller vortices (Baker, 1978a; Baker et al., 1992a), but might these features have instead been formed by the voluminous lava flows (e.g., ~5×10^3 to 2×10^5 km^3 for relatively small channel systems) predicted by some workers to have been erupted at channel heads (Leverington, 2007, 2009; Jaeger et al., 2010)? Indeed, though many examples of Martian longitudinal features have been previously interpreted to be erosional in nature, meter-scale images indicate that some such features were not formed by

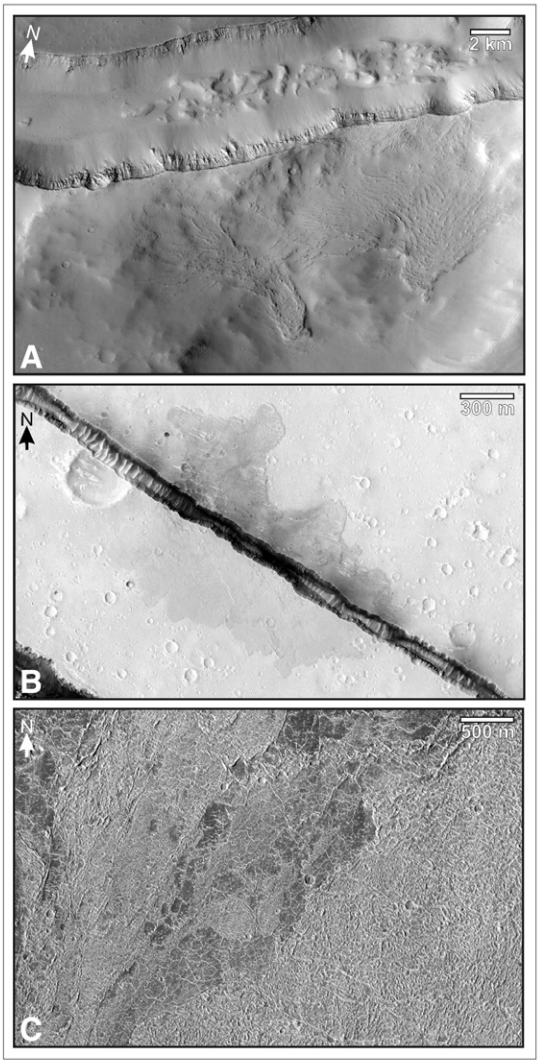

Fig. 6. Many Martian outflow channels head at source features that clearly mark the sites of voluminous effusion of lava from the subsurface. (A) Ridged flows that extend downslope from the head graben of Mangala Valles; these flows are found on both the north and south sides of the graben and are counterparts to those erupted from the same structure at nearby Daedalia Planum (Leverington, 2007). (B) Lobate-margined flows erupted from one of the numerous extensional features that mark the head region of Athabasca Valles, a system that appears to be completely mantled by volcanic flows (Jaeger et al., 2007, 2010). (C) Platy-ridged flows erupted at the head of the largest outflow channel, Kasei Valles (Chapman et al., 2010). Other features commonly found at the heads of Martian outflow channels include mare-style ridged plains and disturbed terrains comparable to the heads of numerous lunar and Venusian systems. Image numbers: (A) Mars Reconnaissance Orbiter Context Camera (CTX) image P08_003991_1636_XN_16S148W; (B) HiRISE image PSP-7197-1910; (C) HiRISE image PSP-10225-1795. Figure centers: (A) 18.15° S., 148.7° W.; (B) 10.7° N., 156.4° E.; (C) 0.4° S., 80.4° W.

erosion at all. Instead they are comprised of the edges of disrupted lava layers and rows of small cones formed during emplacement of associated units, suggesting that the flow dynamics of Martian lavas had the capacity to produce massive parallel systems of ridges and gullies (e.g., Jaeger et al., 2007; Leverington, 2009; Jaeger et al., 2010).

The terminal basins of Martian outflow channels lack clear examples of fluvial deltas and sedimentary shoreline features (e.g., Greeley et al., 1977; Schonfeld, 1977a; Baker et al., 1992a; De Hon, 1992; Carr, 1996; Malin and Edgett, 1999; Burr et al., 2002b; Ghatan and Zimbelman, 2006). Instead, the terminal reaches of channels typically fade into wrinkle-ridged plains with properties consistent with emplacement through mare-style volcanism (e.g., Carr, 1973, 1974; Carr et al., 1976; Greeley et al., 1977; Schonfeld, 1977a, 1979a; Greeley and Spudis, 1981; Baker et al., 1992a; De Hon, 1992; Head and Wilson, 1992; Carr, 1996; Ghatan et al., 2005) (Fig. 8), a property typical of lunar and Venusian outflow systems (e.g., Schultz, 1976; Wilhelms, 1987; Baker et al., 1992b; Leverington, 2004). Though aqueous interpretations are currently widely favored (e.g., Golombeck et al., 1999a,b), the properties of surface materials at the landing sites of the Viking 1, Pathfinder, and Spirit spacecraft are arguably entirely consistent with those expected of volcanic flows subjected to extensive reworking by impact processes, and this interpretation is congruent with inferred emplacement of associated units of Chryse Planitia and Gusev crater by mare-style volcanism (e.g., Carr et al., 1976; Greeley et al., 1977, 2005). Thin sedimentary units that mantle the terminal basins of the northern plains have been hypothesized as lacustrine or oceanic deposits (e.g., Fuller and Head, 2002; Kreslavsky and Head, 2002; Baker, 2006), but the widespread association of boulders with these units (McEwen et al., 2007) and the absence of sedimentary characteristics clearly indicative of deposition at and beyond the mouths of diluvial channel systems are not supportive of these interpretations.

The hypothesis for development of the Martian outflow channels by volcanic flows is testable. For example, well preserved and well exposed outflow systems should have attributes that demonstrate the past occurrence of effusive eruptions at channel heads, the conveyance of volcanic flows along component channels, and the accumulation of voluminous flows at terminal basins (Leverington, 2004). Furthermore, the ages of volcanic units inferred to have been involved in particular episodes of channel development should not be separated greatly in time (Schonfeld, 1977a; Cutts et al., 1978). Notably, however, the presence of volcanic landforms and deposits along Martian outflow channels and at associated terminal basins does not preclude channel development through nonvolcanic mechanisms. Thus, for example, the mantling of Martian outflow systems by volcanic materials does not in itself disprove hypotheses of channel incision involving aqueous mechanisms. Instead, the widespread presence of volcanic features can only be said to be broadly consistent with a requirement of the volcanic hypothesis. However, in the absence of clear evidence for the past occurrence of aqueous flood events, the volcanic hypothesis stands alone as the only proposition that involves liquids known to have flowed in large volumes along Martian outflow channels.

4.3. Could Martian igneous intrusions have supplied channel-forming eruptions?

Past development of large igneous intrusions within rocky bodies of the inner solar system is expected to have been driven by energy derived from early processes of accretion, gradual internal differentiation and cooling, and radioactive decay (e.g., Siegfried and Solomon, 1974; Horedt, 1980; Turcotte, 1980). Large mantle-sourced bodies can buoyantly rise as a result of the density differences that exist between these bodies and surrounding materials. Rising bodies that are not confined by rheological or other mantle barriers will typically become stranded in neutral buoyancy zones (e.g., at the base of a planetary crust), where overlying materials are of lower bulk density (e.g., Head and Wilson, 1992; Ryan, 1994; Grosfils and Head, 1995; Wilson and Head, 2002). Magmas stranded in neutral buoyancy zones can rise beyond these zones as a result of the lithostatic pressure P exerted by overlying materials:

$$P = \rho g h \tag{1}$$

where ρ is the average density of the overlying materials, g is acceleration due to gravity (3.72 m/s^2 for Mars), and h is depth below the surface (e.g., Turcotte and Schubert, 2002). The igneous intrusions

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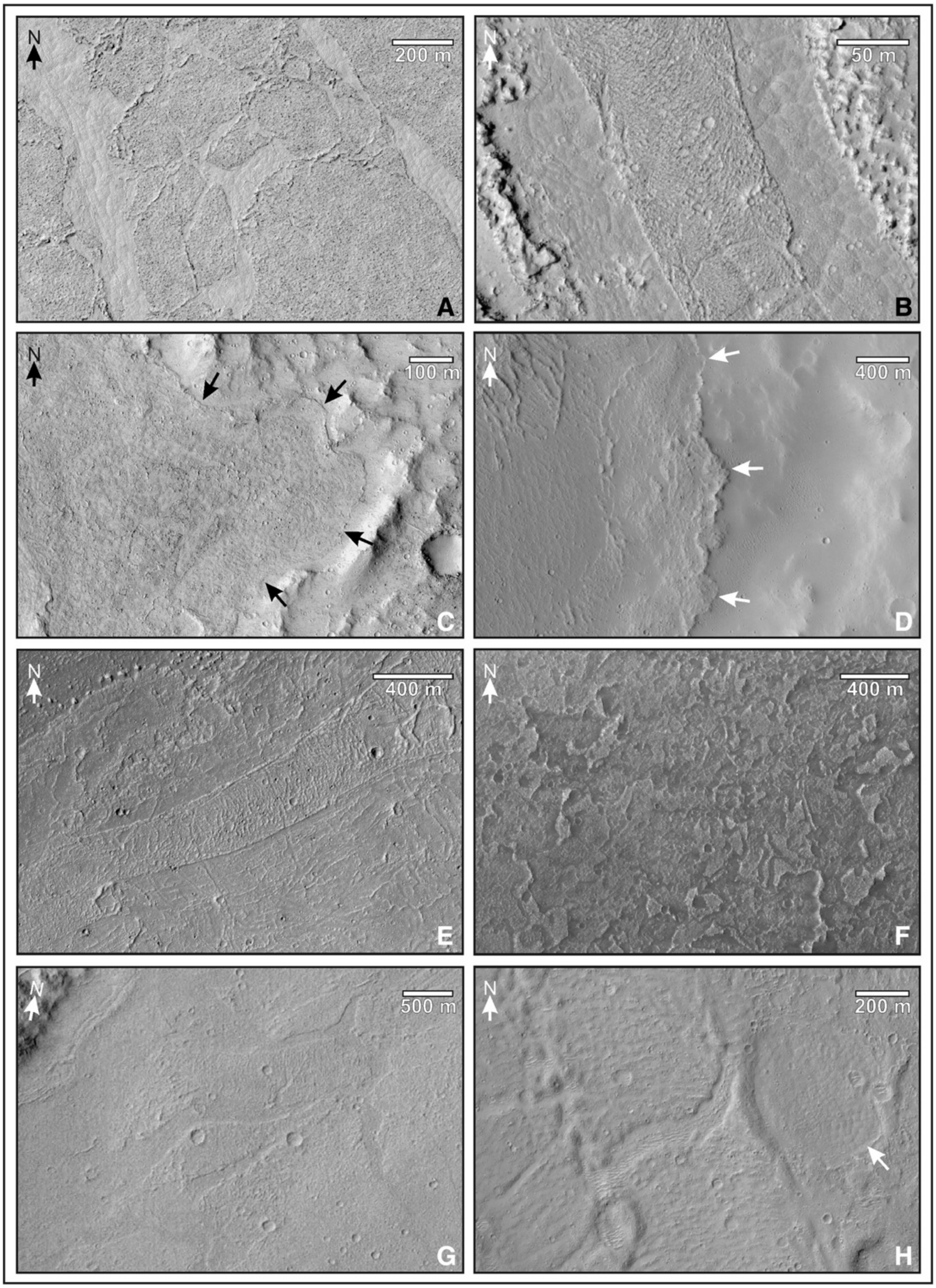

Fig. 7. Though a notable diversity of volcanic landforms and deposits is visible at young and well-exposed systems such as Athabasca Valles (e.g., Keszthelyi et al., 2000, 2004, 2006; Jaeger et al., 2007, 2010), features with clear volcanic affinities are present along most outflow systems. (A) Platy-ridged flows at which plates are separated by materials with polygonal lava-lake textures, Athabasca Valles. (B) Ridged flows separated by shear zones from adjacent units with polygonal lava-lake textures, Athabasca Valles. (C) Overflow unit with lobate margins (flow front marked by arrows), Athabasca Valles. (D) Overflow unit with lobate margins and ridged character (flow front marked by arrows), Kasei Valles. (E) Ridged flow separated by shear zones from adjacent flow materials, Kasei Valles. (F) Inflated lava flows, Kasei Valles. (G) Platy-ridged flows, Labou Vallis branch of Mangala Valles. (H) Thinly mantled flows with platy-ridged character and polygonal texture (arrow), Mangala Valles. Additional examples of features such as these include the roofed subchannels and inflated flows of Mangala Valles (Leverington, 2007); the lobate-margined overflows and slabby flows of Allegheny Vallis and Elaver Vallis (Leverington, 2009); and the wrinkle-ridged units that are discontinuously present along the floors of almost all outflow systems. Image numbers: (A) HiRISE TRA_000854_1855; (B) HiRISE PSP_001540_1890; (C) HiRISE PSP_006472_1855; (D) HiRISE PSP_010607_2040; (E) HiRISE PSP_007825_2010; (F) HiRISE PSP_010000_2055; (G) CTX P13_006180_1717_XN_08S153W; (H) HiRISE PSP_007617_1750.

that develop above a magma chamber as a result of lithostatic pressure will generally be oriented approximately perpendicular to the least compressive stress within the host rock (e.g., Rubin, 1995), ultimately forming intrusions such as dikes. The level to which magmas can be lithostatically driven is in large part constrained by the depth of the parent magma chamber. Equating the lithostatic

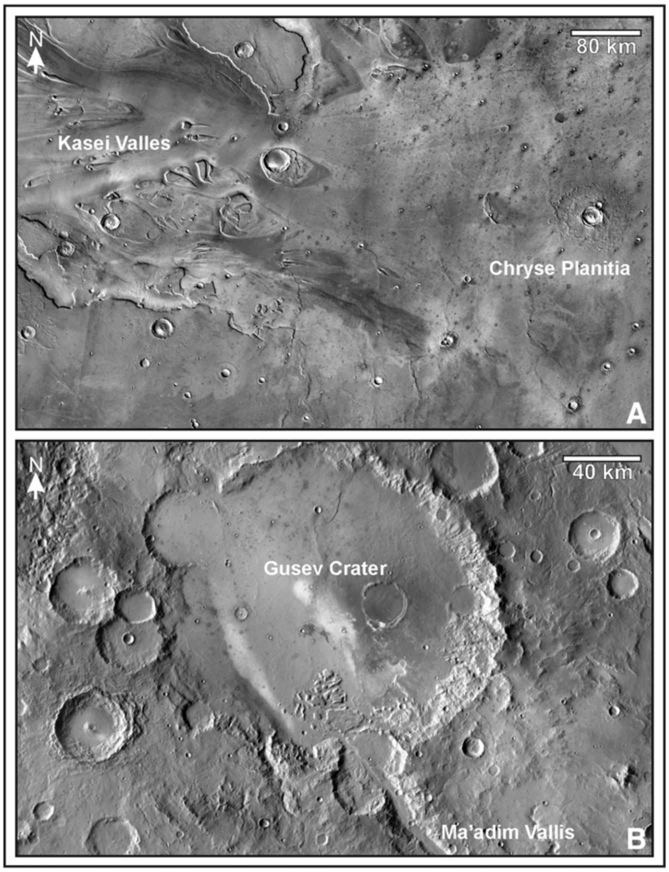

Fig. 8. Along their lower reaches, the Martian outflow channels fade into plains with characteristics consistent with volcanic origins. (A) Kasei Valles and several other large outflow systems of the circum-Chryse region all terminate at the ridged volcanic plains of Chryse Planitia (e.g., Carr et al., 1976; Greeley et al., 1977; Arvidson et al., 1989). (B) The smaller Ma'adim Vallis system similarly terminates at the ridged volcanic plains of Gusev Crater (e.g., Greeley et al., 2005). This relation between outflow systems and volcanic plains on Mars is similar to that recognized on the Moon (e.g., Fig. 4A). Thermal Emission Imaging System (THEMIS) daytime infrared mosaics courtesy of Arizona State University. Figure centers: (A) 25.5° N., 49° W.; (B) 14.6° S., 175.5° E.

pressure on the magma chamber with the hydrostatic pressure at the base of a growing intrusion, the maximum height above a stranded magma chamber that fluids can reach as a result of lithostatic pressure is given by

$$H = (D_h \rho_h)/\rho_1 \tag{2}$$

where H is the height above the stalled position, D_h is the thickness of the material above the magma chamber, ρ_h is the average density of the material above the magma chamber, and ρ_l is the density of the magma (Head and Wilson, 1992). Though lithostatic pressure is a major control on the level to which magma can flow, magma will only rise beyond a stalled chamber if a path already exists and can be opened by magma pressure or if fluid pressures are sufficient to open a new path by overcoming the tensional strength of overlying rock (e.g., Rubin and Pollard, 1987; Rubin, 1995). In the absence of excess pressures such as those related to dynamic inflow into a magma chamber (e.g., Chen et al., 2007) or to favorable gradients of tectonic stress (e.g., Rubin, 1995), magma intrusions will generally reach no shallower than the level given by Eq. (2).

Assuming existing paths to the surface can be exploited or new paths can be developed, a deeper chamber will tend to have a greater capacity to send magma to the surface than a shallower chamber. The deepest magma chambers should include those that become stranded beneath thick crusts and those that become confined by deep traps associated with rheological boundaries in mantle materials (e.g., Head and Wilson, 1992). For magma chambers that become stalled among mantle materials, the maximum height to which magma can rise in a connected path is modified from Eq. (2):

$$H\rho_1 = D_c\rho_c + D_m\rho_m \tag{3}$$

where H is the height above the stalled position, ρ_l is the density of the magma, D_c is the thickness of the crust above the magma chamber, D_m is the thickness of mantle materials above the magma chamber, ρ_c is the average density of the overlying crust, and ρ_m is the average density of overlying mantle materials (Head and Wilson, 1992).

Can Martian igneous intrusions have supplied magma to the surface at volumes and rates sufficient for channel development? Assuming a typical highland crustal thickness of 75 km, a large magma chamber stranded at the base of the crust will be subjected to a lithostatic pressure of ~810 MPa. If the average density of crustal materials is 2900 kg/m^3, and if the density of basaltic magmas is assumed to be 3000 kg/m^3, Eq. (2) implies that a dike could rise no closer to the Martian surface than a depth of ~2500 m (Fig. 9), allowing eruption only along the floors of isolated and deep topographic basins. However, assuming a mantle density of 3500 kg/m^3, a magma chamber stranded at a rheological transition marking the base of a 120- to 200-km-thick lithosphere would be subjected to a lithostatic pressure of ~1.4 to 2.4 GPa. This pressure would be theoretically sufficient to raise a magma column ~5 to 18 km above the Martian surface, if the imaginary chamber were of a volume sufficient to allow formation of an intrusion to the surface and to allow additional magma to be expelled to the Martian surface. For realistic igneous systems, the available magma volume would be a key limiting factor on the volume ultimately expelled to the surface. In the consideration of the capacity of Martian intrusions for driving large channel-forming eruptions, it can be useful to examine past estimates of the volumes of large Martian intrusions. For example, the large individual dikes that are believed to underlie the Tharsis-radial graben system are predicted to have mean widths up to ~400 m and typical lengths of up to ~3000 km (Wilson and Head, 2002). Assuming the height of a giant Martian dike to be 120 km (Wilson and Head, 2002),

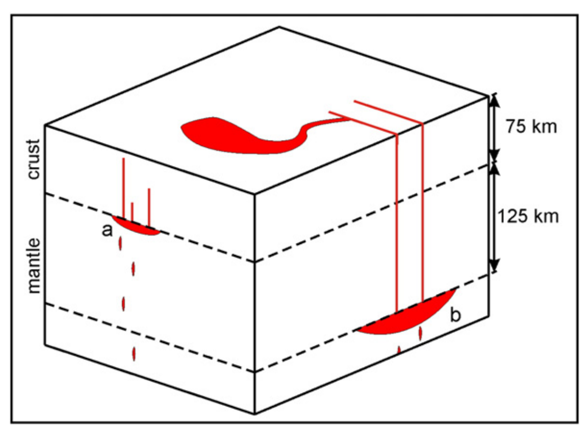

Fig. 9. The lithostatic pressure on a magma chamber stranded beneath a 75-km-thick Martian crust (a) will generally be of insufficient magnitude (810 MPa) to allow intrusions to reach the surface, except at local topographic lows such as impact craters and tectonic features (see discussion in Section 4.3). In comparison, a magma chamber stranded at a rheological boundary such as that at the base of a 120- to 200-km-thick lithosphere will be characterized by sufficient pressure (~1.4 to 2.4 GPa) to drive rapid and voluminous eruptions to the surface; the 200 km scenario is depicted at *b*. The volumes of large individual dikes on Mars should reach or exceed ~50,000 km^3 (e.g., Wilson and Head, 2002), suggesting that relatively rapid eruption of comparable volumes to the Martian surface might have been possible if intrusions such as these were fed by and connected to large reservoirs stranded at similarly shallow depths of the mantle.

a single intrusion with a width of 200 m and length of 2000 km would have a volume of ~48,000 km^3. If only an additional ~17% of this volume were to be erupted to the surface, it would correspond to the ~8000 km^3 magma volume estimated by Jaeger et al. (2010) to have been involved in mechanical erosion of the Athabasca Valles outflow system. Still larger erupted volumes, such as the ~2×10^5 km^3 estimated to have been involved in volcanic development of the Mangala Valles system (Leverington, 2007), might similarly be realistic consequences of eruption from one or more large intrusions extending from reservoirs stranded within the upper mantle. Though the above calculations are crude estimates that ignore all but the most basic considerations, they suggest that igneous processes should indeed be viable mechanisms by which the outflow channels of Mars could have developed. The very existence of large volcanic outflow systems on the Moon and Venus (Figs. 4 and 5) is consistent with this perspective.

5. Discussion and implications

Hypotheses for aqueous development of the outflow channels of Mars do not appear to be viable, on the basis of associated requirements for implausible properties of the Martian near-surface, for water abundances that are orders of magnitude greater than can be independently justified, for long-term near-surface conditions that may not be congruous with the preserved mineralogical record, for prominent aqueous deposits that have not yet been confidently identified along channels and at terminal basins, and for the action of processes for which there are no known solar system analogs. Instead, the volcanic hypothesis for development of the outflow channels of Mars appears to represent the simplest and most effective means by which these systems can be understood. This hypothesis is founded upon the existence of volcanic analog landforms on multiple bodies of the inner solar system and on corresponding analog processes that are high volume and low viscosity variations on familiar terrestrial volcanic themes. The volcanic hypothesis is compatible with available mineralogical data for Mars, is consistent with the nature of landforms and deposits found along the lengths of Martian channels, and brings the outflow channels into conformity with the low near-surface water abundances predicted by geochemical considerations. The volcanic hypothesis is also consistent with recent volcanic interpretations of smaller Martian channels that possess characteristics similar to those of larger systems, such as streamlined erosional residuals and complex sinuous or anastamosing reaches (Leverington and Maxwell, 2004; Williams et al., 2005; Leverington, 2006; Caprarelli et al., 2007; Garry et al., 2007; Keszthelyi et al., 2008; Rampey and Harvey, 2008; Hauber et al., 2009; Bleacher et al., 2010).

More fundamentally, the volcanic hypothesis is attractive as it fits within a wider geological framework that realistically and economically accounts for the existence and nature of outflow systems located on several bodies of the inner solar system. The outflow systems of Mars are presently treated as features distinct from volcanic channels of the Earth, Moon, and Venus, but the volcanic hypothesis suggests that these systems are products of the same processes and are, at their most basic level, variations within the same family of landform. A volcanic origin for the outflow channels of Mars would further highlight the capacity of ancient lava flows for voluminous effusion and significant erosion by thermal or mechanical processes and would imply an important new unification in major effusive volcanic processes and landforms across at least three bodies of the inner solar system. A volcanic origin for the Martian outflow channels would imply that outflow zones are the surface expressions of magmatic systems rather than the locations of past aqueous outbursts from aquifers. As such, the outflow channels would join the large Martian shields as among the most prominent volcanic landforms in the solar system (Fig. 10). The northern lowlands of Mars are characterized by vast volcanic plains of Hesperian age (Head et al., 2002) that mantle densely cratered units of Noachian age (Frey et al., 2002), and a volcanic origin for the Martian outflow channels would imply that these plains are, as with the lunar maria, partly composed of volcanic units sourced from outflow channels. Assuming a missing volume of ~4×10^6 km^3 at the circum-Chryse systems (Carr, 1987; Baker et al., 1992a; Carr, 1996), greatly simplified thermal considerations that neglect the possible action of mechanical processes (Leverington, 2007) imply a corresponding erupted magma volume of 1×10^8 km^3. If distributed across the Borealis basin, this implies an average thickness of ~2.4 km, a value that exceeds the recognized ~0.9 km average thickness (Head et al., 2002) and that leaves no room for contributions by other volcanic sources of the northern lowlands. Assuming volcanic origins for the Martian outflow channels, this thermally derived overestimation suggests that recent predictions of the predominance of mechanical processes in lava incision on Mars (Hurwitz et al., 2010; Jaeger et al., 2010) are likely to be correct.

If the outflow systems of Mars, Venus, and the Moon developed through common volcanic processes, might large analogous systems have developed on the Earth during the Hadean or Archean eras? Even today, terrestrial outflow processes can form the modest channel systems commonly involved in constructive emplacement of volcanic flows on the Earth (e.g., Greeley, 1971a,b; Greeley and Hyde, 1972;

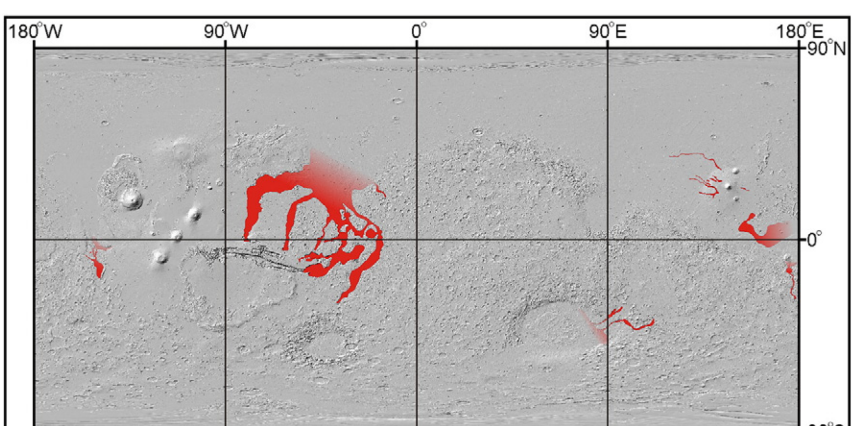

Fig. 10. A volcanic origin for the largest outflow channels of Mars (red) would render these features among the most prominent igneous landforms in the solar system. A volcanic origin would also imply that, as at the lunar maria, the ridged volcanic plains at terminal basins are partly composed of flows derived from these outflow systems. Global map of shaded topography after Smith et al. (2003). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

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Peterson and Swanson, 1974; Greeley, 1977). Unfortunately, the poor preservation of early terrestrial crust (Kamber et al., 2005) — combined with variations in the inner solar system with regard to volatile abundance, crustal thickness, degree of differentiation, thermal history, and activity of plate tectonics (Anderson, 1984; O'Neill et al., 2007) — presently limits inferences of the past existence of large terrestrial outflow systems to the realm of speculation. Although early mantle temperatures were high (Ryder et al., 2000) and mantle processes likely had the capacity to drive periodic resurfacing events (Davies, 2008), formation of outflow systems on Earth could have been inhibited by the existence of voluminous near-surface water reservoirs (e.g., Ryder et al., 2000), the operation of early plate tectonic processes (e.g., Hopkins et al., 2008), or limiting aspects of early mantle chemistry (e.g., Carr and Wänke, 1992).

If the volcanic hypothesis for outflow channel development on Mars is correct, potential implications extend beyond those directly related to the origins of solar system channels. The volcanic hypothesis does not rely upon ancient swings in Martian climatic conditions or atmospheric properties and is therefore consistent with a wide range of possible climatic histories, including the long-term cold and dry conditions implied by global surface mineralogy (e.g., Hoefen et al., 2003; Bibring et al., 2006; Forget, 2007; Koeppen and Hamilton, 2008). Since inferences of the existence of large northern lakes and oceans during the Hesperian and Amazonian are based mainly on aqueous interpretations of outflow systems (e.g., Parker et al., 1989; Scott et al., 1992; De Hon and Pani, 1993; Wharton et al., 1995; Komatsu and Ori, 2000; Williams et al., 2000; Baker, 2001; Head et al., 2001; Jakosky and Phillips, 2001; Masson et al., 2001; Parker and Currey, 2001; Carr and Head, 2003; Fairén et al., 2003; Baker, 2009c), a volcanic origin for the outflow channels would substantially change our understanding of the nature and extent of past aqueous environments on Mars.

Development of life on Mars would have likely required environments within which liquid water was stable over some amount of time (e.g., McKay and Stoker, 1989; McKay et al., 1992; Chyba and McDonald, 1995; Mancinelli and Banin, 1995; Wharton et al., 1995; Brack and Pillinger, 1998; Hiscox, 2001). Among the ancient Martian environments identified as potentially supportive of past biological activity are the outflow channels themselves (e.g., Burr et al., 2002a; Dohm et al., 2004; Levy and Head, 2005; Murray et al., 2005; Warner et al., 2010), oceans at the mouths of outflow systems (e.g., Helfer, 1990; Wharton et al., 1995; Komatsu and Ori, 2000; Parker and Currey, 2001), water bodies and zones of hydrothermal activity at channel heads (e.g., Rodriguez et al., 2005; Coleman et al., 2007; Schulze-Makuch et al., 2007), and a range of other associated surface or subsurface environments (e.g., McKay, 1997; Dohm et al., 2004; Schulze-Makuch et al., 2005). A volcanic origin for the Martian outflow channels would dramatically narrow the possible range of environments once supportive of Hesperian or Amazonian life and could reduce the broader likelihood of past life on Mars by diminishing the geomorphological justification for the existence of immense but hidden near-surface water reservoirs. Because many of the regions of greatest interest on Mars correspond to those most likely to have experienced the prolonged activity of liquid water (e.g., Farmer et al., 1995; Hubbard et al., 2002; Jakosky and Mellon, 2004), recognition of a volcanic origin for the Martian outflow channels could decrease the appeal of outflow sites previously considered to be prime targets for future robotic or human exploration. Despite narrowing the possible range of Martian environments once hospitable to life, recognition of a volcanic origin for the outflow channels of Mars could improve our capacity for discrimination of sites of genuine astrobiological significance by highlighting the greater importance of other Martian sites characterized by clearer records of past aqueous conditions.

Volcanic interpretations of the Martian outflow channels raise important new questions regarding the fundamental nature of all channel systems on Mars. If Martian volcanic processes had the capacity in the Hesperian and Amazonian epochs for generation of outflow systems with characteristics that superficially resemble those of aqueous systems, might other Martian channel systems share common volcanic origins? The properties of highland channel networks of Hesperian age are considered by some workers to be consistent with formation through regional volcanic resurfacing (e.g., Leverington and Maxwell, 2004; Leverington, 2006). Might large low-order sinuous channel networks such as the Nanedi and Nirgal systems, by analogy with remarkably similar lunar and Venusian features, have been formed by low viscosity lava flows rather than conventionally cited aqueous sapping processes or surface overflow? Because volcanic channel origins do not require special atmospheric properties for development, recognition of the widespread predominance of volcanic processes in Hesperian and Amazonian channel formation could further influence our understanding of ancient climatic conditions and volatile stores.

A volcanic origin for the largest channels in the solar system raises questions regarding the basic nature of involved igneous processes. For example, what flow volumes and rates of effusion were involved in the formation of specific Martian systems, and what ranges of physical properties (temperatures of eruption, lava chemistries, lava yield strengths, and Reynolds numbers) were once associated with these flows (e.g., Schonfeld, 1976, 1977a,b; McGetchin and Smith, 1978; Schonfeld, 1979a,b; Wilson and Mouginis-Mark, 2001; Williams et al., 2005; Leverington, 2007, 2009; Jaeger et al., 2010)? How would the properties of Martian flows have compared with those of flows associated with lunar and Venusian outflow systems? To what relative extents were thermal erosion (involving the melting of substrates by thermal energy) and mechanical erosion (involving the physical removal of substrates by kinetic energy) involved in channel incision (Cutts et al., 1978; Leverington, 2007; Jaeger et al., 2010)? Though erosive processes seem likely to have dominated in the formation of major channel systems, what roles might constructive volcanic processes have also played in system development? Still more fundamental questions arise: what kinds of intrusive igneous bodies would have been necessary to supply inferred flow volumes at required rates of effusion (Jaeger et al., 2010)? On the Moon, eruption of large volumes of low viscosity lava should have been possible from deep intrusions associated with magma conduits only slightly wider than those common on the Earth (Wilson and Head, 1981; Head and Wilson, 1991, 1992; Wilson and Head, 2010). Similarly deep and voluminous source magma chambers have been surmised for the outflow channels of Mars (McGetchin and Smith, 1978; Jaeger et al., 2010), but many uncertainties remain.

6. Conclusions

Though the Martian outflow channels are widely interpreted as the products of catastrophic floods from aquifers, aqueous hypotheses suffer from numerous weaknesses. Among the most problematic shortcomings of aqueous models are the following: (i) processes hypothesized to have triggered sudden development of aqueous Martian floods do not appear to be consistent with realistic expectations for megaregolith permeability; (ii) hypothesized sediment concentrations are several orders of magnitude greater than are likely and are inconsistent with the absence of prominent examples of fluvial depositional landforms along outflow channels and at channel mouths; (iii) the location of channel heads far above terminal plains is not consistent with expected variations in the hydraulic head of large well-connected and highly permeable aquifers; (iv) the near-surface water abundances required of aqueous interpretations are orders of magnitude higher than most corresponding estimates derived from geochemical considerations; (v) widespread preservation of minerals that are highly susceptible to aqueous alteration does not appear to be consistent with the required long-term existence of water-saturated

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cryospheric seals and voluminous underlying aquifers; (vi) the absence of clear examples of fluvial deposits along the outflow channels and at terminal basins suggests that correspondence between terrestrial diluvial landscapes and those of the Martian outflow channels may be limited to erosional features; and (vii) no known solar system analogs exist for catastrophic development of large channel systems by outbursts from aquifers. The modest explanatory power of aqueous hypotheses is generally undermined by the limited congruence between expected and observed channel properties and by the numerous requisite assumptions that serve only to increase the complexity of these hypotheses.

The volcanic hypothesis for development of the outflow channels of Mars represents the simplest and most effective means by which these systems can be understood. The basic origin of the outflow channels, involving effusion of fluids from the subsurface, is far more consistent with volcanic mechanisms of development than with aqueous processes such as those that formed the floodways of terrestrial glacial lakes. Excellent correspondence exists between the general characteristics of the Martian outflow channels and those of candidate volcanic analogs of the Moon and Venus, and the Martian outflow channels similarly show abundant evidence for having conveyed large volumes of magma from channel heads to terminal basins. A capacity for incision of channels by magma is in accord with lunar and Venusian analogs, and with the low lava viscosities inferred for Mars based on geochemical considerations and on the character of Martian volcanic deposits. Volcanic development of the Martian outflow channels is compatible with available geochemical and mineralogical data for Mars and does not require the occurrence of special climatic conditions. The volcanic hypothesis is especially attractive as it fits within a wider geological framework that economically accounts for the existence and nature of outflow systems located on several bodies of the inner solar system. A volcanic origin for the outflow channels of Mars reduces the probability that extensive aqueous environments existed along the outflow channels and at terminal basins during the Hesperian and Amazonian and therefore narrows the possible range of Martian environments once hospitable to life.

Acknowledgments

The author thanks Goro Komatsu and an anonymous reviewer for their helpful comments. The editorial recommendations made by journal staff are also appreciated. The author acknowledges the use of High Resolution Imaging Science Experiment images processed by the University of Arizona, Thermal Emission Imaging System images processed by Arizona State University, and Mars Reconnaissance Orbiter Context Camera images archived by Arizona State University.

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Please cite this article as: Leverington, D.W., A volcanic origin for the outflow channels of Mars: Key evidence and major implications, Geomorphology (2011), doi:10.1016/j.geomorph.2011.05.022

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Please cite this article as: Leverington, D.W., A volcanic origin for the outflow channels of Mars: Key evidence and major implications, Geomorphology (2011), doi:10.1016/j.geomorph.2011.05.022

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Please cite this article as: Leverington, D.W., A volcanic origin for the outflow channels of Mars: Key evidence and major implications, Geomorphology (2011), doi:10.1016/j.geomorph.2011.05.022

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Please cite this article as: Leverington, D.W., A volcanic origin for the outflow channels of Mars: Key evidence and major implications, Geomorphology (2011), doi:10.1016/j.geomorph.2011.05.022