Martian volcanism: current state of knowledge and known unknowns

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ABSTRACT

Much has been discovered about volcanism on Mars over the past fifty years of space exploration. Previous reviews of these discoveries have generally focused on the volcanic constructs (e.g., Olympus Mons and the other volcanoes within the Tharsis and Elysium regions), the analysis of individual lava flows, and how this activity has evolved over time. Here we focus on attributes of volcanology that have received less attention and build upon characteristics of terrestrial volcanoes to pose new questions to guide future analyses of their Martian equivalents either with existing data sets or with new types of measurements that need to be made. The remarkable lack of exposed dikes at eroded ancient volcanoes attests to an internal structure that is different from terrestrial equivalents. Enigmatic aspects of the origin of the ridged plains (commonly accepted to be volcanic but with few identifiable flow fronts and only rare vents), the style(s) of volcanism during the earliest period of Martian history (the Noachian), and the possible mode(s) of formation of the Medusae Fossae Formation are considered here. Martian meteorites have been dated and are volcanic, but they cannot be correlated with specific geographic areas, or the chronology of Mars derived from the number of superimposed impact craters. Some of these questions about Martian volcanism can be addressed with existing instrumentation, but further progress will most likely rely on the acquisition of new data sets such as high-resolution gravity data, the return of samples from known localities, the flight of a synthetic aperture imaging radar, penetrators sent to the Medusae Fossae Formation, and detailed *in situ* field observations of selected volcanic sites

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1. Introduction

Carr (1973) and Greeley and Spudis (1981) conducted the first reviews of volcanism on Mars using Mariner 9 and Viking Orbiter images, respectively, and they concluded that volcanism has been one of the most important geologic processes operating throughout Mars' evolution. This is consistent with the results of Viking Orbiter-based (e.g., Scott and Tanaka, 1986; Greeley and Guest, 1987; Tanaka et al., 1988) and more recent (Tanaka et al., 2014a, b) global geologic mapping. In situ chemical analyses of the Martian soil by the Viking Lander spacecraft (Banin et al., 1992) indicated mafic to ultramafic source rocks, consistent with Earth-based remote sensing spectra (McCord et al., 1978; Singer et al., 1979) and Viking Orbiter color data (McCord et al., 1982). Photogeologic analysis showed two main types of volcanic morphologies: (1) central vent volcanoes, and (2) volcanic plains. Early work (Carr, 1973) identified the giant Tharsis and Elysium volcanoes, as well as the circum-Hellas volcanoes, as examples of Martian central-vent volcanoes. Lunae and Hesperia Plana were seen as examples of a second volcanic morphology, based on the assumption that the mare-type wrinkle ridges within these plains were equivalent to the ridges that formed in lunar mare lava flows (Phillips and Lambeck, 1980). General trends in the temporal evolution of volcanism on the planet were also identified early in the exploration of Mars, with early (Noachian) highland volcanism characterized by (potentially) explosive volcanism around Hellas, and younger (Hesperian and Amazonian) shield-building effusive activity first in Elysium and then more recently within Tharsis (Scott and Carr, 1978).

Mouginis-Mark et al. (1992a), Hodges and Moore (1994) and Greeley et al. (2000) updated these Viking Orbiter analyses of volcanism on Mars, but basic knowledge of the distribution of young volcanic features remained unchanged. The similarity between the giant volcanoes within the Tharsis region and terrestrial basaltic shields such as Mauna Loa was recognized early during the exploration of Mars (Carr and Greeley, 1980). Similarly, Viking

Orbiter (and Mariner 9) images enabled the more heavily eroded low-relief Tyrrhenus and Hadriacus Montes in the southern highlands to be identified along with the second large volcanic province, Elysium (Carr et al., 1977). It was only later that the youngest large volcanic province within Cerberus Fossae was recognized (Berman and Hartmann, 2002; Plescia, 2003). Changes in eruptive style over geologic time, perhaps even at a single volcano (Mouginis-Mark et al., 1988), may also have included geochemical evolution of the parental magmas (Baratoux et al., 2013).

More recently, Zimbelman et al. (2020a) documented the current understanding of the major volcanic constructs on Mars, including results of recent geologic mapping investigations at increasingly larger scales, along with compositional analyses and studies of volcanic geomorphology. However, despite almost three decades of improving image spatial resolution, spectral coverage, and topographic analysis, some fundamental aspects of Martian volcanism identified by Mouginis-Mark et al. (1992a) remain poorly understood. These aspects include: What is the internal structure of a typical Martian volcano? Are the ridged plains volcanic in origin and, if so, what were the eruption conditions associated with their emplacement? What was the nature, distribution, and magnitude of volcanism in the Noachian Period? What is the role of explosive volcanism on Mars and were explosive eruptions driven by magmatic gas exsolution and/or interactions with near-surface volatiles? Where on Mars did the Shergotty, Nakhla and Chassigny (SNC) meteorites come from, critical information for calibrating the age of volcanism within Tharsis, Elysium and Cerberus (Werner and Tanaka, 2011)? What is the origin of the Medusae Fossae Formation (MFF), which stretches more than 100° along the Equator between 135° to 240°E (Bradley et al., 2002)?

This paper reviews the issues related to these outstanding questions and offers some suggestions for going forward, both with assets already in orbit around Mars, as well as the new types of instrumentation that might be needed to facilitate improving our understanding of volcanism on the Red Planet. We focus on the questions identified above, but clearly there are others that also remain unanswered. With a subject as broad as volcanism, there will inevitably be topics that are not included in order to maintain a focus to this review. For example, in this review we do not dwell on the petrologic evolution of Martian magmas, or the chemical differences between volcanic rocks at the different landing sites investigated by the rovers on Mars. Interested readers are directed towards the reviews by McSween (2008) and Zimbelman et al. (2020b) for more details on Martian geochemistry and to Christensen et al. (2005) for the spectral evidence for chemical diversity of rocks across Mars. We also do not address the interaction of magmas and lava flows with ground ice, or the potential generation of lahars by this interaction (Christiansen, 1989; Pedersen, 2013), because there is reasonable agreement in the community about the processes associated with this interaction (e.g., Allen, 1979; Mouginis-Mark, 1985; Squyres et al., 1987; Head and Wilson, 2002, 2007; Fagents et al., 2002; Wilson and Mouginis-Mark, 2003; Hamilton et al., 2010; Cassanelli and Head, 2018).

Other volcano-related topics are not discussed principally because they relate to the erosional history of a volcanic construct. The aureole materials that lie beyond the Olympus Mons escarpment is one such example (Harris, 1977; Lopes et al., 1980). These aureole deposits may be related to the collapse of the lower flanks of the volcano, either due to the influence of shallow transient seas at the base of the shield (de Blasio, 2011, 2018), to erosion of the less competent basal materials (King and Riehle, 1974; Head et al., 1976) or to the role of ice within the basal materials (Hodges and Moore, 1979; Tanaka, 1985). Conversely, a tectonic origin for the different aureole lobes (Borgia et al., 1990) has also been proposed.

2. What we think we understand

Here we briefly consider aspects of Martian volcanism that have been the focus of numerous previous investigations. In general, there is reasonable agreement within the community as to the characteristics and significance of these aspects of volcanism on Mars, but questions inevitably remain even for these well-studied topics. These questions are considered in Section 3, and the data sets needed to potentially advance our understanding are covered in Sections 4 and 5.

2.1 Morphology of giant shield volcanoes

The morphology of the giant Tharsis volcanoes (Olympus, Arsia, Pavonis, Ascraeus, and Alba Montes) and Elysium Mons have been extensively studied (e.g., Carr, 1973; Malin, 1977; Hodges and Moore, 1994; Montesi, 2001; Greeley et al., 2000; Plescia, 2004; Ivanov and Head; 2006; Bleacher et al., 2007; Hauber et al., 2011; Platz and Michael, 2011; Richardson et al., 2017; Mouginis-Mark, 2018). The surfaces of these volcanoes are covered by lava flows that resemble terrestrial lavas, typically having much larger volumes (Mouginis-Mark and Yoshioka, 1998; Pasckert et al., 2012; Peters et al., 2021). Terrestrial shield volcanoes have commonly served as useful analogs for the evolution of Martian summit calderas (Mouginis-Mark et al., 2007, 2021), and connections to the volumes and locations of the near-surface magma chambers (Wilson and Head, 1994). Investigations of Hawaiian shield volcanoes (particularly Mauna Loa and Kilauea) led to predictions of magma chamber depth (Zuber and Mouginis-Mark, 1992), the periodicity of activity (Wilson et al., 2001), and the possible volume of magma produced during a single eruption (Mouginis-Mark and Wilson, 2019). Although the timing of the on-set of activity at specific sites remains unclear, treating these Martian volcanoes as giant equivalents of terrestrial shields has been productive (Carr and Greeley, 1980; Mouginis-Mark et al., 2007) and provided pathways for detailed analyses.

There are, however, several aspects of these large Martian shield volcanoes that remain enigmatic. Principal among the “unknowns” are their internal structure, in terms of the “plumbing system” (i.e., the distribution of dikes that once fed magma to the surface), the depth and size of the original magma chamber, and the reason that some volcanoes experience significant changes in their eruption style during the growth of the volcano (e.g., Alba Mons; Mouginis-Mark et al., 1988). Determination of the internal structure of a volcano is often a challenge even for well-studied terrestrial shields (e.g., Ryan et al., 1988), and typically relies upon on-going activity to produce seismic data (Lin et al., 2014) or surface deformation (Rymer and Williams-Jones, 2000).

At the largest scale, the Tharsis Montes and Alba Mons all have substantial secondary or satellitic eruptive centers that have implications for interior processes and structure. The rift aprons of the Tharsis Montes, which post-date the main shields, have extensive flow fields that form the vast complex of volcanic plains surrounding the larger Tharsis edifices (Crumpler and Aubele, 1978; Bleacher et al., 2007), and Alba Mons‟ western flank consists of a shield-shaped accumulation of volcanic materials (Plescia, 2004; Ivanov and Head, 2006) whose surface exhibits a vast flow field with numerous lava flows and lava tube systems (e.g., Cattermole, 1990; Schneeberger and Pieri, 1991; Crown et al., 2019).

We will return to the topic of the internal structure of Martian volcanoes in Section 3, but hints about their internal structure do exist. The pair of summit calderas at Alba Mons (Cattermole, 1990; Schneeberger and Pieri, 1991; Ivanov and Head, 2006) may indicate the migration of the magma chamber, but this does not explain the unusually low cross-sectional profile of the volcano, or the apparent high concentration of extensive lava tubes (Crown et al., 2019). Wrinkle ridges and circumferential graben within the summit caldera of Olympus

Mons are interpreted to indicate that the once-active magma chamber lay at a depth of ~16 km below the caldera floor (Zuber and Mouginis-Mark, 1992) and thus resided above Mars datum within the volcanic edifice. One of the smaller shields within the Tharsis region is the volcano Tharsis Tholus, which is notable not only for its very deep caldera but also because the volcano has, for no obvious reason, experienced major flank collapses on several occasions (Plescia, 2003a; Platz et al., 2011). These flank collapses may indicate that magma intruded into the flanks along rift zones caused over-steepening and subsequent collapse of the side of the volcano in a comparable manner to the generation of massive flank collapses of oceanic volcanic islands on Earth (Fiske and Jackson, 1972; Moore et al., 1994; Oehler et al., 2008). Outside the Tharsis region, the unusually steep upper slopes of Elysium Mons volcano (Malin, 1977) may be due to low-volume summit eruptions (Blasius and Cutts, 1981), although it appears likely that large eruptions also occurred from close to the summit (Wilson and Mouginis-Mark, 2001). Thus, the interplay of low-and high-volume eruptions may dictate the overall shape of a volcano in a comparable manner to that proposed for volcanoes in the Galapagos Islands (Rowland, 1996; Mouginis-Mark et al., 1996).

2.2 Emplacement of lava flows

Since the collection of images of Olympus Mons by Mariner 9, there have been many efforts to use the geometry of lobate lava flows to determine their rheological properties (e.g., Hulme, 1976; Moore et al., 1978; Zimbelman, 1985; Baloga et al., 2003; Hiesinger et al., 2007; Pasckert et al., 2012). Some flows may exceed 300 km in length, and the widths, thicknesses and dimensions of the flows and their associated lava channels can be used to unravel the relative roles of viscosity changes and concurrent formation of levees, stationary margins, and stagnant zones within the flow (Glaze and Baloga, 1998; Baloga et al., 2003; Glaze et al., 2009; Peters et al., 2021). In southern Tharsis, mapping studies revealed lava flows with textures analogous to 'a'ā and pāhoehoe lava flows on Earth, with implications for comparable diversity in flow emplacement styles and rates in Martian flow fields (Crown and Ramsey, 2017). Models for flow emplacement in turn may help to indicate the effusion rate and, hence, the duration of activity (e.g., Lopes and Kilburn, 1990). These latter parameters have inevitably been based upon the growing understanding of terrestrial lava flows, but nevertheless have been used to try to identify the source vents of long lava flows (Rowland et al., 2002, 2004). The results of these models are generally consistent with basaltic lava compositions.

Using terrestrial lava flows to model Martian flows warrants consideration of the effusion rate of Martian lavas, the possible distribution of pāhoehoe lavas, and the inflation of lava flows during eruptions that may have lasted months or years (Bleacher et al. 2017). Bruno et al. (1992, 1994) demonstrated that the fractal dimensions of the margins of lava flows could discriminate between 'a'ā and pāhoehoe flows, but this technique has not been applied to Martian lava flows. For terrestrial lava flows, the occurrence of lava tubes indicates relatively long-duration eruptions (Peterson et al., 1994; Stephenson et al., 1998; Calvari and Pinkerton, 1998) as the development of tubes take time to develop either by roofing over of a lava channel or the development of distributaries beneath a solid lava crust (Peterson et al., 1994). Thus, the identification of an extensive lava tube system on Alba Mons, Mars (Crown et al., 2022) suggests that these eruptions could well have lasted for weeks, months or longer. The characteristic platy-ridged morphology of the very young Cerberus Fossae lavas (Plescia, 2003b; Jaeger et al., 2007, 2010) was attributed to high effusion rate, low topographic gradients, and low viscosity pahoehoe with disrupted surface crusts and sometimes inflated margins (Keszthelyi et al., 2004). Specific aspects of flow morphologies in multiple locations in the Tharsis and Elysium provinces are interpreted as evidence of flow inflation (Giacomini, et al., 2009; Bleacher et al., 2017; Crown and Ramsey, 2017; Hamilton et al., 2018, 2020; Dundas et al., 2019), common to low-viscosity flows on Earth emplaced at relatively low rates (Rowland and Walker, 1990; Walker, 1991; Hon et al., 1994; Thordarson and Self, 1998). A lava flow north of Hrad Vallis presents clear evidence that at least a few Martian lava flows inflated during long-lived eruptions (Hamilton et al., 2018). However, it appears that the majority of individual lava flows on Mars are comparable to terrestrial 'a'ā flows, as well as the relative abundance of 'a'ā vs. pāhoehoe on Mars. This raises questions about what controls the diversity of flow types (and, hence, eruption conditions) that exist on Mars.

2.3 Age of volcanic activity

The general characteristics of the chronology of volcanism on Mars has been determined via decades of counting the number of superposed impact craters on volcanic materials (Neukum and Hiller, 1981; Hartmann and Neukum, 2001; Werner, 2009). The older highland volcanoes (e.g., Tyrrhenus and Hadriacus Montes) are thought to have experienced explosive eruptions (Carr et al., 1977; Greeley and Crown, 1990) which produced flanks made of easily-eroded ash deposits, while the younger volcanoes in the Tharsis and Elysium regions predominately produced lava flows (Carr et al., 1977). Impact crater size-frequency distributions can reveal the youngest lavas on a volcano (Hartmann et al., 1999; Richardson et al., 2017), as well as illuminate the relative timing of activity at Tharsis, Elysium, and the Circum-Hellas volcanoes (Werner and Tanaka, 2011; Tanaka et al., 2014b). Use of impact craters as small as ~16 m in diameter now revealed in high-resolution (25 cm to 3 m per pixel) images allows the degradational history of volcanic surfaces on Mars to be determined and chronologic investigations of Amazonian volcanism to be undertaken (Hartmann and Berman, 2000; Platz and Michael, 2011; Hauber et al., 2011; Berman and Crown, 2019). But a fundamental problem remains in that these ages are only relative or model dependent, and thus attempts to link the stratigraphy of volcanic units with the thermal history of Mars (e.g., Greeley and Spudis, 1981; Vaucher et al., 2009; Baratoux et al., 2011) can produce dramatically different results. The longevity of volcanism at a particular volcano cannot therefore be determined, although models of the cooling rate of magma within the "plumbing system" of a volcano can place constraints on the maximum time allowed between successive eruptions before the pathway to the surface freezes (Wilson et al., 2001). Such calculations suggest that the Tharsis volcanoes were built episodically with active phases lasting less than 1 Myr alternating with ~100 Ma "quiet phases". With assumptions for the impact crater production rate, absolute model ages can be estimated, but obtaining true absolute ages for Mars requires calibration as was done for the Moon with returned lunar samples (Hiesinger et al., 2012).

Calibrating the relative ages of Martian volcanic units is therefore critical not only for the analysis of magma supply rates for individual volcanoes (Wilson et al., 2001), but also for using the distribution of mapped volcanic units to infer the temporal magma production rate across the whole planet (Greeley and Spudis, 1981; Tanaka et al., 2014b). However, an absolute age date from a known location finally has been obtained for Mars by Farley et al. (2014) using radiogenic and cosmogenic noble gases in a mudstone on the floor of Gale Crater. They found that a K-Ar age of 4.21 +/- 0.35 Ga characterized a mixture of detrital and authigenic components and gave an unexpectedly old age for the rim of the crater. Although a remarkable accomplishment, such an age-dating technique depends upon a rover working on the surface of Mars, with the geographic scale of the sampled region probably too small to be of value for extrapolating to the kilometer- to tens of kilometer-scale needed for calibrating orbital crater counts. We further discuss this point in Section 5.2.

2.4 Volcanic meteorites from Mars

Martian meteorites, of which at least 262 have been found on Earth (Udry et al., 2020), are often referred to as the SNC meteorites, named after the meteorites Shergotty, Nakhla and Chassigny, the type meteorites recognized to come from Mars (Wood and Ashwal, 1981; Ashwal et al., 1982; McSween, 1985, 2008; Teiman et al., 2000; Bridges and Warren, 2006). The SNCs have either a basaltic mineralogy or are ultramafic cumulates crystallized from basaltic melts. These 262 samples of Mars evidently originated from at least 11 different ejection events (Udry et al., 2020), so their age range should represent typical surfaces on Mars. The majority of the rocks are of young (Amazonian) age, possibly because their ejection from Mars was sufficiently violent that relatively unweathered rocks could remain intact (Udry et al., 2020). The basaltic shergottites represent surface lava flows and have crystallization ages of ~170 Ma. Olivine-phyric shergottites are similar in mineralogy to other shergottites (i.e., rocks similar to Shergotty), but they contain 7 to 29 vol. % large olivine crystals and range in age from 170 to 575 Ma. Nakhlites and chassignites (named, respectively, after the Nakhla and Chassigny meteorites) have crystallization ages of ~1.3 Ga and are inferred to have solidified at depth either as sills or as cumulates (McSween and McLennen, 2014). The only Martian meteorites that have very old crystallization ages, and thus most likely originated from the ancient Martian highlands, are Allan Hills 84001 (~4.1 Ga) and NWA 7034 (~4.4 Ga) (Santos et al., 2015). The young ages of many of the SNC meteorites, as well as their mineralogy, strongly suggest that they originated from either the Tharsis or Elysium regions (Nyquist et al., 2001; Treiman, 2005), but the question of exactly where on the planet they come from remains unresolved (Mouginis- Mark et al., 1992b; Hamilton et al., 2003; Lang et al., 2009).

2.5 Physics of Martian eruptions

Wilson and Head (1994) provided the theoretical basis for the interpretation of basaltic eruptions on Mars. Their analysis considered not only the emplacement of lava flows once erupted, but also the generation of melt, the longevity of magma chambers, and the propagation of dikes beneath the surface. They proposed that the low atmospheric pressure on Mars would encourage the release and expansion of magmatic volatiles such that most basaltic eruptions on Mars would involve explosive activity at the vent. This could lead to formation of ash-laden convecting eruption clouds and/or much more local dispersal of coarser pyroclasts. Depending upon the eruption rate and size of the ejecta clasts, lava fountains might be opaque or semi-transparent (Head and Wilson, 1989), so that either a brittle scoria deposit consisting of cooled clasts could form, or clasts could land virtually uncooled to coalesce into a lava pond which in turn could feed one or more lava flows (Wilson et al., 2009). Individual lava flows within Tharsis have been modeled to infer the rheology of the melts (e.g., Baloga et al., 2003; Glaze and Baloga, 2006; Garry et al., 2007).

Numerous investigations have focused on the physics behind the widespread dispersal of pyroclasts from explosive eruption plumes that might have formed in the thin Martian atmosphere (Fagents and Wilson; 1996; Glaze and Baloga, 2002; Wilson and Head, 2007; Kerber et al., 2011). Glaze and Baloga (2002) reviewed the conditions under which volcanic plumes on Mars would rise in the atmosphere. They found that convective plume models are only valid to a height of ~10 km because of the low atmospheric density. They inferred that plumes could not rise to a sufficient altitude to produce globally-distributed ash deposits. However, Kerber et al. (2011, 2012) concluded that voluminous, fine-grained, friable deposits such as the Medusae Fossae Formation (MFF) could have a pyroclastic origin, with ash dispersed by global atmospheric circulation.

These numerical models include the possible interaction between magma and ground water and/or ice (Head and Wilson, 2002), as well as the roles of the past and present-day climate conditions (Hort and Weitz, 2001). Most of these investigations modeled generic eruptions on Mars, but Kerber et al. (2011) investigated the modeled dispersal of pyroclastic material from Apollinaris Mons as a possible origin for the MFF. Greeley and Crown (1990), Crown and Greeley (1993), and Gregg and Farley (2006) assessed explosive eruption models for Tyrrhenus and Hadriacus Montes. Crown and Greeley (1993) examined the potential for welding in Martian pyroclastic flows, but as noted by Broz et al. (2021), "the behavior of pyroclastic density currents on Mars is not fully understood and more work on the subject is needed".

2.6 Volcanic input into the atmosphere

Erupting volcanoes on Mars are considered to have been major contributors to the atmosphere, from the Noachian (Carr, 1986; Greeley, 1987; Halevy et al., 2007; Gillmann et al., 2011; Halevy and Head, 2014) until the Amazonian Period (Wilson and Mouginis-Mark, 1987; Plescia, 1993). The discovery of abundant sulfur at the Viking Lander 1 site was inferred to be due to degassing of the Tharsis volcanoes (Settle, 1979). By comparison with data collected from terrestrial basaltic eruptions such as the 2014 – 2015 eruption of Holuhraun volcano, Iceland (Stefansson et al., 2017; Ilyinskaya et al., 2017) and from Masaya volcano, Nicaragua (Rymer et al., 1998; Stix et al., 2018), the long-term impact of the equivalent or larger lava flow eruptions on Mars could have created a much thicker transient atmosphere than is present today (Meunier et al., 2012). Volcanic outgassing most likely included significant amounts of CO_2 and H_2O prior to ~3.5 to 2.0 Ga ago (Grott et al., 2011), and CO_2 could have contributed as much as 1 bar of atmospheric pressure depending upon mantle composition. Outgassing is expected to have strongly declined in the Hesperian

Period, and to have been insignificant during the Amazonian Period (Grott et al., 2011). It has been inferred (Craddock and Greeley, 2009) that the volatile content of Martian lavas could typically have been ~0.5 wt.% water, ~0.7 wt.% carbon dioxide, and ~0.14 wt.% sulfur dioxide, along with several other volatile constituents (HF, HCl, and H_2S). But the effects on the atmosphere depend, to a considerable extent, on the frequency and magnitude of individual eruptions.

2.7 Summary of "knowns" in Martian volcanism

After almost 50 years of detailed analysis of the surface of Mars, there is general agreement that the following aspects of volcanism are known:

1) The surfaces of the younger volcanoes resemble shield volcanoes on Earth;

2) Numerous lobate lava flows on the flanks of these volcanoes appear comparable to terrestrial 'a'ā flows;

3) Volcanic activity on Mars evolved from explosive pyroclastic eruptions in the Noachian to effusive activity in the Hesperian and Amazonian;

4) The SNC meteorites are volcanic and have surprisingly young ages compared with the ages inferred from crater counts;

5) Physics-based models can be constructed to predict eruption dynamics;

6) Volcanism no doubt played a significant role in the evolution of the Martian atmosphere.

3. Unknowns in Martian volcanism

This section raises questions about volcanic features on Mars that are less frequently discussed or are more problematic in terms of understanding their mode of formation and/or age. These questions pertain directly to the aspects of Martian volcanism that are raised in Section 2 and include the internal structure of the volcanic constructs, the emplacement of the ridged plains materials, and the nature of volcanism during the earliest period of Mars history (the Noachian Period). We also consider the potential role of explosive volcanism and the formation of the MFF, as well as the typical effusion rates of lava flows. Finally, the importance of absolute age dates of the volcanic units is discussed in the context of the ages of the meteorites from Mars.

3.1 Internal structure of Martian volcanoes

While the surface morphology of the giant Martian shields is well understood, few constraints exist to infer the internal structures of these same volcanoes. But the internal structure has a dominant role in the determination of where on the volcano individual lava flows might be erupted and will have a controlling influence on the size and depth of the magma chamber (Wilson and Head, 1994; Wilson et al., 2001). This in turn will affect the geochemistry of the erupted magmas (McSween, 2008).

An apparent rift zone was first identified in Viking Orbiter images (Crumpler and Aubele, 1978), and this must indicate some inter-connection between the volcanoes (Bleacher et al., 2007). Several lava flows in excess of 100 km (Baloga et al., 2003) emanate from this apparent rift zone. On Earth, mapping the distribution of vents on a volcano is an effective way to infer the internal structure remotely (e.g., Lipman, 1980; Rowland, 1996), as these vents would be surface expressions of pathways from the shallow magma source to the surface, and typically are represented in eroded volcanoes as exposed dikes. Orbital images of Mars using Context Camera (CTX) (Malin et al., 2007) and High Resolution Imaging Science Experiment (HiRISE) (McEwen et al., 2007) should be sufficient to identify such vents, but in reality it has proven difficult to distinguish flank vents a few tens to hundreds of meters in diameter from eroded impact craters, while surface break-outs from lava tubes may falsely give the impression of flank vents on the Tharsis volcanoes (Mouginis-Mark 2018).

One such attempt was made by Peters and Christensen (2017) to identify potential vents on the flanks of Olympus Mons, and resulted in 60 candidate vents being suggested, with the potential vents clustering at elevations between 8 and 14 km, and below 4 km relative to the Mars datum. But this approach failed to consider the topographic effects of potential flank eruptions building local high points, which have been seen on terrestrial volcanoes (Rowland and Garbeil, 2000) to be a strong indicator of subsurface magma pathways (such as Kilauea's East Rift Zone). In Fig. 1, we show a slope analysis of the flanks of Olympus Mons. It is evident that the potential vents identified by Peters and Christensen (2017) did not result in any change to the local or regional topography of the volcano, which would be expected if lavas had erupted from vents fed by subsurface dikes. Instead, these are places where lava appears to have originated as break-outs from lava tubes associated with lava flows that originated at the summit of Olympus Mons (Mouginis-Mark, 2018). If true, this would have significant implications for the interior structure of Martian volcanoes, as the lack of flank vents may be related to the larger sizes of Martian volcanoes and difficulties with maintenance of complex subsurface pathways at large scales (Wilson and Head, 1994; 2002). Future geophysical exploration may well help understand the nature of subsurface magmatic systems in Martian volcano interiors (Section 5.1).

There are few examples of exposed dikes on Mars; one of the best examples is located east of Hrad Vallis (Fig. 2) and appears to have originated from the NW flanks of Elysium Mons volcano. However, dike emplacement events have been invoked for the formation of several large fractures on Mars, including: graben hosting chains of collapse-craters to the east of Alba Mons (Scott et al., 2002); numerous graben in the Memnonia Fossae region that are radial to Arsia Mons (Wilson and Head, 2002), one of which acts as the source of Mangala Vallis (Wilson and Head, 2004); the complex depression Hrad Vallis (Wilson and Mouginis-Mark, 2003); and the collapse of the Olympus Mons caldera (Mouginis-Mark and

Wilson, 2019). Schultz et al. (2004), using Mars Orbiter Laser Altimeter (MOLA) (Smith et al., 2001) topography across a graben in Memonia Fossae, calculated that the surface expression is consistent with a subjacent dike that extends to ~18 km depth with a maximum dilation of ~95 m.

Terrestrial examples (Fig. 3) show that dikes are most often exposed at the surface when more easily eroded material into which the dike was first intruded is subsequently removed by erosion. It is expected that where deep erosion has taken place around some of the proposed “supervolcanoes” (Whelley et al., 2020), some evidence of the more resistant cores of the volcanoes should be visible. Dikes may also be exposed in vertical faces exposed by mass wasting. Commonly, when a dike is emplaced, it rises to the near surface along most of its length and the top stalls at a depth of a few tens of meters to a few hundred meters, depending upon the local crustal density distribution (Head et al., 2006). Elastic finite-element models of the stress fields of the Galapagos volcanoes suggest that diapiric magma ascent could promote both circumferential and radial dike emplacement (Chadwick and Dieterich, 1995). One of the rare places on Mars where this situation may exist is NW of Elysium Mons (Pedersen et al., 2010; Korteniemi et al., 2010; Hamilton et al., 2018); here, individual raised ridges a few hundred meters wide, ~20 – 30 m high, and tens of kilometers long have been identified by Mouginis-Mark and Wilson (2016).

An older system of possible exposed dikes exists in the northern Hellas region, where ridges ~700 m wide and 10-20 m high are intermittently traceable for ~600 km (Head et al., 2006). En echelon ridges oriented radially to the summit of Tyrrhenus Mons have been interpreted as dikes exhumed by mass wasting of the flanks (Gregg et al., 1998), and the concentric pattern of linear ridges and fractures as indicative of a vast ring dike system on Hadriaca Mons (Korteniemi et al., 2010). Numerous swarms of extensional tectonic structures are observed to be either radial to, or circumferential to, the central Tharsis area and, to a lesser extent, the Elysium Rise (Ernst et al., 2001). Some of these tectonic structures most likely indicate the paths of dikes and have been linked to eruptions at Cerberus Fossae (Berman and Hartmann, 2002; Plescia, 2003b; Head et al., 2003), Hrad Vallis (Wilson and Mouginis-Mark, 2003), and Mangala Valles (Wilson and Head, 2004).

It seems reasonable to expect that the walls of Valles Marineris would be an excellent place to search for exposed dikes. The geologic origin of the 4,000 km long, 200 km wide, and up to 7 km deep Valles Marineris may be a guide to the expected number of dikes. If the canyon evolved in a flood basalt setting (Leone, 2014), then the paucity of dikes seems reasonable. However, if the Canyon grew akin to a continental (or oceanic) rift system (Mege and Masson, 1996; Mongomery et al., 2009), then there is a massive paucity given the lack of currently identified features. As the canyons enlarged through landslides and transport of materials out of the Canyon (Tanaka and Golombek, 1989), it would be reasonable to expect that these old pathways would now be exposed either as vertical blades of more resistant rock, or as linear ridges on the floor of the canyon. However, only a few examples of likely dikes at the far eastern end of the canyon have been found (Flahaut et al., 2011) (Fig. 4). Ridges on the canyon floor remain enigmatic in origin due to the degree of erosion that masks the true thickness of the features as well as their limited horizontal length. It is also possible that the lack of identifiable exposures of dikes is simply due to the predominantly mafic volcanic nature of the crust of Mars: new dikes could have been intruded into older rocks of similar composition and physical properties, so there is limited differential erosion and denudation, resulting in a comparable lack of exposure relative to the Earth.

3.2 Emplacement of the ridged plains materials

Are the ridged plains volcanic? The morphology of the ridged plains within Lunae Planum, Hesperia Planum and Syria Planum is different from that of the volcanic constructs within Tharsis and Elysium as there are very few identifiable lava flow lobes in any of these localities (Fig. 5). Wrinkle ridges in the ridged plains can be explained by either "global-scale" compressional stresses related to the development of Tharsis (Phillips and Lambeck, 1980, Watters and Maxwell, 1986; Watters, 1991), or local "thinskinned" tectonics in Hesperia Planum (Goudy et al., 2005), Syrtis Major Planum (Raitala and Kauhanen, 1989), and Isidis Planitia (Ritzer and Hauck, 2009). Although lacking the lobate lava flows seen within Tharsis and Elysium, a volcanic origin for the ridged plains has been proposed based upon the similarity between the ridged plains on Mars and the lunar maria (Scott and Carr, 1978; Maxwell, 1982). Ridged plains materials within Syrtis Major provide some of the closest associations between the ridges and lava flows from Nili and Meroe Patera (Hiesinger and Head, 2004; Gregg et al., 2020).

It is possible that the lack of numerous identifiable lava flow lobes might be explained if they were formed by long-duration, low effusion-rate, low-viscosity eruptions. Candidate examples of lava flow inflation have been found in the eastern Tharsis plains (Bleacher et al., 2017), in southern Tharsis (Crown and Ramsey, 2017), and near Hrad Vallis (Hamilton et al., 2018). This may imply that some Martian lava flows can inflate and thus mask or remove lobate flow margins, raising questions about the possible origin of the ridged plains materials on Mars. Some insight into this type of lava emplacement may be gained from the mode of formation of the Roza Member of the Columbia River Basalts, which has an area of ~40,000 km^2 and a volume of ~1,300 km^3; initial studies of these flows inferred that they were emplaced as high effusion-rate flows (Swanson et al., 1975). However, detailed field analysis of the Columbia River Basalts flows by Self et al. (1996) and Thordarson and Self (1998) subsequently led to the reclassification of the Cerberus Plains to the south of Elysium Planitia area on Mars as a giant platy-ridged pāhoehoe flow field akin to the 1783 – 1784 Laki Flow Field in Iceland (Keszthelyi et al., 2000, 2004). If this were the case for the ridged plains of

Mars, it is therefore not surprising that surface structures of candidate inflated pāhoehoe flows have been buried or eroded. The implication of this idea that the ridged plains were formed by high-volume pāhoehoe lava flows is that individual Martian eruptions must have continued for many years, with the effects on the atmosphere lasting at least years or decades. Of course, if the ridged plains were formed via long-duration, highvolume, eruptions, there is still the issue of where the vents were located. There is also an apparent absence of exposed lava tubes within the ridged plains, which might be expected to have formed if the eruptions were comparable to large inflated flows on Earth (Stephenson et al, 1998). The type of observation needed to identify inflated lava flows within the ridged plains is described in Section 4.2.

3.3 Volcanism in the Noachian

What type of volcanism took place in the earliest preserved period of Mars history, namely the Noachian Period? Probable exposures of Noachian volcanic units or features include numerous constructs 50 – 100 km in diameter in the southern highlands (Xiao et al., 2012), plains units within the south polar region (Tanaka and Kolb, 2001), and the lower walls of Valles Marineris (Flaut et al., 2012). Noachian volcanic landforms within expanses of ancient cratered terrain may be partially to completely eroded by numerous degradational processes occurring over Martian history. However, the existence of many floor-fractured craters on Mars, which were probably formed by intrusive igneous activity, hints at widespread local activity during the Noachian (Schultz and Glicken, 1979; Bamberg et al., 2014).

At the local scale, the Spirit Rover encountered differentiated alkaline rocks in the Columbia Hills within Gusev crater, for which McSween et al. (2006) suggested a local magma source beneath the crater, while Lang et al. (2010) suggested that the flows in Gusev crater may have originated Apollinaris Mons. Rocks on the floor of Gusev are also basaltic in composition, and may date from ~3.65 Ga (Greeley et al., 2005). Silicic volcanic rocks have been found at Gale Crater by the Curiosity rover (Morris et al., 2016). The earliest eruptive records of the Tharsis and Elysium provinces, which are Hesperian in age, (Tanaka et al., 2014b) are poorly preserved due to overprinting by the voluminous younger volcanic units that characterize those regions. Spectral studies suggest that some Noachian volcanic units are distinctly olivine-rich, whereas the volcanics of the Hesperian Period show diagnostic absorptions of olivine-pyroxene (Wilson and Mustard, 2013) and that flood lavas may have characterized parts of Noachis Terra (Rogers and Nazarian, 2013).

A volcanic origin for many of the areally extensive bedrock exposures throughout the Martian highlands has been proposed (Edwards et al., 2009), but Rogers et al. (2018) favor the interpretation that they are either impact-generated materials or detrital sedimentary rocks, or both. Dohm et al. (2009) proposed that the Claritas Rise to the SE of the Tharsis region was the result of tectonic uplift during the Noachian Period. Magnetic anomalies in the ancient southern highlands could have been formed by repeated dike intrusions (Nimmo, 2000), although alternative hypotheses have been proposed for these anomalies (Fairen et al., 2002). Regional geologic mapping from Viking Orbiter images revealed many volcanic features of the western hemisphere of Mars, including those in the southern highlands (in the Phaethontis and Thaumasia quadrangles) (Scott and Tanaka, 1981; Scott, 1982).

Some insight into the earlier phases of Martian volcanism can be obtained from the compositional data collected by the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) data (Murchie et al., 2007). Such data have been used to investigate the interior walls of Valles Marineris (Flahaut et al., 2011, 2012) and the central peaks of impact craters within plains units (Quantin et al., 2012). Both of these settings provide the opportunity to explore the spectral properties of materials that may originate at depths of several kilometers below the present-day surface. Flahaut et al. (2012) concluded that the western part of Valles Marineris may be cut into materials with physical properties consistent with them being lavas or volcanic sediments. Uplifted materials within the central peaks of impact craters (Quantin et al., 2012) reveal that there may be an accumulation of ~18 km of volcanic materials within the Tharsis region and that there is a major geologic discontinuity beneath this volcanic pile.

The longevity of any volcanic center in the Noachian is also poorly known. Viking Orbiter-based geologic investigations of the highland volcanoes and Alba Mons (e.g., Greeley and Spudis, 1981; Mouginis-Mark et al., 1988, 1992a; Crown and Greeley, 1993) discussed potential transitions from early explosive eruptions to later effusive eruptions, both at individual volcanic centers and across the Martian surface in general. Potential "super-volcanoes" in northern Arabia Terra were proposed by Michalski and Bleacher (2013). Such super-volcanoes would be consistent with the interpretation of the friable materials in Arabia Terra being volcanic ash deposits (Chapman, 2002; Kerber et al., 2012; Whelley et al., 2021). However, even though Arabia Terra displays considerable evidence for erosion of the surface that might amount to hundreds of meters of material removed, there is no evidence for intrusives around these structures, and indeed there is no consensus yet regarding whether these features are even volcanic in origin, casting the role of central-vent volcanism during the Noachian Period somewhat in doubt.

3.4 Source Craters for the SNC meteorites

Where on Mars did the SNC meteorites come from? Early calculations (Vickery and Melosh, 1987) suggested that the SNC meteorites were ejected from a large (>100 km dia.) impact crater on Mars about 200 million years ago, but more recently smaller (<30 km diameter) impact craters have been suggested (e.g., Lang et al., 2009). Even the number of impact events remains a question of debate when searching for the source crater(s). Udry et al. (2020) inferred at least 11 discrete ejection events but made no predictions of where their craters might be located.

Attempts have been made to identify the source crater(s) using Viking Orbiter images (Mouginis-Mark et al., 1992b), Thermal Emission Spectrometer (TES) spectra (Hamilton et al., 2003) and a combination of multiple remote sensing data sets (Lang et al., 2009). Indeed, Hamilton et al. (2003) identified the best spectral matches to the SNCs in regions of Mars that lack young volcanic flows, such as Nili Fossae and Eos Chasma. But using Thermal Emission Imaging System (THEMIS) infrared (IR) and TES data, Lang et al. (2009) noted a lack of correlation between Tharsis lava flows and the mineralogy of the SNCs. However, most of the Tharsis region is obscured by a mantle of dust >1 m thick (Christensen, 1986), making it difficult to obtain comprehensive compositional measurements there.

The issue of where the SNC meteorites originate is crucial to understanding the absolute age of relatively recent volcanism on Mars. Because of the clustering of meteorite ages, with the vast majority having young (<~1.3 Ga) ages, the chronology derived from crater-counts of different geologic surfaces (Hartmann, 2005) would require that the SNCs all originated from the same (young) chronostratigraphic surfaces. Werner and Tanaka (2011) defined the different age epochs on Mars based upon crater size-frequency data, and inferred those surfaces produced during the Late Amazonian are younger than 0.235 Ga, surfaces in the Middle Amazonian are younger than 0.880 Ga, and Early Amazonian surfaces are younger than 3.0 Ga. This classification would place most of the SNCs either in the Late or Middle Amazonian. Ejection ages can also be derived for the SNCs using cosmic-ray exposure ages (the time the rocks were in space after their ejection from Mars) and their terrestrial ages (the residence time on Earth prior to collection) (Nyquist et al., 1998; McSween and McLennan, 2014; Udry et al., 2020). There are three distinct ejection events (at ~11 Ma, ~2.6 Ma, and ~0.5 Ma), which would all imply that the source impact craters would be morphologically fresh (and, of course, be located on young volcanic targets) (Bogard et al., 1984). Zunil crater (Tornabene et al., 2006) and Tooting crater (Mouginis-Mark and Boyce, 2012) may fit the criteria of young crater age, young volcanic targets, and abundant secondary craters, while their diameters (10.1 km and 28.5 km, respectively) mean that their excavation depths (~5% of the crater diameter; Cockell and Barlow, 2002, their Fig. 1) may have been entirely within a sequence of volcanic flows. It seems plausible that the youngest SNCs may either come from Cerberus (i.e., Zunil crater) or from Tharsis (i.e., Tooting crater). Age differences in the meteorites may be representative of the two volcanic regions, with the basaltic and lherzolitic shergottites (age ~165 to 475 Ma) coming from Tharsis; and naklites and Chassigny (ages of ~1.30 to 1.35 Ga) coming from Elysium. This would require that the Hesperian volcanic rocks mapped by Tanaka et al. (2014b) are in fact much younger than previously inferred. Alternatively, the shergottites may come from the young lava flows within Cerberus Fossae (Berman and Hartmann, 2002) and the naklites and Chassigny from Tharsis. Thus, the Elysium volcanics may not be represented in the meteorite collection at all, potentially because they are sufficiently weathered that they could not survive the stresses of ejection from Mars or the landing on Earth (Fritz et al., 2005; Udry et al., 2020).

3.5 Explosive volcanism

What was the role of explosive volcanism in the geologic evolution of Mars, and where might pyroclastic deposits be observed? Early evidence from Mariner 9 images (West, 1974) and Viking Orbiter images (Reimers and Komar, 1979) indicated that during early Martian history certain volcanoes had erupted explosively. Based upon numerical models (Wilson and Head, 1994), pyroclastic deposits on Mars should be mafic and unwelded. If they existed, large volumes of mafic pyroclasts would be expected to be converted to eolian materials very quickly (Ye and Michalski, 2021). After several decades of Mars exploration (starting with

West, 1974, and many authors since then) the evidence that Martian volcanoes erupted explosively is less convincing than that for effusive activity (Broz et al., 2021, and references therein). This may be due in part to better preservation of primary characteristics in lava flows than in pyroclastic deposits, but also possibly due to a true predominance of effusive volcanism relative to explosive volcanism or to the prevalence of early explosive volcanism that is now obscured by degradation and burial.

The models mentioned in Section 2.5 do not provide sufficient detail to confidently identify landscapes produced by explosive activity in early Martian history, although investigations into possible recent explosive eruptions (such as one in Cerberus Fossae; Horvath et al., 2021) may allow quantitative analyses of eruption characteristics. Nevertheless, the identification of areally extensive air-fall deposits (Hynek et al., 2003) or pyroclastic flow deposits is important as the thin atmosphere and lower gravity on Mars should favor magma fragmentation and, hence, the production of friable pyroclastic materials (Wilson, 1984; Wilson and Head, 1994). For example, if Tyrrhenus and Hadriacus Montes, which are the most often cited candidates for extensive explosive eruptions (West, 1974; Peterson, 1978; Greeley and Spudis, 1981; Greeley and Crown, 1990) and are referred to as the Highland Montes, were formed by explosive eruptions, it would be expected that thick layers of material would be seen at the summits, and that for airfall deposits these layers would thin with radial distance from the summits. CTX images of the flanks of Tyrrhenus Mons (Fig. 6) and Hadriacus Mons reveal individual layers, but the physical properties of this material are hard to determine, as the erosional mechanism exposing these layers is not apparent. Age constraints from crater counts suggest a long duration of activity for Tyrrhenus Mons, with older explosive eruptions forming the edifice (Late Noachian) and later (Hesperian) eruptions emplacing the lava flow field on the southwest flank (and likely including eruptions at the summit) (Crown and Greeley, 1993; Crown et al., 2005b; Williams et al., 2008). However, the edifice-building eruptions appear to have taken place over a limited time period; inspection of the layers exposed on the surface by flank erosion shows that there are no partially exhumed impact craters revealed beneath the current surface. Thus, it seems unlikely that paleosurfaces were exposed on flat surfaces for extended time periods, because few impact craters formed on these earlier surfaces. On a more regional scale, the fact that there is a lack of distinct pyroclastic layers across the surface of Mars (in caldera walls, flank scarps, etc.) limits our ability to study explosive volcanism there.

How could we confirm or refute the idea that certain volcanoes on Mars experienced explosive eruptions, and hence were formed by different magma types (either possessing a higher volatile content or a different silica content) from those forming the giant shields such as Olympus and Arsia Montes? Francis and Wood (1982) suggested that the Highland Montes (Hadriacus and Tyrrhenus Montes) represent a unique style of volcanism driven by volatiles derived from mantle sources by a mechanism that stopped coincident with the period of maximum planetary degassing. Basaltic highly-explosive (plinian) eruptions were likely to have occurred throughout Martian history (Wilson and Head, 2007), but the height to which the plumes were ejected and thus the regional dispersal of the ejecta remains enigmatic (Fagents and Wilson, 1996; Glaze and Baloga, 2002). Good insight into the (relatively) recent activity at Arsia Mons can be gleaned from inspection of a graben to the west of the summit caldera (Mouginis-Mark and Rowland, 2008). Here (Fig. 7) can be seen a section almost 1 km high that includes >1,400 individual lava flows. Contrary to the suggestion that there has been explosive activity at this volcano (Mouginis-Mark, 2002; Hynek et al., 2003), this graben expose no units that could be interpreted as thick ash layers.

The identification of pyroclastic deposits from early in Martian history becomes particularly difficult, but knowledge of the spatial extents and thicknesses of putative ash deposits would enable the magnitude of the eruption to be estimated. Pan et al. (2021)

proposed that olivine enrichments in certain crater floors may arise from widespread pyroclastic materials. Whelley et al. (2021) and Chu et al. (2021) may have identified several large and deep craters in western Arabia Terra that could be explosive calderas, capable of producing vast layers of ash. Using spectral data, mineral signatures of ash deposits have been interpreted to imply thinning from 1 km to 100 m thickness with increasing radial distance from the depression (Whelley et al., 2021). However, the interpretation of these depressions as calderas remains highly controversial and highlights the problems in the identification of ancient pyroclastic materials.

What would the alternatives be? Highly energetic phreato-magmatic eruptions due to magma/water interactions such as the one inferred for Hrad Vallis (Wilson and Mouginis-Mark, 2003) may have been possible, and might have produced landforms that are morphologically different from other types of explosive volcanism; for example, excess water during these eruptions at Hrad Vallis may have led to water channels being formed. The general environments and geological settings of magma-water interactions were reviewed by Head and Wilson (2002), and it appears reasonable to ask if all of the volcanoes on Mars where explosive volcanism has been hypothesized (e.g., Hecates Tholus; Mouginis-Mark et al., 1982) absolutely require explosive volcanism. For example, could Hecates Tholus and Ceraunius Tholus be mantled by glacial deposits that mask the underlying lava flows? Fassett and Head (2006, 2007) proposed that the valleys on Hecates Tholus were formed by basal melting and run-off from a summit snowpack, consistent with the interpretation by Gulick and Baker (1990) that these valleys are fluvial in origin, with the summit morphology subdued by a glacial lag-deposit. This contrasts with the hypothesis of Mouginis-Mark et al. (1982) that Hecates Tholus experienced major explosive eruptions; the interpretation of Fassett and Head (2006, 2007) now seems the more plausible origin for the summit morphology and the valleys.

3.6 Origin of the Medusae Fossae Formation (MFF)

The potential for extensive pyroclastic volcanism on Mars may also pertain to the origin of the Medusae Fossae Formation (MFF), which is one of the most enigmatic features on Mars? The MFF is an unconformable deposit covering at least ~2.2 million km^2 (Bradley et al., 2002; Mandt et al., 2008); this is >1% of the entire surface of Mars. The MFF consists of several large outcrops that are prominent across almost 100 degrees of longitude (Scott and Tanaka, 1982, 1986; Greeley and Guest, 1987). There is a diverse array of hypotheses for the origin of the MFF (Zimbelman et al., 2020c), with the leading ideas including a volcanic origin, either as ignimbrites (Scott and Tanaka, 1982), or ashfall deposits (Hynek et al., 2003; Kerber et al., 2012). Even if the MFF is volcanic ash (originating either from Tharsis and/or Apollinaris Mons), the question remains as to why there is no evidence for significant ash deposits elsewhere on Mars. Specifically, if the MFF originated from volcanoes within Tharsis, graben on the flanks of Arsia Mons (Mouginis-Mark and Rowland, 2008), as well as sections of Valles Marineris, should reveal thick deposits of ash, but they do not. Conversely, Kerber et al. (2011, 2012) proposed that Apollinaris Mons was the site of this explosive activity and that the MFF was dispersed by the global atmospheric circulation. Why Apollinaris Mons experienced this type of activity when other volcanoes evidently lacked such an explosive history is not clear. Also, the volume of the Apollinaris Mons construct is orders of magnitude smaller than the estimated volume of the MFF deposits, which does not seem consistent with the formation of even the largest-volume terrestrial caldera formation mechanisms (Linsay et al., 2001; Willcock et al., 2013).

Ojha et al. (2018) argued that all of Mars‟ dust may have originated from the MFF. Indeed, Gale Crater (5.4°S, 137.8°E), whose floor is the landing site of the Curiosity rover, contains a 5.2 km-high central mound of layered material that is largely sedimentary in origin and may be part of the MFF (Thomson et al., 2014). The age of the layered mound is estimated to be ~3.6 – 3.8 Ga which straddles the Noachian-Hesperian time-stratigraphic boundary (Thomson et al., 2014). As can be seen in Fig. 8, material at the summit of the mound superficially resembles the MFF. If this analogy is correct, then it raises the question of the origin of crater mounds around the planet (Bennett and Bell, 2016).

There is good evidence that the current exposures of the MFF are the erosional remnant of a previously more extensive deposit (Chapman, 2002; Zimbelman et al., 2020c). Allen (2007) used pedestal craters (craters surrounded by a crenulated-margin plateau; McCauley, 1973) within the MFF to determine that the deposits were once 1.2 to 1.9 times more voluminous than mapped MFF exposures. Mapped outliers of MFF materials indicate that the MFF once covered an area 2 to 4 times greater than that of the present exposures (Dunning, 2018), and that at least the western portion of the MFF is Hesperian in age (Zimbleman and Scheidt, 2012). Campbell et al. (2021) employed the SHAllow RADar (SHARAD) radar to measure the real permittivity and loss tangent of the MFF and concluded that the unit is a two-layer deposit, with 300 – 600 m of fine-grained, self-compacting material above up to 2 km of minimally compacting, low-loss material. The 1.4 x 10^6 km^3 volume of the current MFF exposures (Bradley et al., 2002) should clearly be considered as a minimum.

Indeed, the MFF may be comparable to a terrestrial ignimbrite deposit (Scott and Tanaka, 1982; Mandt et al., 2008). Ignimbrites are produced when a dense mixture of volcanic gas and pyroclasts is erupted into an atmosphere and forms a collapsed, fountain-like structure over a vent feeding ground-hugging flows (Sparks et al., 1978). An ignimbrite flare-up in the southwestern United States involved multiple pulses between 18 and 49 Ma resulting in 0.4 x 10^6 km^3 of erupted materials (Best and Christiansen, 1991; Cather et al., 2009). The Sierra Madre Occidental accounts for three-fourths of this total, with a peak eruptive flux of 3 x 10^4 km^3 Ma^{-1} at 32 Ma and an average eruptive flux of 1.8 x 10^4 km^3 Ma^{-1} between 27 and 38 Ma

(Cather et al., 2009). Toba represents the largest single explosive deposit on Earth, and emplaced 2,800 km^3 of material during a caldera-forming event at 74 Ka (Rose and Chesner, 1990; Chesner et al., 1991). The current volume of the MFF would require 500 eruptions equivalent in volume to Toba over the 2.9 Ga interval within which the MFF was emplaced. Similarly, nearly 100 Sierra Madre Occidental events would be required to produce the current MFF volume. The sheer size of the MFF remains a challenge for the ignimbrite origin hypothesis, and Ohja et al. (2018) have suggested that it may once have a much larger volume and be the source for much of the dust on Mars, even though there is a lack of caldera structures associated with it (Francis and Wood, 1982). Some depressions in the Martian highlands may be the remnants of giant calderas (Bernhardt and Williams, 2021; Whelley et al., 2021), although this interpretation remains controversial. Add to this volume issue the likely basaltic composition of the Martian deposits, as compared to the predominantly silicic terrestrial ignimbrites, and this also casts doubt on the ignimbrite origin of the MFF (Mandt et al., 2008; de Silva et al., 2010).

Thus, there is still great uncertainty regarding the origin of the MFF. The ashfall hypothesis appears unlikely given eruption column stability considerations and the great thickness of the deposit at great distances from potential sources, although transport and redeposition by the wind is possible. An origin as multiple ignimbrites can better explain erosional morphology and accumulation of thick deposits, but the lack of vents is a serious issue with this hypothesis. More speculative origins have also been suggested for the MFF, including paleo-polar deposits (Schultz and Lutz, 1988) or layers of rafted pumice (Mouginis-Mark and Zimbelman, 2020). An approach to resolving the origin of the MFF is discussed in Section 5.4.

3.7 Summary of "unknowns" in Martian volcanism

1) We have virtually no idea of the internal structure (i.e., the magma pathways) within a Martian volcano;

2) Current datasets are limited in their ability to search for evidence of inflated lava flows within the ridged plains materials. Thus, their potential origin as volcanic flows, as well as the location of the associated vents, remains unresolved;

3) Volcanic rocks of Noachian age have been found across the highlands and specifically at Gusev crater. But the style of volcanism (as basic as were the eruptions explosive or effusive) cannot be identified due to extensive erosion;

4) It is not clear how widespread pyroclastic deposits without associated source vents might be identified given the rapid weathering on Mars;

5) The origin of the Medusae Fossae Formation remains unresolved;

6) We have little understanding of where on Mars each of the 11 craters that ejected the SNC meteorites are located. This places the absolute calibration of the impact cratering flux into uncertainty, and an explanation must be found to explain why almost all of the meteorites are of Amazonian age.

4. Prospects for progress using existing data sets

Further analysis of Mars Odyssey, Mars Reconnaissance Orbiter, ExoMars Trace Gas Orbiter and Mars Express data will no doubt reveal interesting new aspects of the geology and volcanology of Mars. Here we focus on topics where progress might be made with these data sets to answer some of the questions outlined in Section 3. Topics include searching for dikes within the walls of Valles Marineris to investigate the internal structure of the canyon, the morphology of the ridged plains materials to help determine if they are volcanic, and the chemical composition of the MFF.

4.1 Internal structure of volcanoes: Investigating the walls of Valles Marineris

The internal structure of volcanic regions may be further explored by additional imaging from sensors already in orbit around Mars. HiRISE images, with a spatial resolution of ~25 cm/pixel, are essential for the recognition of dikes a few meters in width. New HiRISE images would also allow testing of the idea proposed by Lucchitta (1990) that there was late-stage volcanism along the walls of Coprates Chasma, or could extend the area of analysis of Brustel et al. (2017) who found more than 100 dikes in eastern Coprates Chasma. Color HiRISE images (e.g., Fig. 4) are particularly helpful in recognizing these dikes. Such a search should not only include the areas identified by Lucchitta (1987, 1990), but also include dark, spatially continuous layers that might be extensive ash deposits (Geissler et al., 1990). Mangold et al. (2010) studied two recent volcanic units in Noctis Labyrinthus, which are typically basaltic in composition. They proposed that the Valles Marineris experienced a phase of mafic or ultramafic pyroclastic volcanism, supporting the idea of Lucchitta (1987, 1990) who suggested that pyroclastic activity might have been geologically quite young (Amazonian in age). Fig. 9 illustrates what was interpreted to be a line of pyroclastic deposits erupted from vents aligned along a basal fault (Lucchitta, 1987, 1990). Ubiquitous horizontal layering to depths of at least 8 km within the canyons was identified by McEwen et al. (1999), who concluded that either volcanic or sedimentary processes were much more important early in Martian history than previously accepted, as prior to these observations it was hypothesized that the walls should include mainly unlayered or coarsely layered megabreccias from impact cratering events. The identification of columnar joints in impact crater walls elsewhere on Mars (Milazzo et al., 2009) has been used to argue for lava flows that were rapidly cooled by liquid water, and similar features may also have existed at times when the units now exposed in the canyon walls were emplaced.

4.2 Searching the ridged plains to identify vents and flow lobes

To test if the ridged plains are indeed volcanic, one of the most effective methods for the detection of subtle topographic features such as lava flow fronts is the use of digital elevation data to produce de-trended topographic maps. Head et al. (2002) employed this technique with high-pass filters applied to gridded MOLA topographic data to identify structural features within the North Polar region. This approach also proved successful in finding low-amplitude topographic features on the Moon from de-trended Lunar Orbiter Laser Altimeter topographic data (Kreslavsky et al., 2017). However, this methodology has only been applied to regional- to global-scale topographic data sets. In recent years, much higher spatial resolution digital topographic data sets have been produced for areas on Mars using either HiRISE or CTX data, and the potential exists for using the Colour and Stereo Surface Imaging System (CaSSIS) on the ExoMars Trace Gas Orbiter to also produce local digital elevation models (DEMs) (Thomas et al., 2017). A systematic campaign of CTX stereo coverage might also be used to get the same results. Such DEMs for areas of ridged plains materials could be used to search for vents and flow lobes.

A second method to interpret the morphology of the ridged plains would be the acquisition of low-illumination-angle images of these units. Such images might then be used to apply statistical methods developed by Glaze et al. (2005) to investigate the potential distribution of tumuli on candidate pāhoehoe surfaces within the ridged plains. At least for terrestrial volcanism, the main differences between pāhoehoe and 'a'ā are controlled by volumetric flow rate (Rowland and Walker, 1990), and it is likely that a similar situation would exist on Mars, even if the transitional flow rate was different. However, because the Martian atmosphere might pose a problem due to atmospheric scattering at these incidence angles, few images from instruments such as THEMIS visible (VIS) and CTX have been collected under low solar lighting. And yet, experience with imaging lunar maria (Moore, 1972; Head and Lloyd, 1972, 1973) shows that subtle topographic features associated with lava flows and tectonic structure can be identified under shallow illumination. It might also prove possible to identify different units within plains units (e.g., Fuller and Head, 2002) based on surface morphology coupled with age estimates from crater counts, to help determine the spatial evolution of the units within the larger expanse of ridged plains. However, it is recognized that to collect such low-illumination-angle data, the timing of the equatorial-crossing of the spacecraft would need to be changed, and this might impact the operation of other sensors on the spacecraft.

4.3 Composition of ejecta from fresh craters formed within the MFF

Clues to the origin of the Medusae Fossae Formation might well be gained if there were better compositional data for the deposit, but a major issue is how such data could be collected given the ubiquitous dust cover? One approach might be to collect these spectral data soon after the detection of new impact craters on the MFF. Daubar et al. (2013) have demonstrated that new impact craters formed during the spacecraft era can be identified in certain parts of Mars where albedo variations make for easy detection. Spectral data from CRISM and TES data might show whether the MFF is silicic and therefore more likely to be the product of explosive volcanism. If the MFF is comprised of fresh basalt, this would suggest that it is an ash deposit, potentially produced by explosive basaltic volcanism (Hynek et al., 2003; Kerber et al., 2011); deposits formed by basaltic plinian activity are likely to be common on Mars, so mafic ash fall deposits should be common (Wilson and Head, 1994).

This approach has been taken by Viviano et al. (2019) to investigate the composition of lava flows within Tharsis and Elysium using CRISM data. Comparable data for craters formed within the MFF might reveal its composition. As an example, a 600 m diameter impact crater within the MFF is shown in Fig. 10. The secondary craters from this primary impact indicate that this is a relatively recent impact event. Unfortunately, the infrared detectors on CRISM are no longer operational (Kestay, pers. Comm., 2021), but if infrared spectral data comparable to CRISM could be obtained for a fresh crater before it becomes buried by dust, then perhaps the underlying composition could be detected. Of course, this approach would have wide applications to other volcanic regions as well (Viviano et al., 2019).

5. What new data sets are needed?

This section is more speculative, and draws on the methods employed to study volcanoes on Earth with instrumentation that has not yet been flown to Mars. None of the techniques are totally new, and some have already been developed with the specific objective of planetary exploration. They therefore offer some possibilities for the further analysis within the coming few decades of some aspects of volcanism discussed in Section 3.

5.1 Higher-resolution gravity data

The internal structure of volcanoes on Earth is often inferred in part from the gravity signatures of rift zones and magma chambers (Rymer, and Williams-Jones, 2000; Geist et al., 2006). For example, using 3,300 field, shipborne and airborne gravity measurements, Kauahikaua et al. (2000) developed a simplified three-dimensional model of the subsurface structure of the island of Hawai'i, and they were able to identify linear gravity anomalies that were interpreted to be expressions of dense cores of previously unrecognized rift zones lacking surface expressions. Such regional gravity data with a spatial resolution of a few hundred meters to over a kilometer, for example, the area to the NW of Elysium Mons, might confirm the existence of giant dikes extending towards Hrad Vallis (e.g., Pedersen et al., 2010). However, collecting such gravity data over volcanic regions of Mars could prove challenging, as the Martian atmosphere precludes the deployment of low-altitude satellites comparable to the GRAIL gravity mission to the Moon (Zuber et al., 2013). Surface measurements would be needed using a specifically targeted rover. Alternatively, gravity data at the spatial resolution needed (most likely in the hundreds of meters horizontal scale) might be obtained from a long-duration drone like the Dragonfly drone to be sent to Saturn‟s moon Titan in 2028 (Lorenz et al., 2018). Engineering problems would arise here, as such a drone might have to fly at an elevation of a few kilometers above Mars datum in order to accommodate the elevation variability typical of volcanic regions on Mars (e.g., SW of Arsia Mons or NW Elysium). However, the ubiquitous thick dust cover around Tharsis (Christensen, 1986) may also be hazardous to drone flights that could require landing areas on the dusty MFF surface. As a result of the thin Martian atmosphere, drone-based gravity data may never be obtained except for the areas at elevations close to the Mars datum (Smith et al., 2001) such as the low-lying and relatively flat young lava flows in the Cerberus Fossae region.

5.2 Imaging radar data for the ridged plains and the MFF

Synthetic aperture radar (SAR) data could be employed to search for structures within the ridged plains and MFF that might aid analysis of their origin. Not only would additional information on surface texture be obtained, but also subtle structures such those associated with lava pits and lava-rises (Walker, 1991) could be mapped. Imaging radar could be used to map differences in lava textures that are related to their effusion rate (Gaddis et al., 1989).

This use of SAR proved to be spectacularly successful during the Shuttle Imaging Radar-A (SIR-A) experiment (McCauley et al., 1982, 1986), which demonstrated that radar penetration to a few meters in the hyper-arid sands of the Sahara Desert, and such penetration depths should also be possible on Mars. Some radars have already flown in Mars orbit. The

Mars Advanced Radar for Subsurface and Ionospheric Sounding (MARSIS) (Jordan et al., 2009) and SHARAD (Seu et al., 2007) are both sounding radars that can probe the subsurface of Mars to depths of from hundreds of meters to a few kilometers, and have revealed some important attributes of the MFF (Watters et al., 2007; Carter et al., 2009). However, neither MARSIS nor SHARAD produce an image comparable in resolution to one obtained by a SAR.

The ability of an imaging radar to penetrate a dry surface is a function of the wavelength used. The SIR-A data were collected at a wavelength of 24-cm (L-Band), which is longer than that of other planetary radars such as the Mini-RF (Moon), Cassini SAR (Titan) and Magellan (Venus) radars. The exciting possibility of peering beneath the layers of dust that cover much of Mars (Christensen, 1986), so that near-surface volcanic structures were detectable, could reveal textural differences in lava flows that might provide additional information on their emplacement (e.g., Gaddis et al., 1989; MacKay and Mouginis-Mark, 1997). A Mars radar system that would have a penetration depth of a few meters would require a longer wavelength radar than that flown to Titan or Venus and thus a proportionally larger antenna but operating at this wavelength would make the orbital control needed for interferometric passes easier to accomplish.

Another advantage of a SAR is the ability to operate in an interferometric mode (Massonnet and Feigl, 1998). On Earth, this method has been particularly successful at quantifying volcano deformation following a new eruption (Amelung et al., 2000; Jung et al., 2011), as well as mapping the change in the surface structure due to the emplacement of new lava flows (Zebker et al., 1996). While it is highly unlikely that current activity would be detected on Mars, an additional product derived from an interferometric radar is the digital topography generated at the spatial resolution of the radar image (Farr et al., 2007). Such high resolution (potentially, 10 – 30 m/pixel) digital topography for all of Mars would be a significant advance over the MOLA laser altimeter data set and would undoubtedly provide greater insight into volcanism on the Red Planet (Glaze and Baloga, 2007; Hauber et al., 2009; Wilson et al., 2009).

5.3 Return samples from known locations and their ages

One of the most interesting discussions of the crater-density and absolute-age boundaries of units on Mars was presented by Werner and Tanaka (2011). As was discussed in Sections 2.3 and 3.8, the ages obtained for the SNC meteorites have an array of values, but in general they indicate volcanic activity in a period that crater chronology would classify as the Amazonian (Udry et al., 2020). But until returned samples can calibrate the ages of any materials from known locations on Mars, such as has been done for units on the Moon (Hiesinger et al., 2003; Robbins, 2014), there will be significant uncertainties in the ages of the latest activity (e.g., Hartmann et al., 1999), the longevity of activity at any one volcano (Vaucher et al., 2009), or the timing of volcanism across the planet (Tanaka et al., 2014a, b). Following the successful landing of the Perseverence rover in February 2021, a Mars sample return is planned by the United States and Europe, but the likelihood seems low that these samples will come from a landing site within one of the key volcanic stratigraphic units on Mars, due to the incentive to return samples with potential bio-signatures. Returning samples from known locations, preferably from at least three areas would be preferable. Samples from Cerberus Fossae (Amazonian), the ridged plains (Hesperian) and Tyrrhenus Mons (Late Noachian/Early Hesperian) would provide important absolute calibration points for the middle period of Martian geology.

An alternative to age dating returned samples from Mars is to develop and land spaceflight instruments specifically for dating surfaces in situ (Cohen et al., 2019). The continuing efforts to develop a compact resonance ionization mass spectrometer for in situ age dating of a planetary surface (Anderson et al., 2015, 2000) offers a potentially cheaper method to provide this age calibration. Missions flying this spectrometer would still need to be targeted to geologic areas identified by regional mapping as important chronostratigraphic units (e.g., Tanaka et al., 2014b; Werner, 2019), and would require surface mobility to acquire samples from multiple localities, but this target selection would not be complicated by the need to also identify potential bio-signatures at the same site.

5.4 Penetrators within the MFF

Some key geographic areas of high volcanological interest will almost certainly lie outside the realm of either a rover or a return sample mission, at least for a few decades to come. Resolving what the MFF really is would be one such area, as it is very large (see Section 3.5) and holds potential answers to the diversity of magma chemistry and styles of eruption on Mars (Zimbelman et al., 2020c). Either the elevation might be too high to allow parachutes to help decelerate a lander, or the rover-scale topography might be too rough for a safe landing. If mobility is not an option, one alternative mission concept might be to employ technologies from earlier decades and send one or more surface penetrators to this area (Greeley and Bunch, 1976). Such an approach was employed with the Mars Microprobe Mission in 1999 (Smrekar et al., 1999) as part of the Mars Polar Lander, but unfortunately no data at Mars were returned by this mission. Tests with terrestrial analog materials suggested that a penetration depth of at least 0.5 m could be achieved by these penetrators (Lorenz et al., 2000), and if the MFF is as unconsolidated as its morphology (Bradley et al., 2002), radar (Carter et al., 2009), and gravity/topography (Ojha and Lewis, 2018) characteristics suggest, then a penetrator might bury itself to at least a meter within the MFF. A diverse instrument payload for such a penetrator array would be possible based upon concepts for lunar penetrators (Ahrens et al., 2021), but at a minimum the determination of the MFF composition should be a high priority, as would the determination of the particle size and shape distribution, as well as the possible degree of welding of the particles.

5.5 Field observations for many locations

Prior to the landing of humans on Mars, the best future observations of volcanic processes on the planet could well be made using in situ data from rovers on the surface. This was advocated for the analysis of possible volcanic tephra deposits by Wilson and Head (2007), and it was subsequently demonstrated by Crumpler et al. (2015) that significant structural, petrographic, and stratigraphic information could be gained from rover observations. In addition, collecting age dates from a rover site (Farley et al., 2014; Anderson et al., 2020) would help refine knowledge of the thermal history of Mars by placing this volcanic activity into the broader chrono-stratigraphic history of Mars (Werner and Tanaka, 2011). The identification of specific lithologies would also be a great step forward, such as was the identification of pyroclastic activity from the probable bomb sag seen by the MER Spirit rover at Home Plate in Gusev Crater (Squyres et al., 2007).

Of course, the capabilities of the rover, in terms of the instrument package as well as the range over diverse terrains, would be of importance. Unfortunately, the most interesting examples of Martian volcanism are often located at high elevations on the planet, where the thin atmosphere creates challenges for the method used to land the rover. But as an example, the potential to resolve the style of activity at Tyrrhenus Mons from field observations can nevertheless be envisaged from analysis of phreatomagmatic cones such as Koko Crater, O'ahu, Hawai'i (Fig. 11). Although no layering has been observed in the walls of Tyrrhenus Mons (Williams et al., 2008), the probability that the volcano was formed by explosive eruptions suggests that layers should be visible in the field; detailed observations of a volcanic section would be instrumental in terms of identifying air-fall and/or pyroclastic flow sequences. Koko Crater is a “post-erosional cone” (Wentworth, 1926; Rottas and Houghton, 2012) that displays several features of the kind that may possibly be identified from a rover at Tyrrhenus Mons. In some places infilled gullies can be seen (Fig. 11c), implying deposition as well as erosion took place. Lithified climbing dunes within the section (Fig. 11d) may indicate relatively high-speed emplacement of materials via base-surges. Such base-surge deposits at tuff rings, including Koko Crater, have been described by Crowe and Fisher (1973) and Fisher (1977). Ejecta blocks may be visible and would be suggestive of vent material being incorporated into the eruption. If the blocks rest conformably within the pyroclastic layers (Fig. 11e) then the eruption was relatively dry. If, on the other hand, the layers were emplaced wet, then bomb sags might be found (Fig. 11f).

6. Conclusions

This review has identified several pressing questions about volcanism on Mars that warrant continuing analysis using existing data sets. These questions have often received less attention than others associated with the large shield volcanoes, but as such could become the focus for a future mission concept. For instance, the remarkable absence of identifiable exposed dikes around ancient, eroded volcanoes raises unanswered questions about the internal structure of these volcanoes and, by implication, the internal structure of younger (Amazonian) volcanoes. Resolving the absolute age of volcanic provinces will require the calibration of the impact crater cumulative size-frequency distribution curves. This would require returned samples from known locations, with samples from at least three areas such as Cerberus Fossae (Amazonian), the ridged plains (Hesperian) and Tyrrhenus Mons (Late Noachian/Early Hesperian). Multiple hypotheses exist for the formation of the MFF (Zimbelman et al., 2020c). A mission employing multiple penetrators to different parts of the

MFF might well resolve whether this material is an ignimbrite, ash fall, rafted pumice, or has a different (non-volcanic) origin.

Further analysis of terrestrial volcanoes may also shed new light on volcanic processes on Mars. Morphological comparisons between Martian calderas and their terrestrial equivalents have contributed to the quantification of the size scales and geologic histories of shallow magma bodies on Mars, but we still have difficulty in identifying the products of explosive volcanism commonly associated with such features on Earth (Mouginis-Mark et al., 2007). A high priority should be to understand the processes associated with the formation of volcanic landforms during the Noachian Period. Tyrrhenus and Hadriacus Montes are classic examples of this type of activity, and so continuing analysis of terrestrial volcanoes produced by the types of pyroclastic volcanism that appears likely on Mars, such as Koko Crater, Hawai'i, (Fig. 11), could guide future mission objectives and the analysis of existing high-resolution image and remote sensing data sets.

And what of the four questions outlined at the beginning of this review? They still remain, and alas the planned future missions to Mars do not place volcanic processes as the highest priority. Nevertheless, the successful landing of the Perseverance rover no doubt commits future Mars missions to the recovery of cached samples from Jezero Crater, which hopefully will provide sample ages from the Noachian into the Amazonian, depending on the true age of the delta deposits. The Perseverance rover carries the RIMFAX ground penetrating radar (GPR), which is providing the first glimpses into the upper 10 m of the floor of Jezero crater (Hamran et al., 2020). The Zhurong rover from the Chinese Tianwen-1 mission is currently on the surface, with results from its GPR still awaited, and the European Rosalind Franklin rover, which was due to launch in September 2022 but will most likely be delayed, would also carry a GPR that could provide additional subsurface information, but neither of the GPR-equipped rovers is being sent to a volcanic region. Thus, it seems that the next 30 years, comparable to the time since the 1992 review of Martian volcanism by Mouginis-Mark et al. (1992a), may not resolve all of the unknowns identified in this review, but we leave this as a challenge for the next generation of Martian volcanologists to address and answer.

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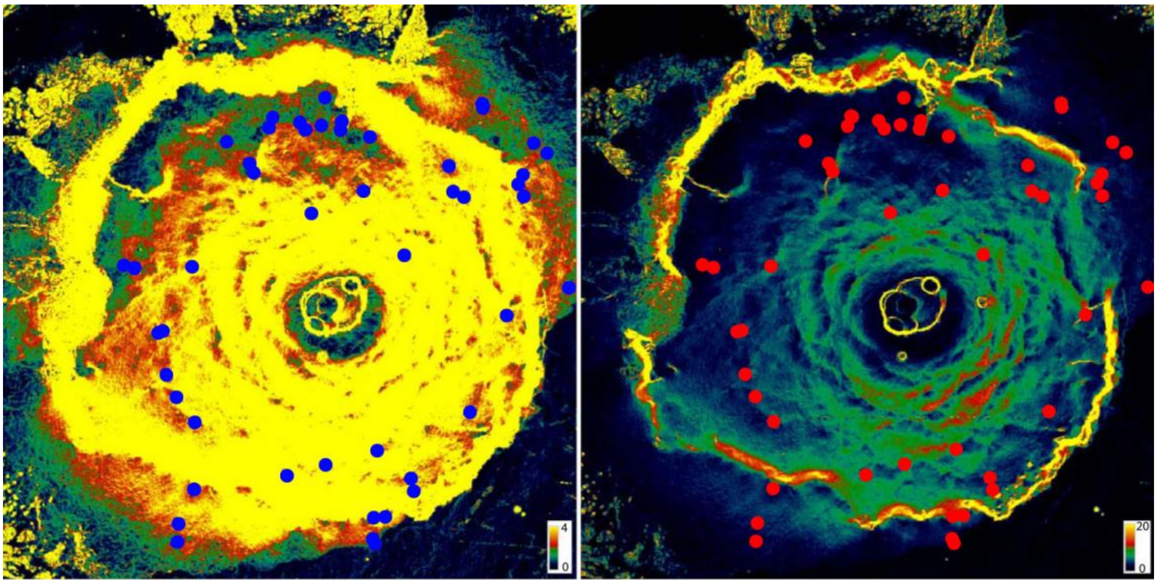

Fig. 1: Slope maps of Olympus Mons, with the vents hypothesized by Peters and Christensen (2017) identified (for clarity, these vents are shown in blue at left and red at right). Two different stretches of the slope map are shown: at left the range of slopes is 0^o to 4^o, at right the range is 0^o to 20^o. The volcano is ~600 km in diameter and rises ~22 km above the NW edge of the Tharsis Rise. These maps illustrate that there is little evidence for constructional volcanism at places down-slope from each of the proposed vents, implying that major volumes of lava were not erupted from vents on the flanks.

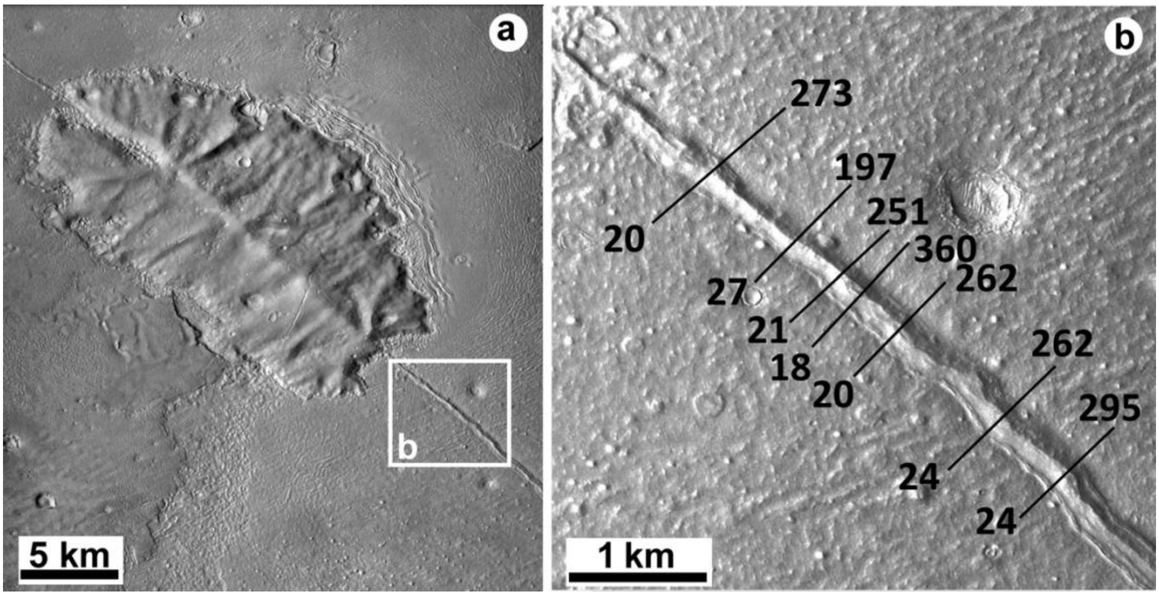

Fig. 2: Exposed dike to the SE of Galaxias Mons, near Hrad Vallis (35.2ºN, 142.6ºE). The dike appears to have originated from the vicinity of Elysium Mons, a considerable distance to the lower right of this image. (a) Location image, box shows the location of the higher resolution image at right. (b) Height of the exposed dike (left of lines) and maximum width (right of lines), both in meters. From Mouginis-Mark and Wilson (2016). Both images are parts of CTX frame G20_026093_2145.

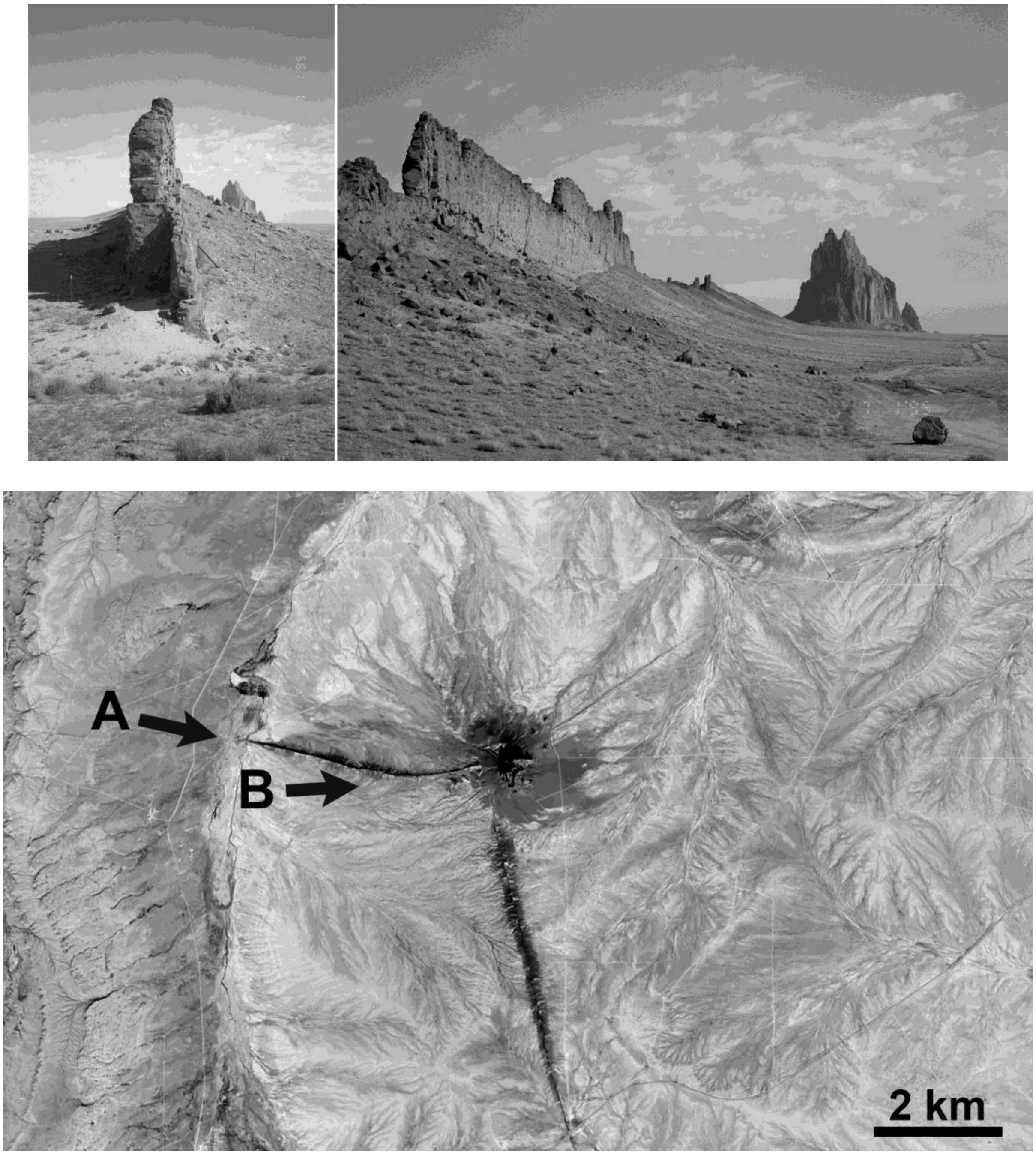

Fig. 3: What one might expect to see as an exposed dike on Mars. Where erosion has removed less resistant materials into which the dike was originally intruded, high-standing ridges of resistant intrusive material should be seen. This example is Tse Bit'a'i ("Ship Rock") in New Mexico, which formed ~27 million years ago. In places, the dikes can rise as much as 50 m above the surrounding plain but are typically <2 m wide. Images taken in 1995 before access restrictions were introduced by the Navajo Nation (in 2016). Bottom image is a Google Earth view of the three dikes. Ground photos taken
approximately from point "A" and "B".

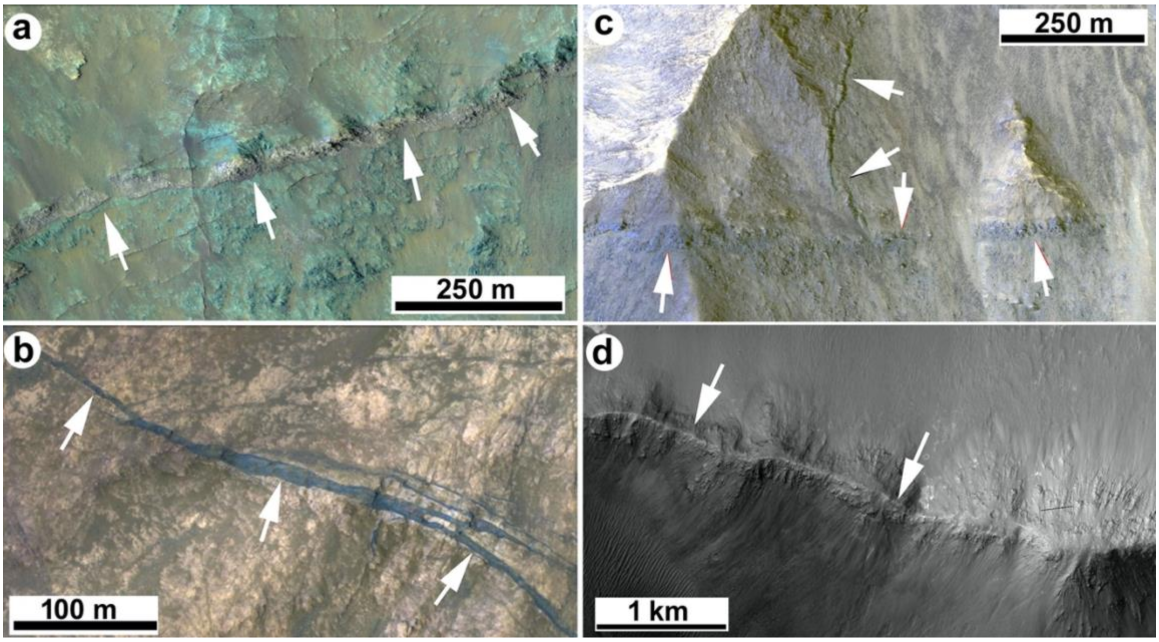

Fig. 4: Probable dikes (identified with arrows) in the canyon walls at the eastern end of Valles Marineris. a) HiRISE image ESP_013903_1650_Color (15.0°S, 303.4°E). b) HiRISE image ESP_040461_1670_Color (12.7°S, 298.7°E). c) HiRISE ESP_013191_1660_Color (14.0°S, 304.9°E). d) Linear ridge (arrowed) on the floor of the canyon, which could be the surface expression of a dike. The topography is not clear in this image, but the ridge is higher than both the area toward the top and bottom of the image. HiRISE image ESP_036518_1735 (6.2°S, 290.7°E).

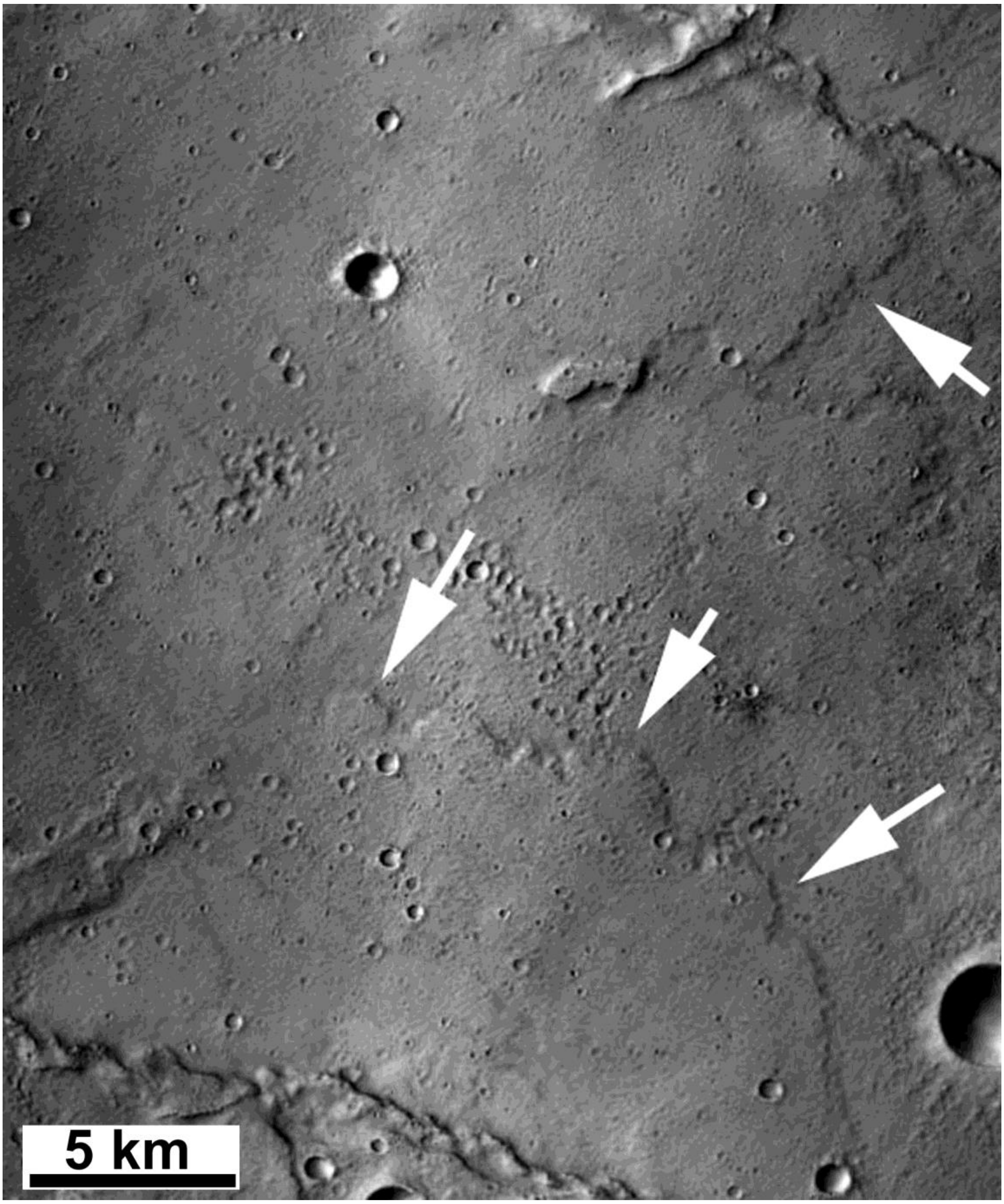

Fig. 5: Flow lobes (arrowed) to the east of the volcano Tyrrhenus Mons. There are very few examples of lava flows within the ridged plains of Mars, which is remarkable considering that the extensive ridged plains are considered to be volcanic in origin. Image centered at 20.65ºS, 108.8ºE. CTX image K05_055355_1594.

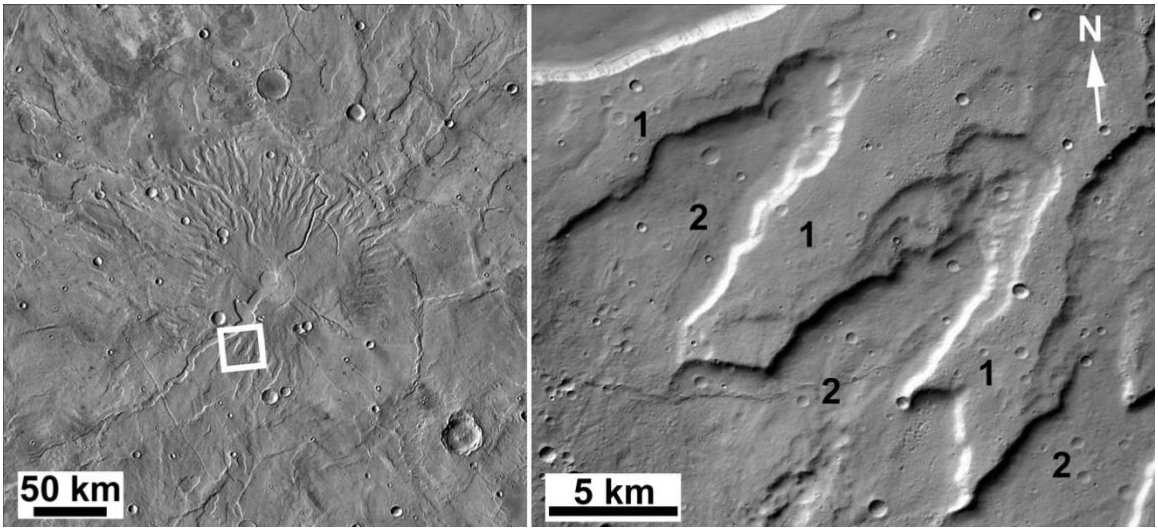

Fig.6: Tyrrhenus Mons. Left, THEMIS VIS mosaic, box shows location of image at right. The main edifice is ~215 x 350 km in size, and the summit rises ~1.5 km above the surrounding plain. At right, at least two layers ("1" and "2") exist on the upper southern flanks of the volcano and could represent different epochs of activity. However, no partially exhumed craters are seen at the boundary of the upper unit, suggesting that there was no protracted repose period between the two phases of eruption that produced these layers. CTX image G17_025000_1583.

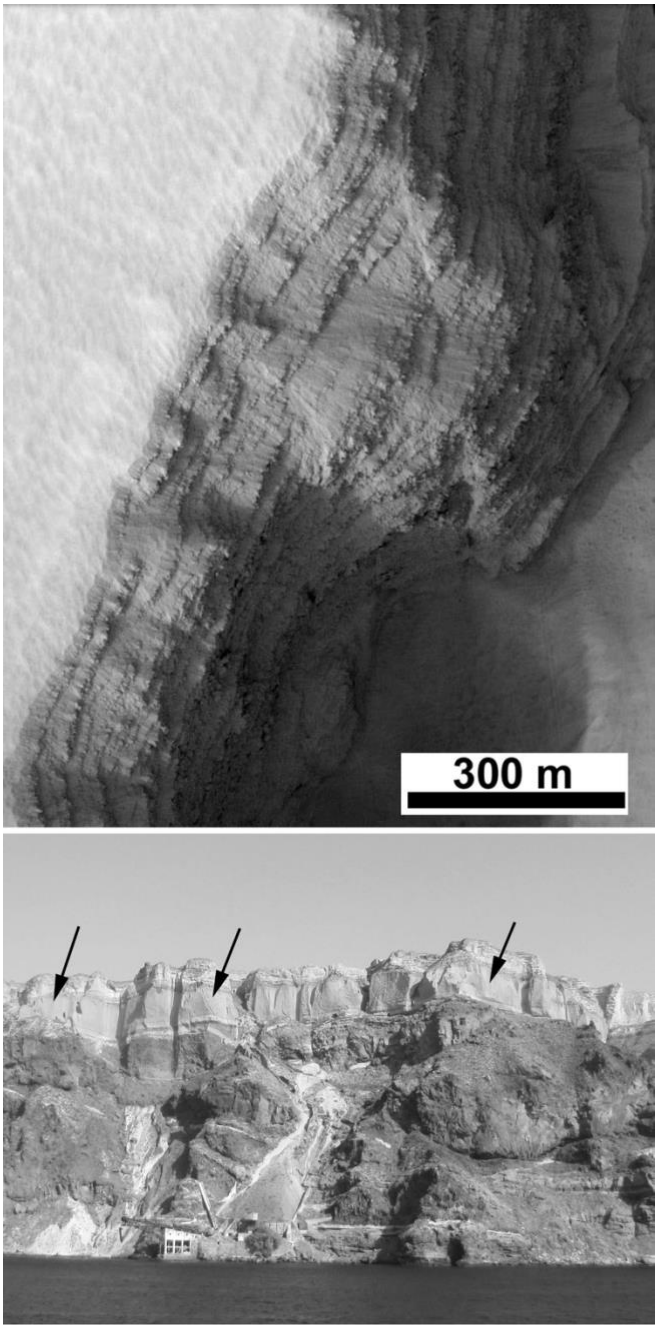

Fig. 7: Top: Numerous lava flows can be seen in section within a graben on the western flank of Arsia Mons, at 8.57ºS, 236.37ºE. The depth of this graben is ~885 m. Notice that there is no evidence for thick ash layers within this section, arguing against the idea that there have been extensive explosive eruptions at this volcano. HiRISE image PSP_004412_1715. Bottom: Something comparable to this scene of the caldera wall of Santorini, Greece (viewed from just off-shore), might be expected if Arsia Mons had produced copious ash. The thick light-colored unit near the rim (arrowed) is ~10 m thick, produced during the 1610 B.C. "Minoan eruption". The excavator at the foot of the cliff also provides the approximate scale.

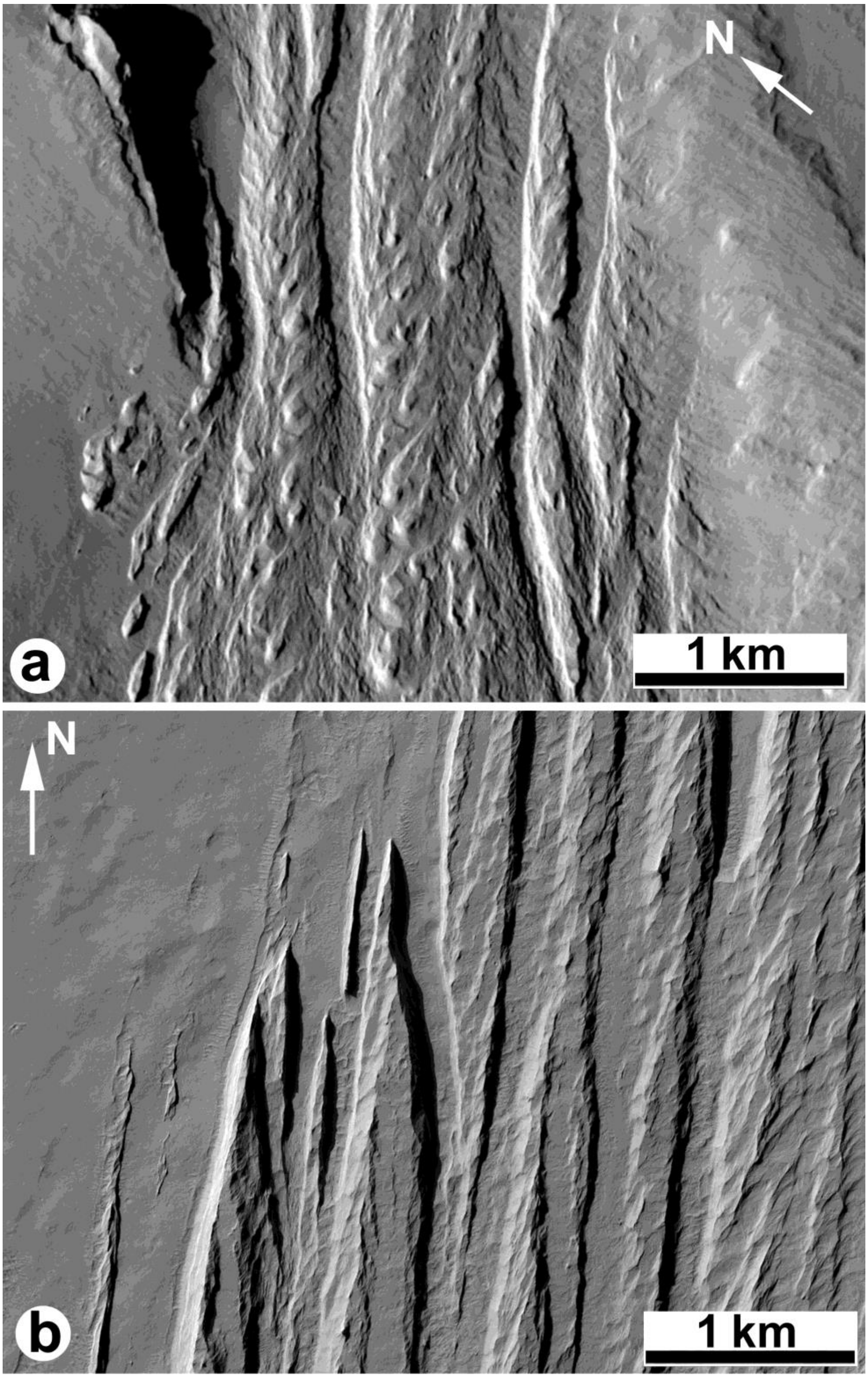

Fig. 8: Is there Medusae Fossae Formation material within Gale Crater? (a) Part of the upper mound etched unit 2, 4.8ºS, 137.0ºE. as mapped by Thomson et al. (2014). CTX image B01_009927_1752. (b) Contact between Apollinaris Mons (at left) and the MFF (yardangs at right), at 10.2ºS, 176.4ºE. HiRISE image PSP_009464_1695.

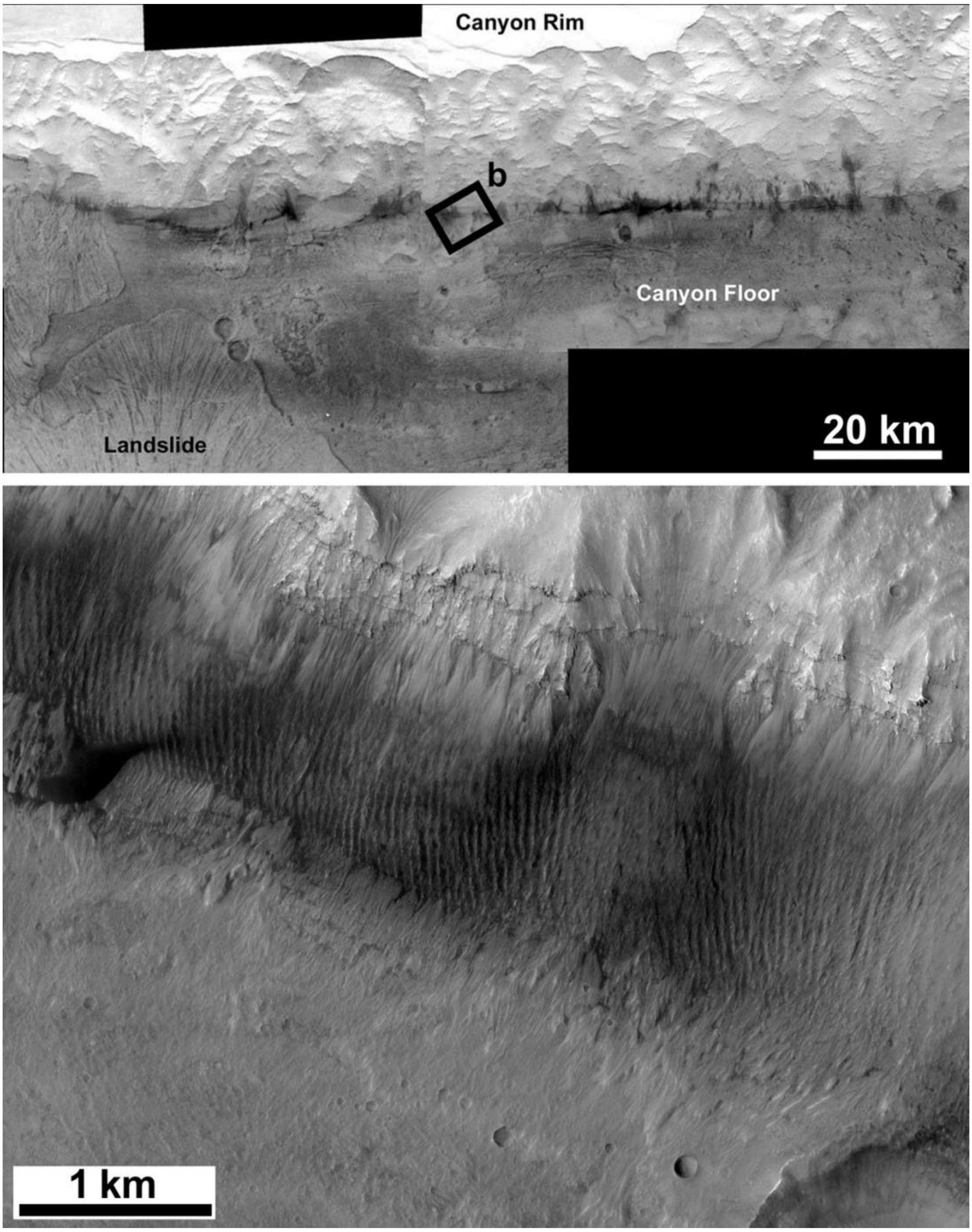

Fig. 9: Top: A line of dark patches at the base of the north wall of Coprates Chasma, as identified by Lucchitta (1990). These were interpreted from Viking Orbiter images to be pyroclastic materials, but subsequent inspection of HiRISE images (bottom image) reveal that they are accumulations of eolian materials. At this point, the canyon walls are ~7 km high. Viking Orbiter images 80A02, 80A01, and 81A04. Bottom: HiRISE image ESP_022369_1685, centered at 11.4ºS, 293.5ºE. Note lack of clear indication that this material originates from a specific continuous layer within the Canyon walls. The wall rises from a low point at lower left to the high point at top right.

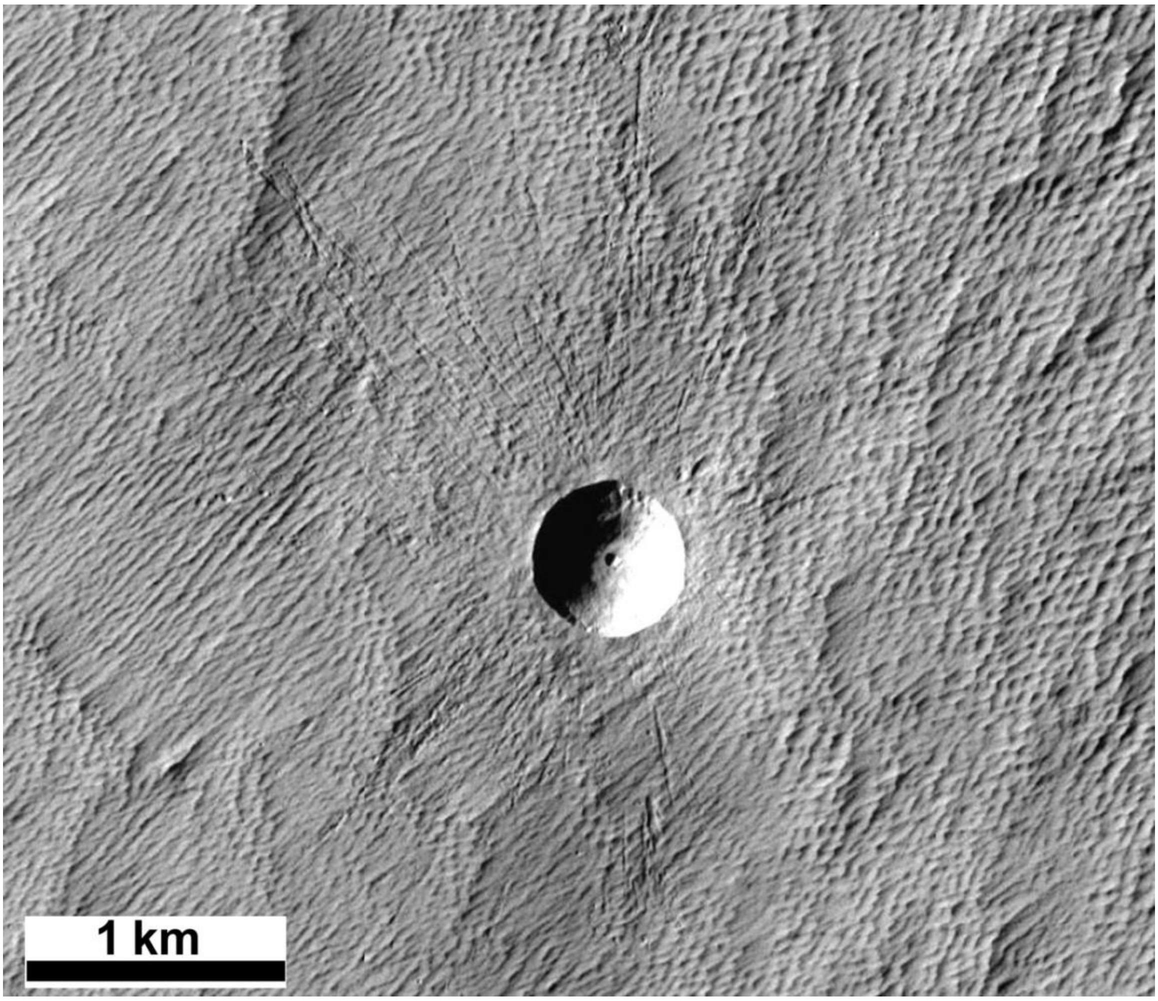

Fig. 10: Relatively fresh 600 m diameter impact crater within the Medusae Fossae Formation, located at 2.32°S, 203.96°E. Note the abundant strings of secondary craters to the north, south and west of the crater, which must indicate the relatively youthful nature of this crater because otherwise it is expected that they would be rapidly removed by eolian processes. If it were possible to collect CRISM data over small craters soon after they formed, it is possible that the composition of the MFF could be constrained. CTX image N02_063132_1777.

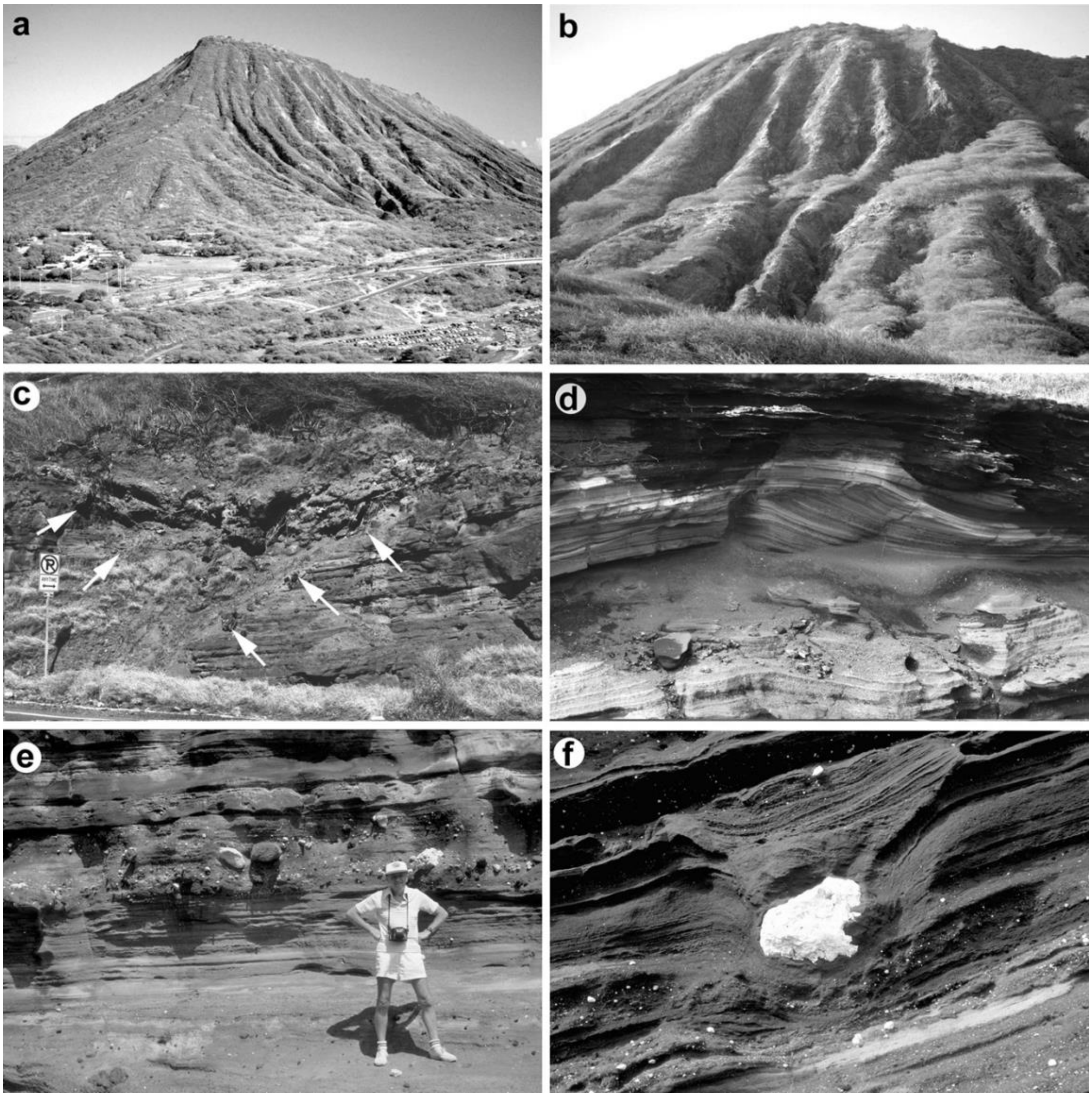

Fig. 11: Various views of Koko Crater, O‘ahu, Hawai‘i, which could serve as an analog to the objectives of a rover exploring the flanks of Tyrrhenus Mons. (a) General view of cone, which is ~200 m high. b) Rainfall has carved deep gullies on the outer slopes, which implies that surface materials became sufficiently impermeable that surface runoff is more effective than water seeping into a friable, porous deposit. (c). Infilled gully cuts through layers of ash (border denoted by arrows), implying deposition as well as erosion took place. d) Road exposure showing numerous ash layers produced by turbidity currents speeding down the slopes. Dunes and anti-dunes are visible in this section, which is ~2 m high. e) In places, ejecta landed in dry ash deposits and made no indentation in the strata, in contrast to the bomb sag in (f), where ejecta landed in very wet sediment. Cliff Mark provides scale in (e). Block is ~30 cm across.