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An overview of explosive volcanism on Mars

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Abstract

Decades of space exploration reveal that Mars has been reshaped by volcanism throughout its history. The range of observed volcanic landforms shows that effusive and explosive eruptions have occurred, albeit unevenly in time and space. Evidence for explosive volcanism—characterized as eruptions in which magma is disrupted by the expansion of gases dissolved in magma or by an interaction of magma with external volatiles—is less common than evidence for effusive activity. Nonetheless, some indications of explosive volcanism have been identified at various sites. For instance, old, rimless depressions, termed paterae, on the summits of broad topographic rises with very

gentle flanks located mainly around the Hellas impact basin. Also various fields of kilometre-sized cones interpreted as scoria cones, tuff rings and tuff cones, and extensive clusters of sub-kilometre-sized pitted cones in the northern lowlands proposed to be rootless cones, are thought to be constructional features caused by accumulation of volcanic fragments. Finally layered deposits widely spread in equatorial areas (e.g., the Medusae Fossae Formation), and layered stacks of ash and a putative volcanic bomb observed by rover, point to a protracted history of explosive volcanism on Mars. Yet some of these interpretations remain the matter of scientific debate. The discovery of explosive volcanism on Mars has triggered an interest in theoretical aspects of such volcanism under gravitational and atmospheric conditions different from those on Earth. These studies indicate that explosive eruptions on Mars would behave differently from those on Earth. This is because a lower atmospheric pressure and gravity can affect all stages of the eruption including the ascent of magma, the process of degassing and magma fragmentation, the transport and deposition of the pyroclasts, and also in some cases the formation of explosive volcanoes themselves. On Earth, explosive eruptions are responsible for the formation of most volcanoes on land, and so a relatively sparse occurrence of explosive volcanism on Mars is surprising, especially considering the martian environmental conditions as well as wide occurrence of external volatiles on Mars. This is because the lower atmospheric pressure than on Earth ought to favour magma fragmentation and hence the formation of pyroclasts and associated explosive volcanic edifices, even if lower volumes of dissolved gases were present in martian magma than is usual on Earth. The relative dearth of explosive activity on Mars therefore represents a gap in our understanding of martian volcanism, suggesting that there may be considerable compositional differences between Mars and Earth or that evidence of explosive volcanism on Mars manifests differently to on Earth. Understanding these differences is important, as explosive volcanism provides insight about the planet's composition and plays a crucial role in the evolution of a planet's atmosphere by the release of magmatic gases, which have the ability to affect geological and even biological processes operating on the surface of the planet. In this paper, we present an overview of explosive volcanism on Mars—from both observational and theoretical perspectives—and discuss the implications of explosive eruptions for the evolution of the Red Planet.

1. Introduction

The first evidence for extensive volcanic products on Mars comes from images by the Mariner 9 mission. Mariner 9 as well as the subsequent Viking missions, revealed that the martian landscape is dominated by mountains dozens of kilometres high which have craters at their summits – later interpreted as large shield volcanoes with multiple calderas – and by lava flows extending over hundreds of kilometres across the surface (e. g., McCauley et al., 1972; Carr, 1973; Carr et al., 1977; Greeley and Spudis, 1981; Tanaka et al., 1988; Hodges and Moore, 1994; Greeley et al., 2000; Tanaka et al., 2014a; Zimbelman et al., 2015). Investigations were initially focused on the large volcanic edifices, e.g., Olympus, Arsia, Pavonis, Ascraeus, and Elysium Montes, which are associated with extensive volcanic plains, together forming two prominent volcanic provinces – Tharsis and Elysium (Grott et al., 2013 and references therein). The morphological, morphometric and spectroscopic data revealed that the Tharsis, Elysium, as well as Syrtis Major complexes represent prominent large-scale volcanic provinces formed by effusions of low-viscosity basaltic lavas (e.g., Mouginis-Mark et al., 1992; Zimbelman, 2000; Hauber et al., 2011).

Besides these large effusive shield volcanoes and associated extensive lava plains, several broad edifices have been discovered in the southern hemisphere of Mars (Scott and Carr, 1978; Peterson, 1978; Greeley and Spudis, 1981; Greeley and Crown, 1990; Crown and Greeley, 1993; Williams et al., 2007, 2008). These edifices, generally called *paterae*, are different in shape from martian large shield volcanoes. Their flanks are very gentle ($<1^\circ$) and heavily eroded, suggesting that they were not formed by solid volcanic rocks, but instead by friable materials. The existence of these *paterae* was the first evidence that explosive volcanism may have occurred on Mars.

After several decades of Mars exploration the body of evidence for explosive eruptions still remains significantly smaller than that for effusive ones (e.g., Tanaka, 2000; Platz et al., 2015 and references therein). This observation is of particular interest as a) the martian environment should favour magma fragmentation caused by magma degassing (Wilson and Head, 1994) and b) evidence for the presence of external volatiles on the surface of Mars, which can give rise to hydrovolcanism, are relatively common (e.g., Baker, 2001; Feldman et al., 2004; Smith et al., 2009; Byrne et al., 2009;

Vincendon et al., 2010). On Earth, to trigger an explosive eruption a magma has usually to interact with external volatiles or to have a volatile content higher than 2 wt% mainly consisting of CO₂ and H₂O (e.g., Carroll and Holloway, 1994; Wallace et al., 2015). On Mars, however, only ~0.03 wt% of volatiles are required to result in such eruptions (Wilson and Head, 1981; 1994). This is because the relatively low surface atmospheric pressure, which averages around ~7 mbars (~0.7% Earth's surface pressure) should cause a rapid expansion of any gases dissolved within the ascending magma (e.g., Wilson and Head, 1981; 1994; Grott et al., 2013). This would result in explosive volcanism for smaller volumes of dissolved gases than is common on Earth, and it has been theorized that all steady eruptions under present-day martian conditions should be explosive (Wilson, 2009).

On Earth, volcanic structures formed by explosive volcanism are common and widespread. Examples include scoria or cinder cones, kilometre-sized explosive edifices formed by accumulation of pyroclasts resulting from low-energy volcanic eruptions of Strombolian type (e.g., Wood, 1979a,b; Vespermann and Schmincke, 2000; see also Kerezić and Neméth, 2012 for review) together with hydrovolcanic edifices in the form of tuff rings, tuff cones and maars (e.g., Sheridan and Wohletz, 1983; Wohletz and Sheridan, 1983; Vespermann and Schmincke, 2000). They represent the overwhelming majority of terrestrial volcanic edifices situated onshore (e.g., Wood, 1979a; 1980; Sheridan and Wohletz, 1983; Vespermann and Schmincke, 2000). On Mars, however, as we show below, such landforms are surprisingly sparse.

Despite the progress in our understanding of the evolution of Mars in the last several decades, the reason for this sparsity still remains unclear. The aim of this review is therefore to summarize the available knowledge about explosive volcanism on Mars and provide pathways to resolving this outstanding issue. Our aim is to summarise the state of knowledge at the time of writing and cite references to support key points, rather than provide a complete historical documentation of the literature, from discovery to present-day understanding. Interested readers can use these key citations as gateways into the deeper literature. First, a general background on explosive volcanism is provided (Section 2) followed by an overview of localities where the presence of surface manifestations of explosive volcanism has been proposed based on remote (Section 3) or in-situ investigations (Section 4). Then, the effect of the martian environment on the course of volcanic eruptions, pyroclast

transport, and the way this would change the expression of resulting surficial features is discussed (Section 5). Finally, we address the way explosive volcanism might have affected the evolution of Mars (Section 6). However, it has to be noted that this overview is limited to explosive eruptions resulting from silicate melts, and does not discuss those associated with sedimentary volcanism and cryovolcanism, which have been proposed to operate on Mars as well (e.g., Bargery and Wilson, 2010; Boyce et al., 2012) and which might also lead to explosive eruptions (Brož et al., 2019; 2020). These processes are beyond the scope of this work.

In discussing how volcanism has evolved, we use the geologic timescale for Mars, which is divided up into three periods of decreasing age, Noachian (4.1-3.7 Ga), Hesperian (3.7-3.1 Ga), and Amazonian (3.1 Ga to present) (Tanaka, 1986).

2. Why do volcanoes erupt explosively?

Whether a volcanic eruption is explosive or effusive depends on whether the ascending magma breaks up into individual pieces or not (e.g. Gonnermann and Manga, 2013). In principle, there are two ways to fragment magma. One is associated with volcanic gases dissolved within the magma, henceforth referred to as a “dry” process, and the other one depends on the availability of external volatiles that magma can interact with during its ascent, henceforth referred to as a “wet” process (Cashman et al., 1999). The dry process occurs because magma can evolve bubbles of compressible volcanic gases, on Earth predominantly H_2O and CO_2 , which can rapidly increase in volume when the magma is decompressed (e.g. as it ascends). Depending on the viscosity of the magma, rapidly expanding gases can either fragment the magma, or in high viscosity melt cause a build-up of gas pressure within the magma. A build-up of pressure can deform the magma viscoelastically and finally cause glassy-fracture resulting in violent explosions and a higher degree of magma fragmentation (Gonnermann, 2015). On the other hand, the “wet” process is independent of the amount of dissolved volcanic gasses as it relies on an external source of volatiles. On Earth the most important one is water, therefore this type of eruption is often termed as hydromagmatic. Water can be in the form of groundwater, ground ice, or a surficial body of liquid water (e.g., Sheridan and Wohletz, 1983; Cashman et al., 1999). However, on other bodies within the Solar System different volatiles, like CO_2

or sulphur, may contribute (e.g. Davies, 2007). On Earth, the basic principle of this interaction is the rapid heat transfer from the magma to the water leading to vaporization of the water, steam expansion and pressure build-up, and subsequent fragmentation of the magma and explosive eruption processes (Basaltic Volcanism Study Project, 1981, p. 729). These types of eruptions are characterized by the production of steam and fine-grained fragments of magma and surrounding rocks ejected from the central vent in a series of energetic eruptive pulses (Sheridan and Wohletz, 1983).

Whether the magma fragmentation is caused by dry or wet processes has a direct impact on the final morphologies of resulting volcanic structures. This is because volcanic gases or external volatiles, in sufficient volumes, are able to transport exploded rock fragments (pyroclasts) from the vent to various distances according to their sizes either ballistically and/or by turbulent jets (e.g., Riedel et al., 2003; Wilson and Head, 1994). Based on the degree of fragmentation of pyroclasts and their dispersal different types of eruptions can be devised. The examples of low energetic eruptions include so-called Hawaiian or Strombolian types during which the relatively large-sized erupted material, which is almost completely magmatic in origin, is predominantly transported ballistically. The pyroclasts mainly within the range of millimetre- to centimetre-sized fragments fall close to the vent and hence form kilometre-sized positive topographic features, like scoria cones or spatter cones (e.g., Riedel et al., 2003; Kereszturi and Nem  th, 2012 and references therein).

On the other hand, the pressure build-up of exsolved gases within a more viscous magma can trigger very energetic and violent eruptions of sub-Plinian, Plinian, or Ultra-Plinian types forming finer pyroclasts, which can be entrained in buoyant convective plumes of hot gas high into the atmosphere. At moderate to high discharge rates, on Earth corresponding to sub-Plinian and Plinian eruptions, convective plumes can reach a height of 10-40 km, hence penetrating the stratosphere, or at very high discharges associated with Ultra-Plinian eruptions can be up to 50 km high (e.g., Woods, 2013 and references therein). The erupted volcanic gases and pyroclasts, which can emerge from the vent at speeds up to 600 m/s^{-1} with temperatures of 800-1200 °C, move through the atmosphere as a highly turbulent flow into which surrounding air is entrained. This mixing causes rapid heat transfer from the hot pyroclasts to the air causing a dramatic decrease of the convective plume density and promoting its vertical propagation (Woods, 2013). Depending on the size of the pyroclasts the material

can fall out from the plume close to the vent or be transported far away from the vent by the wind (e.g., Parfitt and Wilson, 2008 and references therein). The fallout from these plumes can give rise to unconsolidated airfall pyroclastic deposits, which can be spread over hundreds or thousands square kilometres (e.g., Kerber et al., 2012).

When the convective plume becomes denser than the surrounding air, it becomes unstable and collapses. Such collapses lead to the formation of pyroclastic density currents (see e.g. Dufek [2016] for an overview), which are hot clouds composed of a mixture of volcanic gases, pyroclasts and surrounding air. They can move over the surface at high speeds and travel tens of kilometres from the eruption site. Over the course of their path, these currents can form variously thick ignimbrite deposits (e.g., Criswell and Elston, 1981; 1982) that can be distributed over thousands of square kilometres. As the temperature of these currents can be in the range of hundreds of degrees Celsius, the deposited pyroclasts can become welded into ignimbrites. On Earth, the largest known ignimbrite-forming eruptions are associated with the formation of the caldera and hence with the presence of large magma chambers containing unusually gas-rich magma (e.g., Cashman and Giordano, 2014 and references therein). However, it has to be noted that the presence of calderas on terrestrial bodies can not be taken as unequivocal evidence of explosive activity. This is because calderas can also be formed in an effusive setting, by the extensive withdrawal of magma from shallow magma chambers causing subsequent collapse of the summit area of the volcano (see Cole et al., 2005 and Cashman and Giordano, 2014 for review).

Pyroclastic density currents can also result from hydrovolcanism, even when relatively small amount of magma is available. This is because wet processes are generally more energetic than dry processes. This is because the conversion of water from liquid/solid to steam causes a more than 2,000-fold increase in its volume (Parfitt and Wilson, 2008). This causes both magma and surrounding rocks to fragment and be transported from the vent ballistically or by convective plumes in the atmosphere. Depending on the volume of ejected material to the close vicinity of the vent a positive topographic feature can be formed, namely tuff rings or tuff cones, or if the amount of material is insufficient a depression called a maar would be formed (e.g., Sheridan and Wohletz, 1983). Further from the vent pyroclastic deposits can be formed as well.

Another type of water-lava interaction can lead to the formation of rootless cones, formerly termed as pseudocraters (Allen, 1979; Thorarinsson, 1953; Thordarson and Self, 1993; Hamilton et al., 2011). These cones are known as rootless due to lack of a direct feeder from the location of the eruption (Hamilton et al., 2011; Fagents and Thordarson, 2007). They form when moving lava interacts with underlying external water sources forming water vapour capable of rapid expansion (Fagents and Thordarson, 2007; Thorarinsson, 1953). The source of water can be groundwater, ground ice (Head et al., 2005), or a surficial body of water (Wohletz and Sheridan, 1983). The resulting near-circular edifices commonly form in groups containing hundreds or thousands of edifices from a few tens to a few hundreds of meters in diameter. They often have a well-developed central crater with a partially or fully preserved rim (Hamilton et al., 2011; Fagents and Thordarson, 2007).

3. Expressions of explosive volcanism on the surface of Mars

Volcanic activity has played an important role in the evolution of Mars, yet has been unevenly distributed in time and space (e.g., Tanaka, 1986; Hodges and Moore, 1994 and references therein). While evidence for volcanic activity has been observed almost globally (see Tanaka et al., 2014 for details) suggesting that volcanism was once widespread, the size-frequency distribution of impact craters superposing these volcanic edifices varies significantly over different areas suggesting that volcanism was not evenly distributed in time (e.g., Plescia and Saunders, 1979; Tanaka, 1986; Crown et al., 1992, 2005, 2007; McEwen et al., 1999; Leonard and Tanaka, 2001; Mest and Crown, 2001; Werner, 2005; Crown and Greeley, 2007; Werner, 2009; Robbins et al., 2011). Very young surfaces are observed only in the two large volcanic provinces of Tharsis and Elysium, which suggests that as time went on, volcanic activity localized into these two provinces (e.g., Tanaka, 1986; Werner, 2009; Robbins et al., 2011; Grott et al., 2013 and reference therein). These provinces contain numerous volcanic vents from which countless young lava flows emerged (e.g. Bleacher et al., 2007a,b; Hauber et al., 2009; Vaucher et al., 2009a,b) suggesting that effusive volcanism was a dominant style of volcanic activity, particularly later in Mars' history (e.g., Carr et al., 1977; Greeley and Spudis, 1981; Vaucher et al., 2009a; Hauber et al., 2011). On the other hand, various sites on Mars, mainly the southern highlands, contain volcanic edifices which may have resulted from explosive activity (e.g.,

Peterson, 1978; Greeley and Spudis, 1981; Greeley and Crown, 1990; Crown and Greeley, 1993; Gregg and Farley, 2006; Williams et al., 2007, 2008). While effusive volcanic structures within Tharsis and Elysium seem to be relatively pristine and hence young (e.g., Vaucher et al., 2009a; Hauber et al., 2011; Robbins et al., 2011), this is not the case for many volcanic structures within Mars' southern highlands, which are, as described below, heavily eroded and covered by large quantities of impact craters pointing to their relatively old age. This indicates that explosive volcanism was an important process in the early stages of Mars' history (e.g., Plescia and Saunders, 1979; Crown et al., 1992, 2005, 2007; Leonard and Tanaka, 2001; Mest and Crown, 2001; Werner, 2005; Crown and Greeley, 2007). This suggests a temporal transition between older, pyroclastic-dominated highland paterae and more recent effusive central-vent volcanoes of Tharsis and Elysium (e.g., Tanaka, 1986; Greeley and Crown, 1990 and references therein). Below, this section contains a summary of putative explosive edifices of various origins and sizes which have been reported on the surface of Mars (Section 3.1), namely paterae, tholi, scoria cones, snatter cones, and hydrovolcanic features, in decreasing order of the amount of ejected material required for their formation, as well as putative pyroclastic deposits (Section 3.2). They show that explosive edifices are a broad family of features with various ages and a global distribution (Fig. 1).

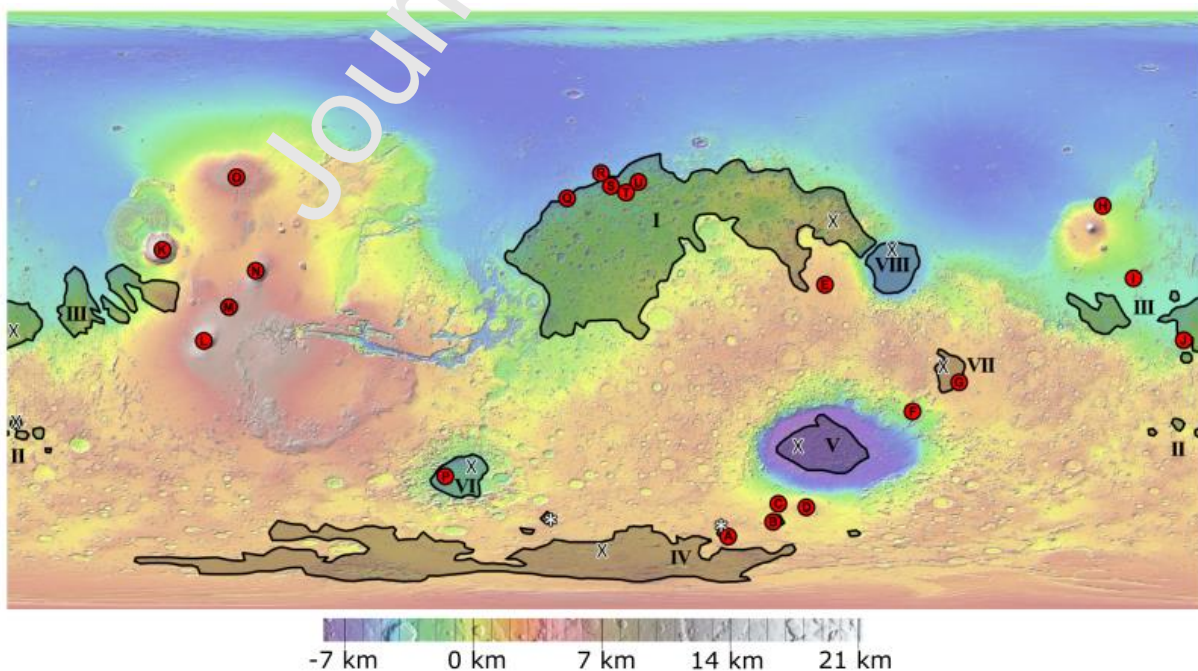


Figure 1: Global MOLA topographic map showing deposits larger than 10^5 km^2 of potential

pyroclastic origin (black outlines; see Table 1 in Section 3.2 for Roman numerals). Black crosses and white asterisks indicate the locations shown in Fig. 8 and 9, respectively. Red dots mark potential centers of explosive volcanism as mentioned in the references listed in Table 1. Letters are: A) Pityusa Patera; B) Malea Patera; C) Peneus Patera; D) Amphitrites Patera; E) Nili/Meroe Paterae; F) Hadriaca Patera; G) Tyrrhena Patera; H) Hecates Tholus; I) Cerberus Fossae; J) Apollinaris Patera; K) Olympus Mons; L) Arsia Mons; M) Pavonis Mons; N) Ascræus Mons; O) Alba Patera; P) Argyre Mons; Q) Eden Patera; R) Oxus Patera; S) Ismenia Patera; T) Siloe Patera; U) Euphrates Patera. Freely adapted and updated from Tanaka (2000) and Kerber et al. (2012).

3.1. Edifice-building by pyroclastic deposits

On Earth, volcanic edifices can be formed by effusive or explosive volcanic activity, or by combination of these processes leading to the formation of a wide variety of edifices. This section discusses martian features for which an explosive origin resulting from magma degassing and/or interactions between magma and external volatiles has been proposed. First, the largest landforms, paterae, are discussed, followed by the descriptions of kilometre-high mountains termed tholi, hundred meters high scoria cones, metre-sized spatter cones, and finally a wide group of variously-shaped and -sized hydrovolcanic features believed to result from water-magma interactions.

3.1.1. Paterae

The most spectacular evidence for explosive volcanism on Mars is the presence of two ancient edifices, Hadriacus (Fig. 2a) and Tyrrhenus Montes, which host large paterae, Hadriaca and Tyrrhena Paterae, on their summits and are situated northeast of the Hellas impact basin (Peterson, 1978; Plescia and Saunders, 1979; Greeley and Spudis, 1981; Greeley and Crown, 1990; Crown and Greeley, 1993; Gregg and Farley, 2006; Williams et al., 2007, 2008). Both are hundreds of kilometres wide and have been interpreted as volcanic edifices due to the presence of shallow, low-relief, central summit, caldera-like paterae and heavily eroded flanks with very gentle slopes ($<1^\circ$) (Plescia, 2004) on which numerous slope-parallel channel-like valleys can be found (Greeley and Crown, 1990; Gulick and Baker, 1990; Crown and Greeley, 1993; Crown et al., 2005). The heavy erosion of the flanks implies

that they consist of friable material, which suggests that the eruption style differed from the style of volcanic activity that formed the large, mainly effusive volcanic plains like Tharsis, Elysium, Hesperia Planum, or Malea Planum (Greeley et al., 2000; Williams et al., 2007, 2008). Their flanks show signs of modifications by fluvial, aeolian, and/or other processes. Their flanks also lack primary lava flow features, which motivated an interpretation as accumulations of pyroclastic material emplaced by airfall and/or gravity-driven pyroclastic flows. Based on the shape of Tyrrhenus Mons the airfall scenario was rejected as the volcano does not have an asymmetrical shape, which should arise from such deposition because of the prevailing direction of winds within this area (Greeley and Crown, 1990). In addition, to form friable airfall deposits at distances up to several hundreds of kilometres from the summit area of Tyrrhena Mons, the convective plume would have to rise to an altitude of 250 kilometres. However, as shown by numerical modelling, such a plume height would be unlikely in a past, possibly warmer Mars atmosphere (for details see Crown and Greeley, 1993 and references therein). Therefore, the formation scenario associated with the emplacement of pyroclastic density currents, namely with pyroclastic flows, was favoured. It was proposed that these flows could be produced by large ancient explosive eruptions that were triggered by water-magma interactions (Plescia and Saunders, 1979; Greeley and Spudis, 1981; Greeley and Crown, 1990; Crown and Greeley, 1993; Gregg and Farley, 2006; Williams et al., 2007, 2008, 2009). Tyrrhenus Mons is also associated with extensive friable deposits, which have been interpreted as pyroclasts (see Section 3.2.7 for details).

It was proposed that explosive activity associated with the formation of these two paterae might have been caused by the interaction of ascending magma with a substantial groundwater zone (Carr, 1979) leading to the generation of a vast amount of pyroclasts (Greeley and Crown, 1990; Crown and Greeley, 1993). This explosive phase might then have been followed by partial collapses of the central parts of these volcanoes due to the magma withdrawals that gave rise to the paterae, i.e., central, complex, caldera-like depressions on the summits of the Montes (Greeley and Spudis, 1981). However, it should be noted that the origin of the water (magmatic or meteoric) for water-magma interactions is a subject of debate (Hauber et al., 2009). Therefore, it remains unclear whether magma

degassing or magma-water interactions caused the explosive eruptions associated with these volcanoes.

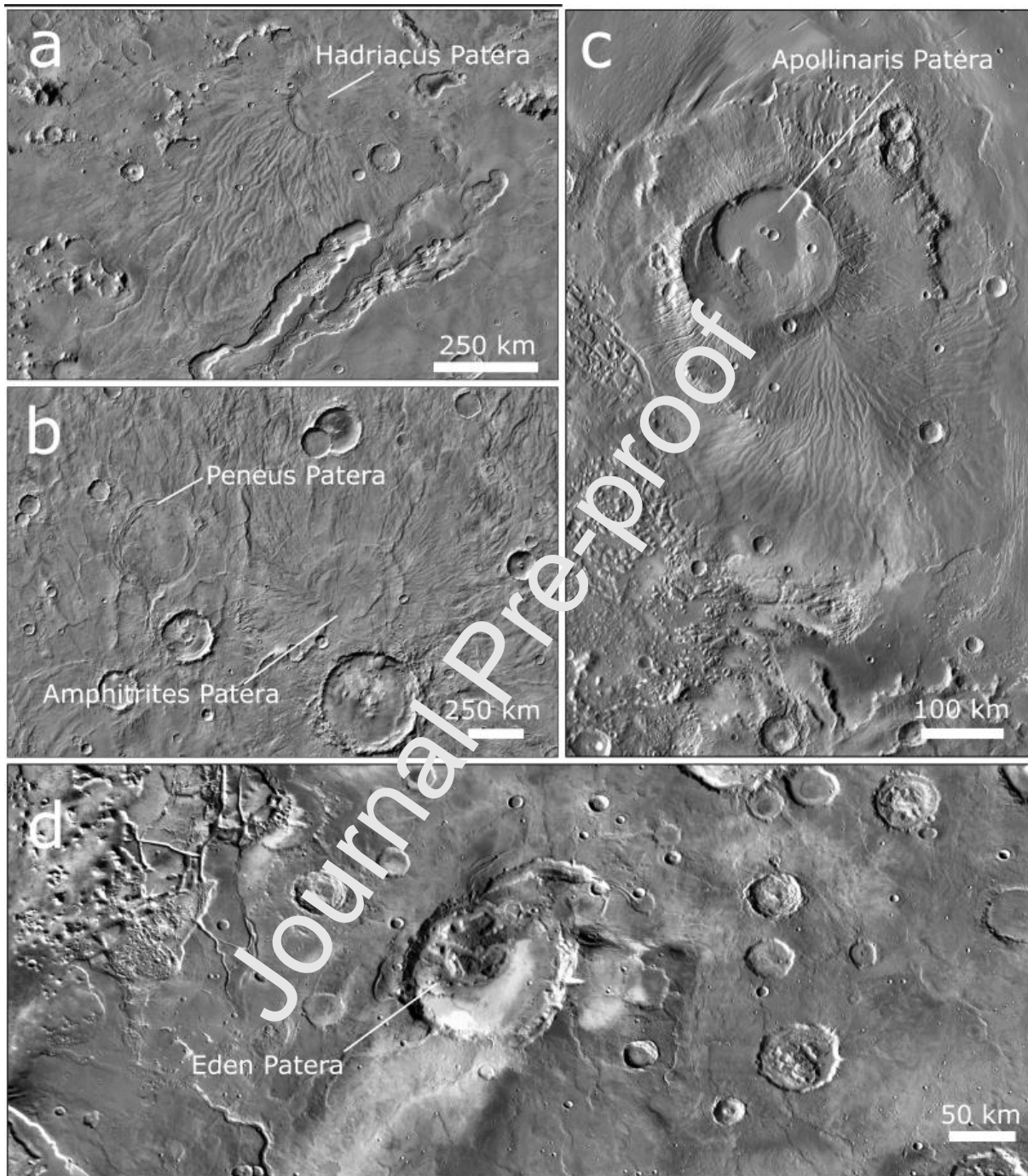


Figure 2: Image showing examples of paterae on Mars. Panel (a) shows Hadriaca Patera on the summit of Hadriacus Mons (centered 31.5°S , 91.8°E), (b) Peneus Patera and Amphitrites Patera (centred 58.7°S , 57.4°E), (c) Apollinaris Patera on the summit of Apollinaris Mons (centred 9.3°S , 174.8°E) and (d) Eden Patera. Note the characteristic heavily eroded flanks on panels (a-c). All panels are based on THEMIS-day mosaics and north is up.

A group of four additional putative ancient volcanic centres represented by large circular caldera-like depressions – Peneus, Malea, Amphitrites and Pityusa Paterae – are found on Malea Planum, an ancient part of the southern highlands south of the Hellas basin (Greeley and Guest, 1987, Tanaka and Scott, 1987, Tanaka and Leonard, 1995, Tanaka et al., 2002, Plescia, 2003a, Plescia, 2004, Larson, 2007; Williams et al., 2009; Bernhardt and Williams, under review). While Amphitrites Patera (Fig. 2b) is located on a broad topographic rise similar to Hadriacus and Tyrrhenus Montes, the remaining three paterae are characterized by little or no appreciable positive topographic relief (Fig. 2b). All four of these potential calderas could be of explosive origin and may represent martian equivalents to large terrestrial calderas such as Yellowstone (Peterson, 1977; 1978). Furthermore, they show signs of heavy modification, probably by periglacial/permafrost, glacial, fluvial, and aeolian processes (Williams et al., 2010a; Bernhardt and Williams, under review). It was also proposed that the absence of channelled flanks and the lack a ‘softened’ appearance indicative of pyroclastic fallout deposits around the two southernmost paterae, Malea and Pityusa, might imply that the plains surrounding these paterae were formed by more consolidated materials, such as lava or ignimbrites (Williams et al., 2009). Recent stratigraphic analyses by Bernhardt and Williams (under review) also suggest that the wrinkle-ridged plains of the area actually superpose Malea and Pityusa Paterae, thereby obscuring any potential ring fractures and flow features, and casting a genetic relationship between these potential calderas and the surrounding plains materials into doubt. Additionally, there have been two formation scenarios for Malea and Pityusa Paterae that would not require explosive volcanism. Crumpler et al. (1996) proposed that the large paterae on Mars might be the result of collapses caused by the evacuation of magma chambers via parasitic, not necessarily explosive eruptions, albeit without specifically referring to Malea and Pityusa Paterae (which lack any visible parasitic cones or vents). Another scenario was lithospheric subsidence caused by the loads of sills and/or (partially) filled, crystallized magma chambers, i.e., without involving explosive volcanism (Larson, 2007). While such a scenario is similar to the suggested formation of Idaho’s Snake River Plain (e.g., DeNosaquo et al., 2009), it is not supported by the lack of any significant, positive free-air gravity anomalies beneath either patera as well as the occurrence of potential pyroclastic deposits inside Pityusa Patera (Bernhardt and Williams, under review).

Together these six potential volcanoes, historically termed as highland paterae, form the Circum-Hellas Volcanic Province, which covers more than 2.1 million km² (see Williams et al., 2009 for detailed review, Williams et al., 2010a) including the wrinkle-ridged plains of the Hellas basin floor and Malea Planum (e.g. Bernhardt et al., 2016; Bernhardt and Williams, under review). All of them are within the same range of age (Middle Noachian to Early Hesperian) and they appear to be the oldest recognized and preserved volcanoes on Mars (Plescia and Saunders, 1979; Crown et al., 1992, 2005, 2007; Leonard and Tanaka, 2001; Mest and Crown, 2001; Werner, 2005; Crown and Greeley, 2007). The results of crater-size frequency distribution analysis suggest that Tyrrhenus Mons and Peneus, Malea, and Pityusa Patera were formed around 3.8 Ga, and Hadriacus Mons and Amphitrites Patera at 3.7 Ga and 3.6 Ga, respectively (Williams et al., 2007, 2008; 2009), hence in the Late Noachian to Early Hesperian. This would make them the oldest central vent volcanoes on Mars. Furthermore, as previously mentioned, Malea and Pityusa Patera appear to be superposed by the surrounding wrinkle-ridged plains, and could thus be even older than the derived surface ages (Bernhardt and Williams, under review). It was also proposed that some of the constituent units of the Circum-Hellas Volcanic Province might have Hesperian to Early Amazonian ages (Leonard and Tanaka, 2001), but these young ages were later challenged by Williams et al. (2009). Stratigraphic relationships imply that these paterae formed after the Hellas impact (~4 Ga) and it is thought that they mark a transition from flood volcanism in central Malea Planum to more localized central vent volcanism. In the case of Hadriacus Mons its formation began even before the planetary dynamo ceased (Lillis et al., 2006). As already mentioned, several resurfacing events of various origins unevenly modified the flanks and paterae within the Circum-Hellas Volcanic Province after their formation. The least modified are Hadriacus Mons and Tyrrhenus Mons (Williams et al., 2009). Flow-like features in the vicinity of their central calderas are visible, which suggests that a transition from dominantly explosive to dominantly effusive eruptions may have occurred over time (e.g., Carr et al., 1977; Greeley and Spudis, 1981; Williams et al., 2007 and references therein).

The southern martian highlands also contain a few dozen isolated Noachian mountains, which may also be ancient volcanoes associated with explosive activity (e.g., Scott and Tanaka, 1981; Dohm and Tanaka, 1999; Xiao et al., 2012). Some of them are associated with unique knobby terrains which

based on their morphologies and thermophysical properties, are thought to be a result of explosive eruptions as they bear similarities to terrestrial eroded ignimbrites (Huang and Xiao, 2014). These ancient volcanoes seem to be associated with the Tharsis rise or Circum-Hellas Volcanic Province (Dohm and Tanaka, 1999; Xiao et al., 2012).

The most notable isolated mountain within the southern highlands is Apollinaris Mons, which is approximately 190 km-wide and 5 km-high (Fig. 1c), and is interpreted to be a shield volcano (Plescia, 2004). This volcano is situated near the equator, south of Elysium volcanic province, on the transition between northern plains and southern highlands in a local basin (Robinson et al., 1993; Scott et al., 1993), ~350 km north of the center of Gusev crater.

Apollinaris Mons has a ~80 km wide, flat-topped summit area interpreted to be a two-stage central caldera, named Apollinaris Patera, which is breached on the south-eastern side. In general, the flanks have an average slope of 5° (Plescia, 2004). A large fan of more resistant material appears to originate from this breach and covers the southern flank of the volcano. The origin of the fan covering the southeastern part of the volcano's flank was initially attributed to lavas (Robinson et al., 1993), but this interpretation was later challenged (see El Maarry et al., 2012 and references therein, El Maarry et al., 2014). Instead, it was proposed that the fan deposits are the result of pyroclastic flows, fluvial processes, or lahars formed due to drainage of a caldera lake, which may have formed within the summit area due to the hydrothermal activity. Based on its overall shape, surface texture and degraded flanks, it has been suggested that parts of the volcano's flanks are made up of friable materials, such as pyroclasts produced during large-scale explosive events (Robinson et al., 1993; Crumpler et al., 2007; Chuang et al., 2019). Age estimates derived via crater counting suggest that Apollinaris Mons was volcanically active during the Noachian and Hesperian periods (Werner 2009; Tanaka et al. 2014a), involving water-magma interactions and long-lived hydrothermal activity (El Maarry et al., 2012 and references therein). The volcano may have released more than 10^{15} kg of water vapour into the atmosphere (Robinson et al., 1993) during its activity and large amounts of pyroclasts. Large-scale explosive eruptions of this kind may have been responsible for extensive pyroclastic deposits on Mars (Kerber et al., 2012), such as the Medusae Fossae Formation (MFF) located on the northern and eastern flanks of this volcano and described in detail below in Section 3.2.3. III.

In Arabia Terra, several irregularly shaped and low-relief craters, namely Eden (Fig. 2d), Ismenia, Siloe and Euphrates paterae have been associated with evidence for explosive volcanism (Fig. 1, letters Q-U; Michalski and Bleacher, 2013). They are associated with lava plains and friable, layered deposits often bearing signs of the presence of sulphates. Based on the morphological and morphometrical characteristics, it was suggested by Michalski and Bleacher (2013) that these paterae may be the martian analogues of terrestrial ‘supervolcanoes’, although it should be noted that these interpretations are not universally accepted (Garcia, 2018). However, if such an interpretation is correct, these volcanic centres may be the source for the large quantity of fine-grained, layered deposits that are exposed throughout Arabia Terra (Malin and Edgett, 2000; Edgett and Malin, 2002; Hynek et al., 2003), including the characteristic units formed from tephra deposits, allophane and imogolite, identified at Mawrth Vallis (Bishop and Rampe, 2016). If this interpretation is true then other large depressions throughout the southern highlands, which are presently presumed to be impact features, could be calderas formed by explosive volcanism. Stratigraphic relationships revealed that they are likely Late Noachian to Early Hesperian in age (Michalski and Bleacher, 2013).

An atypical example of a patera is Alba Mons, an enormous shield with a diameter similar to that of Olympus Mons, is situated on the northern edge of the Tharsis volcanic province. The volcano is characterized by wide and low-relief flanks with slopes around 1.2° at the bottom part of the edifice, and a steeper shield in the upper part reaching an elevation of only 6.8 km at its highest point (Plescia, 2004). At the top of the shield is a complex caldera, known as Alba Patera, estimated as being active between 3.0 Ga and 300 Ma (Robbins et al., 2011). The flanks are covered by large lava flows, some extending nearly 1,000 km from the summit region, but they are also eroded by dendritic valleys indicating friable pyroclastic material modified by surface-runoff processes (Gulick and Baker, 1989) suggesting an explosive phase in the volcano’s history. Based on thermal inertia measurements and theoretical estimates, Mouginis-Mark et al. (1988) proposed that these deposits were emplaced by pyroclastic flows and that the presence of a small valley network within the summit area pointed to the presence of the thin exposed layer of pyroclasts. However, this was not confirmed by later data (Plescia, 2004). Still, the combination of extensive lava flows together with friable material and its low relief makes Alba Mons a unique martian volcano, which is believed to be an example of the transition

between older pyroclastic-dominated highland paterae and more recent effusive central-vent volcanoes dominating Tharsis and Elysium volcanic provinces (Mouginis-Mark et al., 1988; Raitala and Kauhanen, 1989; Ivanov and Head, 2006; Crown et al., 2019).

3.1. 2 Tholi and other shields

The surface of Mars also contains other mountains of various sizes interpreted to be volcanoes ostensibly showing signs of explosive activity. These include, for example, Ceraunius Tholus and Uranus Tholus situated within the Tharsis region, whose flanks seem to be covered by mantling deposits suggesting that explosive volcanism participated in their formation (Crumpler et al., 1996; Plescia, 2000). Pyroclastic material was also suggested to be present on the flanks of Tharsis Tholus (Robinson, 1993), but later refuted (Plescia, 2003b). Similarly, it was proposed that Hecates Tholus (Fig. 3a) lying at the northern edge of Elysium Planitia may be partly composed of pyroclastic deposits (see Section 3.2.9 for details). But again, this conclusion was questioned as alternative explanations including, for instance, glacial erosion or the degradation of surface lava flows by an active hydrothermal system were proposed (Plescia, 2004; Mouginis-Mark and Christensen, 2005 and references therein). Later it was recognized that the north-western flank of Hecates Tholus contains a 10 km wide depression (Fig. 3b), interpreted to be a peripheral caldera formed by an explosive event ~350 million years ago (Hauber et al., 2005; Werner, 2009; de Pablo et al., 2013). Therefore, Hecates Tholus is one of the best examples for large, recent (Middle Amazonian) explosive eruptions on Mars.

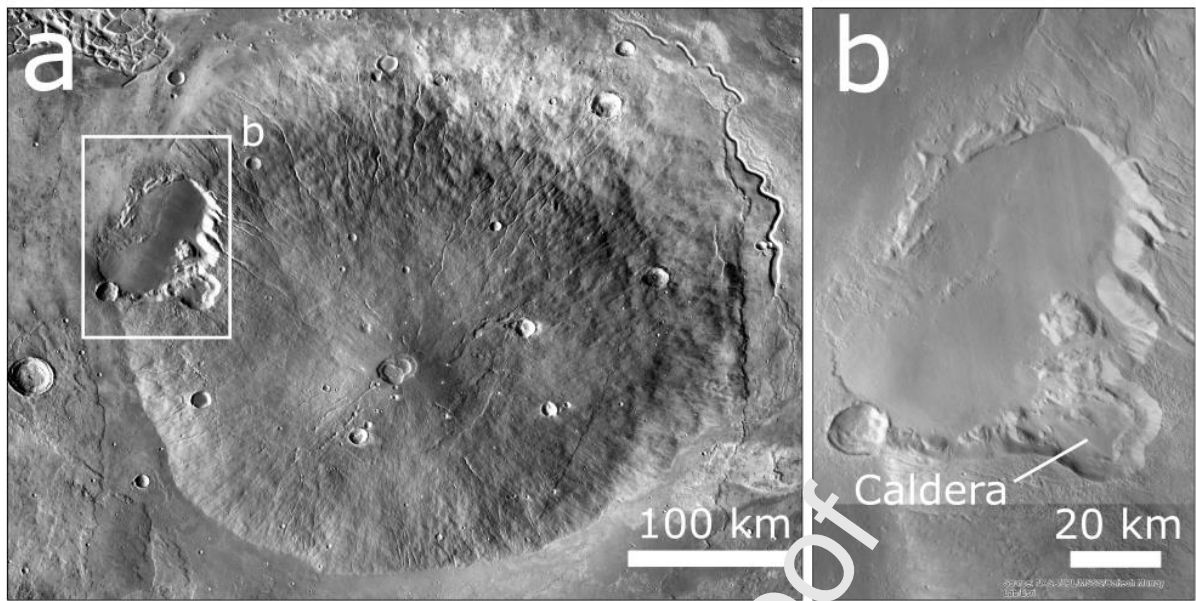


Figure 3: Image showing an example of a Tholus, Hecates Tholus (a) situated at the northern edge of Elysium Planitia. Panel (b) is a detailed view of the proposed flank caldera that is thought to be associated with pyroclastic deposits. Panel (a) is centered at 31.9°N , 150.1°E and uses the THEMIS IR Day 100m Controlled Partial Mosaic provided by USGS, panel (b) is based on THEMIS-day mosaic and Murray Beta CTX (Dickson et al., 2018) mosaic and north is up in all panels.

Another interesting example is Zephyria Tholus situated within the Aeolis region. This is a ~ 36 km wide and ~ 3 km high circular edifice with a ~ 8 km wide summit crater. The volcano has steep flank slopes up to 14° . Based on its shape, size and stratigraphy it was proposed that this edifice is akin to a terrestrial composite volcano (formerly termed stratovolcanoes in the literature) suggesting alternating explosive and effusive volcanic activity in the Early Noachian period (Stewart and Head, 2001). However, this interpretation is based on Viking images accompanied by MOLA altimetry and limited coverage of images obtained by the Mars Orbiter Camera with relatively low spatial resolution, and a more detailed study using newly available data has yet to be done to confirm this interpretation.

Several studies also linked the large martian volcanoes, such as Olympus Mons, commonly classified as typical effusive volcanoes, with explosive volcanism based on theoretical and observational evidence (e.g., Bleacher et al., 2007b; Wilson and Head, 2007; Kerber et al., 2012; Peters and Christensen, 2017). Based on the enormous height of this volcano – Olympus Mons is

around 22 km high – it was suggested that the atmospheric pressure would be so low at the summit that most of the eruptions would be explosive even with a small amount of volatiles (Head and Wilson, 1998a,b). Additionally it has been hypothesized that due to the height of the volcano ice may have accumulated and later been buried on the middle to upper flanks of the volcano (e.g., Tanaka, 1985; Neukum et al., 2004). This scenario should give rise to water-magma interaction resulting in hydrovolcanic features. However, a detailed study characterizing the morphology and distribution of the dozens of flank vents and grabens on Olympus Mons, based on high-resolution images, revealed that effusive eruptions have dominated and that explosive eruptions were rare at least in the Late Amazonian (Peters and Christensen, 2017). This suggests that magmas in the Late Amazonian were volatile-poor and/or there were no external volatiles available to trigger the explosive activity. However, the question of whether explosive activity has played a role in the formation of Olympus Mons, is still open (Mouginis-Mark, 2018).

Explosive activity has also been proposed to have played a role in the construction of another large shield volcano within the Tharsis region, Arsia Mons, which shows widespread deposits (Scott and Zimbelman, 1995), mainly on its western flank, interpreted to be pyroclasts (Mouginis-Mark, 2002; Richardson et al., 2017), and the presence of iron-bearing glass (Lai et al., 2019). The pyroclastic deposits might have been produced in such quantities that they are responsible for the preservation by burial of ice-rich deposits on its flanks as well as providing the debris required to form relict glacial landforms (Hean and Weiss, 2014). The other Tharsis Montes also host landforms interpreted to originate from cold-based glaciation (e.g., Shean et al., 2005; Kadish et al., 2014), yet the apparent lack of explosive volcanism associated with the purported long-lived presence of this ice seems contradictory (e.g., Cassanelli and Head, 2019).

The volcanic activity associated with the predominantly effusive Syrtis Major, a low-relief volcanic shield situated in the martian highlands on the western edge of Isidis Basin (Meyer and Grolier, 1977), was probably partly explosive as documented by the presence of pyroclastic deposits within the central irregular, 200 km × 400 km wide caldera (Fawdon et al., 2015) and the small topographic difference between the caldera floor and its rim (e.g., Robbins et al., 2011). This complex caldera consists of two separate smaller edifices, Meroe Patera to the south and Nili Patera to the

north. Both calderas have been active in the Hesperian (Robbins et al., 2011). Fawdon et al. (2015) showed that pyroclastic deposits are present within Nili Patera and proposed that the distinctively bright floor of the caldera may be remnants of welded ignimbrite(s) associated with caldera formation, remnants of lava flows, or part of a felsic pluton. If this bright material is an ignimbrite, this would point to large-scale explosive eruptions in the past. Olivine-rich ash-like deposits have been also detected in the vicinity of Nili Fossae (Bramble et al., 2017; Rogers et al., 2018; Kremer et al., 2019; Mandon et al., 2020); however, it remains unclear whether they are related directly to volcanic activity of Syrtis Major or to volcanic activity which occurred in the greater Syrtis Major–Isidis Planitia region on early Mars (Kremer et al., 2019). More details may be provided by a planned in situ investigation of the Jezero crater by the Mars 2020 rover. Syrtis Major may also be a source of pyroclastic surges, a kind of pyroclastic density currents capable of moving hundreds of kilometres from their source and possibly modifying the surface of Isidis Basin by their devolatilization (e.g., Ghent et al., 2012). However, other interpretations of the small conical edifices, believed to be evidence of such devolatilization within the Isidis Basin, exist (e.g., Guidat et al., 2015; Souček et al., 2015). Despite these numerous observations, the evidence for explosive eruptions in association with Syrtis Major remains spatially and temporally limited, with the edifice being dominated by lava flows, which together with the general shape of the volcano suggest that explosive volcanism played only a minor role in the evolution of this volcano (Robbins et al., 2011).

3.1.3. Scoria cones

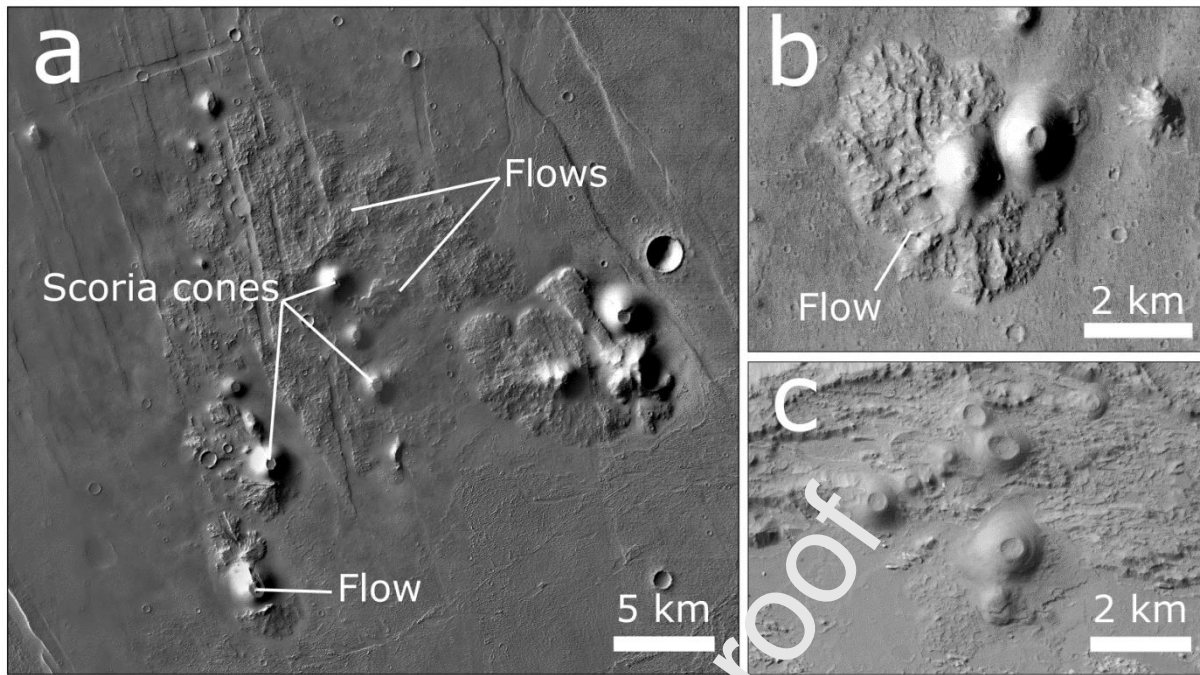


Figure 4: Image showing three different areas containing promising candidates for martian scoria cones. (a) A CTX mosaic of a group of cones in Ulises Colles situated within Tharsis (centered at 5.9°N , 237.1°E), (b) two cones superposed on flow-like unit in Hydraotes Colles situated within Hydraotes Chaos based on a CTX image (centered 0.2°N , 326.2°E) and a (c) cluster of small conical features within Coprates Chasma based on HiRISE image ESP_034131_1670 (centered 12.73°S , -29.72°E). North is up in all panels.

The presence of scoria cones (also called “cinder cones”) was first suggested for several regions of Mars based on Viking data (Carr et al., 1977; Frey and Jarosewich, 1982; Edgett, 1990; Hodges and Moore, 1994; Plescia, 1994), but the interpretations were ambiguous due to the low resolution of available images. However, the lack of certainty in interpretation did not prevent study of their possible shapes and the way the ejected material would be emplaced by theoretical and numerical studies (Wood, 1979a,b; Dehn and Sheridan, 1990; Wilson and Head, 1994; Fagents and Wilson, 1996).

More recent high-resolution images allow us to recognize several sites where martian scoria cones may be present, and reveal details on their morphologies and morphometries. Good examples are found:

1. on the flanks of the larger shield volcanoes, such as Pavonis Mons (Bleacher et al., 2007a; Keszthelyi et al., 2008), Olympus Mons (Peters and Christensen, 2017), and Tharsis Tholus (Platz et al., 2013),
2. as clusters in large volcanic fields, e.g., Ulysses Colles within the Tharsis bulge (Fig. 4a, Brož and Hauber, 2012; Brož et al., 2014; 2015a), Hydraotes Colles in the chaos terrains (Fig. 4b, Meresse et al., 2008; Brož et al., 2015a; 2017),
3. on the floor of chasmata, including Coprates Chasma, the deepest trough of Valles Marineris, (Fig. 4c, Harrison and Chapman, 2008; Brož et al., 2015a, 2017), within the outflow channel of Dao Valles (Korteniemi and Kukkonen, 2018),
4. in the northern lowlands (Lanz et al., 2010),
5. as individual cones in volcanic provinces in Tharsis (Keszthelyi et al., 2008; Hauber et al., 2009).

However, putative scoria cones were mainly described as a subordinate feature accompanying other principal processes (Meresse et al., 2008; Hauber et al., 2009; Lanz et al., 2010), or are one of several alternative explanations, including for instance, sedimentary volcanism (e.g., Skinner and Tanaka, 2007; Harrison and Chapman, 2008; Okubo, 2016). Sometimes, such as for the large field of pitted cones on the floor of Coprates Chasma, the debate as to whether the observed cones are sedimentary volcanoes or scoria cones is still not settled (e.g., Okubo 2016; Brož et al., 2015a, 2017; Wheatley et al., 2019; Kumar et al., 2019).

Recent high resolution data allow the study and characterization of the shapes of martian putative scoria cones in detail. Meresse et al. (2008) firstly reported the basic characteristics for a field of circa 40 cones of Hydraotes Colles on the floor of the chaos terrain of Hydraotes Chaos located 5 km below the martian datum. These cones are predominantly located in the southern part of the chaotic terrain over a 40×30 km area. They have central craters and often form small subclusters separated by 5 km. Meresse et al. (2008) showed that the cones are small, 0.5 to 1.5 km in diameter, with heights between 180 and 230 m, and with a mean slope from 15 to 31°. Those values showed that studied cones are within the same range of values as those of their terrestrial analogues (e.g. Wood, 1979a, 1979b, Brož

and Hauber, 2012 and references therein). Additional details came from the study of the field of 29 cones named Ulysses Colles likely formed in the period 1.5–0.44 Ga (Brož and Hauber, 2012). This volcanic field is located at a height of 4.5 km above the martian datum over an area of about 80×50 km at the southern edge of Ulysses Fossae within the Tharsis region, an area dominated by igneous volcanism, making this field unique in its lack of plausible alternative hypotheses. Detailed morphometrical investigation of these cones revealed that they are wider (up to 4080 m), higher (up to 650 m) and more voluminous (up to $4.2 \times 10^9 \text{ m}^3$) than terrestrial scoria cones, yet have smaller average slopes ranging from 7° to 18° (Brož and Hauber, 2012; Brož et al., 2014; 2015a). This finding was later confirmed by a further study performed on 28 cones within three clusters in Ulysses Colles, Hydraotes Colles and Coprates Chasma (Brož et al., 2015) and was explained by the different environmental factors between Earth and Mars (see details below).

However, despite the increasing evidence for martian scoria cones, they are not as common as on Earth. The small number of sites containing these features suggests that either pyroclastic volcanism of this type was never dominant on Mars, or that it was in the past, but the surface expression has been covered by younger effusive volcanism or destroyed by erosion (Zimbelman et al., 2015). Both are interesting conclusions deserving more detailed investigation in the future.

3.1.4. Spatter cones

The presence of spatter cones was expected in martian volcanic provinces because they are common features on terrestrial basaltic volcanoes and in associated volcanic plains (e.g., Blackburn et al., 1976). However, due to their relatively small size they are not easy features to unambiguously recognize and study from remotely sensed data. To make a good identification of a spatter cone, the spatter particles consisting of fine tephra or some other fragmented material (Hauber et al., 2009a) should be identifiable, which is at the limit of our current satellite imaging capability. Despite these difficulties, spatter cones have been reported on the flanks of Alba Mons (Cattermole, 1986), Arsia Mons (Head et al., 2005), Olympus Mons (Peters and Christensen, 2017) and on edifices accompanying the low shield volcanoes within the Tharsis region (Keszthelyi et al., 2008; Hauber et

al., 2009). An additional complication in their recognition on Mars is that the lower atmospheric pressure and gravity should cause a wider dispersion of ejected material, suggesting that shapes of spatter cones may be different from those of their terrestrial analogues. On Earth, spatter cones are generally less than several meters wide, up to 5 meters high, and have steep flanks exceeding the angle of repose of the fragmental material because of fusion of the molten ejected particles. A dedicated study of putative martian spatter cones addressing their formation mechanism and shapes under martian conditions should be a priority for future work.

3.1.5. Small Hydrovolcanic features

Given the abundance of volcanic activity and ground ice on Mars over a wide range of latitudes (e.g., Baker, 2001; Feldman et al., 2004; Smith et al., 2009; Byrne et al., 2009; Vincendon et al., 2010), interaction between magma and water is expected throughout martian history (e.g. Head and Wilson, 2007 and references therein). This is because ascending dikes would have to propagate through the martian cryosphere causing a rise of hydrovolcanism (Sheridan and Wohletz, 1983). As magma-water interactions may lead to the formation of active hydrothermal environments, hydrovolcanic features are promising targets for future astrobiological studies (e.g., Cousins and Crawford, 2011; Cousins et al., 2012). On Earth, hydrovolcanism can produce various morphological features depending both on the physical state of the water and the style of the volcanism (for summary see Keszthelyi et al., 2010). The full spectrum of hydrovolcanic edifices has been suggested to be present on Mars; namely dikes, sills and ridges, large-scale ground collapses and chaotic terrains, outflow channels caused by the release of meltwater, lahars, subglacial edifices like tuyas (discussed in Section 3.1.5.1) and pyroclastic deposits. Also, constructional edifices such as tuff rings, tuff cones, maars (discussed in Section 3.1.5.2) or rootless cones (discussed in Section 3.1.5.2) have been suggested to be present on the surface of Mars (e.g. Head and Wilson, 2007; Keszthelyi et al., 2010) based on orbital data as well as based on ground observations taken by rover Spirit within Home Planitia (discussed in Section 4).

It has also to be noted that the water may not be the only external volatile that can trigger explosive eruptions on Mars. This is because the martian environment allows CO₂ ice to be stable on the surface and below, and therefore magma can potentially interact with this volatile as well. However, how such interactions as well as the shapes of resulting features would look like is currently unknown.

3.1.5.1. *Tuyas*

The hydrovolcanic features expected to be most commonly present on the surface of Mars are tuyas (e.g., Chapman and Tanaka, 2001; Smellie and Chapman, 2002; Head and Wilson, 2007). Tuyas are flat-topped and steep-sided volcanic edifices formed by subglacial eruptions (Martínez-Alonso et al., 2011) during which pillow lavas, tuffs and lava caps might be formed (e.g., Jones, 1969, 1970; Gudmundsson et al., 1997). When the amount of lava is sufficient to completely melt the overburden of ice, explosive activity may occur forming pyroclasts. Tuyas were widely postulated to occur on Mars in order to explain the abundant and enigmatic mesas commonly found on the martian surface (e.g., Hodges and Moore, 1979; Nedell et al., 1987; Croft, 1990; Lucchitta et al., 1994; Chapman and Tanaka, 2001; Farrand et al., 2008; Keszthelyi et al., 2010; Kress and Head, 2015; Campbell et al., 2016; Scanlon et al., 2014). The most characteristic edifices are situated in Chryse and Acidalia Planitiae (Fig. 5; Martínez-Alonso et al., 2011). However, confirmation of their interpretation as tuyas, such as observations of pillow lavas and columnar jointing, is still missing (Keszthelyi et al., 2010) due to image resolution problems. Hence, this interpretation remains uncertain. Other promising candidates include the hills near Arsia Mons (Scanlon et al., 2014) and hills in southern Sisyphi Montes located between the Argyre and Hellas impact basins (Farrand et al. 2008), which show mineralogic evidence supporting subglacial volcanism (Campbell et al., 2016).

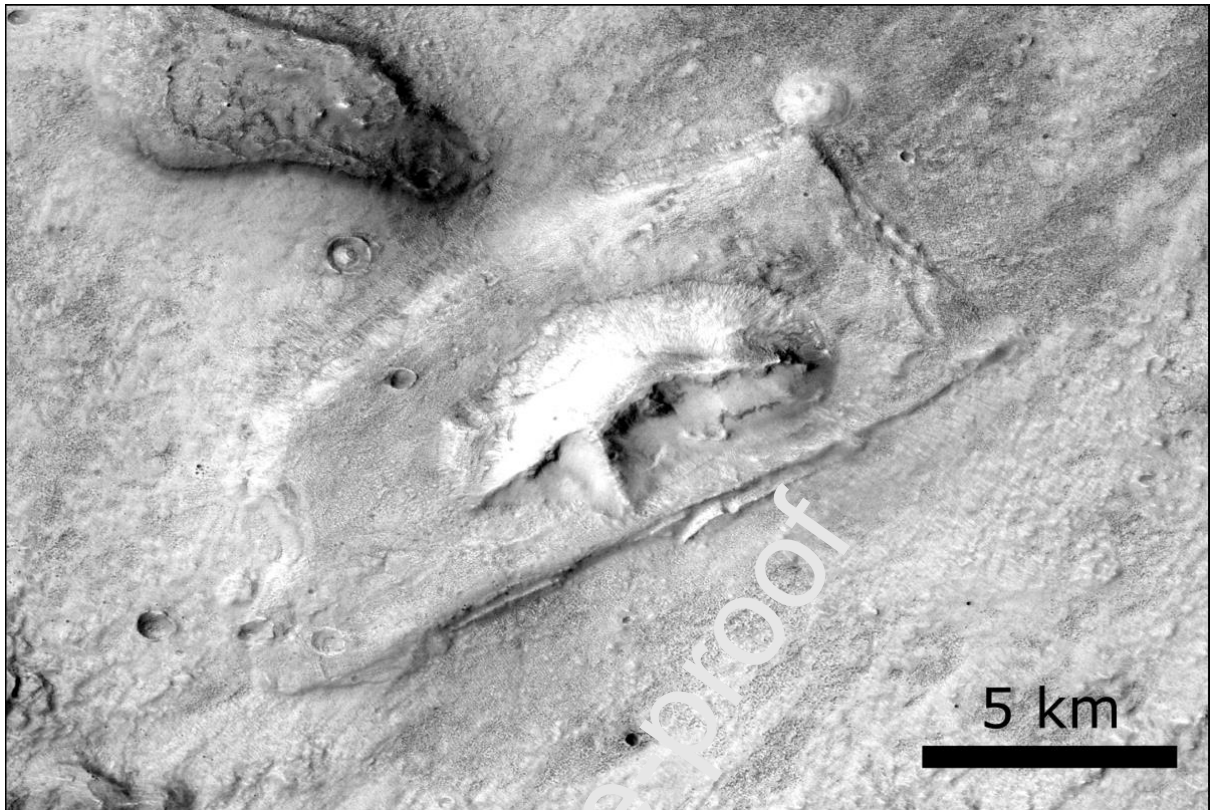


Figure 5: An example of a possible martian tuff cone from Acidalia Planitia represented by a flat-topped and steep-sided mesa with a central knob. CTX image P19_008430_2212_XN_41N036W centred at 41.0°N, 323.57°E, north is up.

3.1.5.2. Tuff rings, tuff cones and maars

The explosive eruptions resulting from hydrovolcanism can fragment the ascending magma together with the surrounding rocks into fine-grained pyroclasts because of the conversion of groundwater or buried water ice into the expanding steam. Then these fragments can be ejected from the vent to the close surrounding where they can accumulate and build a constructional edifice. It has been proposed that such mechanism played a role in the formation of various features on Mars. For example, it has been suggested that hydrovolcanism played a role in the formation of the fretted channel of Ismenia Fossae containing a crater-like feature (Carruthers and McGill, 1998), outflow channels on the western flanks of Elysium Mons (Russell and Head, 2003), several linear ridges at the edge of the Olympus Mons aureole materials (Wilson and Mouginis-Mark, 2003a), various martian depressions and pits surrounded by ballistically-emplaced deposits (Pedersen et al., 2010; Levy et al.,

2017), on the south-western flank of Ascræus Mons (Scott and Wilson, 1999), around the Hellas basin (see Bernhardt et al., 2016 and references therein) or in the opening of Mangala Valles outflow channel (Wilson and Head, 2004). Also the formation and evolution of Hrad Vallis, an ~800 km long and ~370 m deep depression in Elysium Planitia, which is associated with distinctive deposits surrounding the valley for at least 100 km, has been associated with hydrovolcanism, large aqueous outbursts from the subsurface and with the formation of subsequent lahars (Wilson and Mouginis-Mark, 2003b; Morris and Mouginis-Mark, 2006). However, Hopper and Leverington (2014) who proposed that the valley was incised by low-viscosity lavas later challenged this interpretation. Later it was proposed that Hrad Vallis formed by combination of both processes (Hamilton et al., 2018).

The deposition of pyroclasts from hydrovolcanism may also be more localized and form kilometre-sized features, such as tuff rings, tuff cones, or maar. The detection of these features on the surface of Mars was historically expected (Carruthers and McGill, 1998), but insufficient image resolution does not allow unambiguous identification of such features. This is because various geological processes may form similar looking features, for example processes such as meteorite impacts, which subsequently degrade, sublimation of ice, or sedimentary volcanism. This ambiguity can be illustrated by cratered cones situated near the northern martian pole, which were originally interpreted as martian tuff cones (Hodges and Moore, 1979), but high-resolution data revealed no supporting evidence for this interpretation (Keszthelyi et al., 2010). However, even with the availability of high-resolution data, the detection of such features is still a complicated task. Consequently, observations of constructional landforms formed by hydrovolcanism on Mars are sparse and only a handful of dedicated studies exist.

The most promising site where tuff cones and tuff rings may be present is the Nephentes/Amenthes region (Fig. 6), in which a large field of kilometre-sized pitted cones and domes have been described and their morphologies and morphometries have been measured (Brož and Hauber, 2013). These features show many similarities in shape with terrestrial tuff rings and tuff cones. They are wider and taller than their terrestrial counterparts, which is attributed to the lower martian gravity and atmospheric pressure allowing wider dispersion of ejected particles. This area

contains a candidate maar structure (Fig. 6d). Similar looking conical edifices were also described by the same authors in Lederberg crater situated on the edge of Chryse Planitia and in the Arena Colles region on the north-western edge of Isidis Planitia, hence in localities with a different geological context, suggesting that the studied features do not require a unique setting, like sedimentary basins, to be formed. However, alternative explanations still exist; for example, the features within the Nepenthes/Amenthes region could instead be sedimentary volcanoes (Skinner and Tanaka, 2007). At the moment the available data are not sufficient to resolve this dispute.

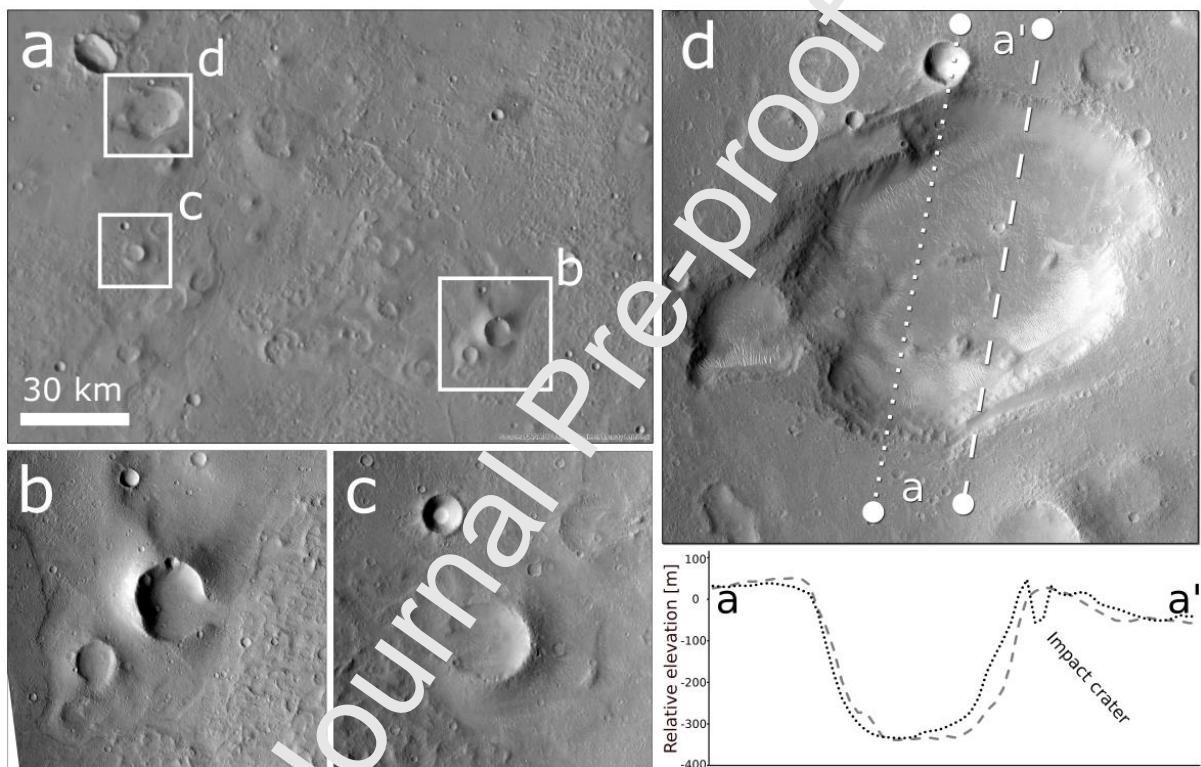


Figure 6: An example of pitted cones in the Nepenthes/Amenthes region (centred 17.4°N , 102.6°E) interpreted as possible martian analogues to terrestrial tuff rings and tuff cones by Brož and Hauber (2013) or as sedimentary volcanoes by Skinner and Tanaka (2007). Cones are often breached (b,c) and they have a wide central crater with floor elevations sometimes below the surrounding surface level. Panel (d) shows a non-circular composite depression with slightly elevated rims, which may represent a martian maar. Topography based on two individual MOLA track profiles. North is up on all panels.

The presence of maar-diatreme structures and tuff cones has also been proposed in Simud–Tiu Valles (Pajola et al., 2016), Argyre Planitia (Williams et al., 2017), the floor of Coprates Chasma (Brož et al., 2017), and Niger and Dao Valles (Korteniemi and Kukkonen, 2018). However, these studies do not contain detailed morphometrical and/or morphological evidence to support their interpretation. Conical edifices within the northern Terra Sirenum were suggested to result from hydrovolcanism, but sedimentary volcanism was subsequently favoured (Hemmi and Miyamoto, 2017).

3.1.5.3. Rootless cones

Rootless cones are the best described yet highly controversial potential explosive edifices on Mars. On Earth, they are formed by magma-water interaction when lava interacts with groundwater, ice or wet sediments and produces pyroclastic material, which is then deposited on the surface of a lava flow (e.g., Fagents and Thordarson, 2007). The first observations of possible rootless cones on Mars were noted from Viking orbiter images (Frey et al., 1979; Allen, 1979) in Acidalia, Utopia, Elysium and Isidis Planitiae, Hephaestus Fossae, the north of Olympus Mons aureole and Arrhenius province (Frey et al., 1979; Frey and Jaosewich, 1982; Mouginis-Mark, 1985; Hodges and Moore 1994). However, the low resolution of the Viking images did not allow for an unambiguous interpretation. Later on, Mars Orbiter Camera (MRO) acquired high-resolution images (Malin et al., 1992) of similar regions providing more detail on the morphology of these cones (Fig. 7). Similar cone groups were also identified from the mid- to northern latitudes (0°-50°N) of Mars (Fagents and Thordarson, 2007; Mouginis-Mark, 1985; Keszthelyi et al., 2010) that stimulated the debate about the presence of volatiles and potential lava flows in this region. The cone groups of Amazonis Planitia are up to ~100 m in diameter (Fagents and Thordarson, 2007) and well- preserved in compared to those which are identified on Acidalia Planitia and Isidis Planitiae (Fagents and Thordarson, 2007; Ghent et al., 2012). However, these interpretations are not unambiguous as similar-looking landforms can be formed by non-volcanic processes, namely by the action of ice or mud (Tanaka et al., 2003; Farrand et al., 2005; Burr et al., 2005), and therefore the debate about their origin is still not settled.

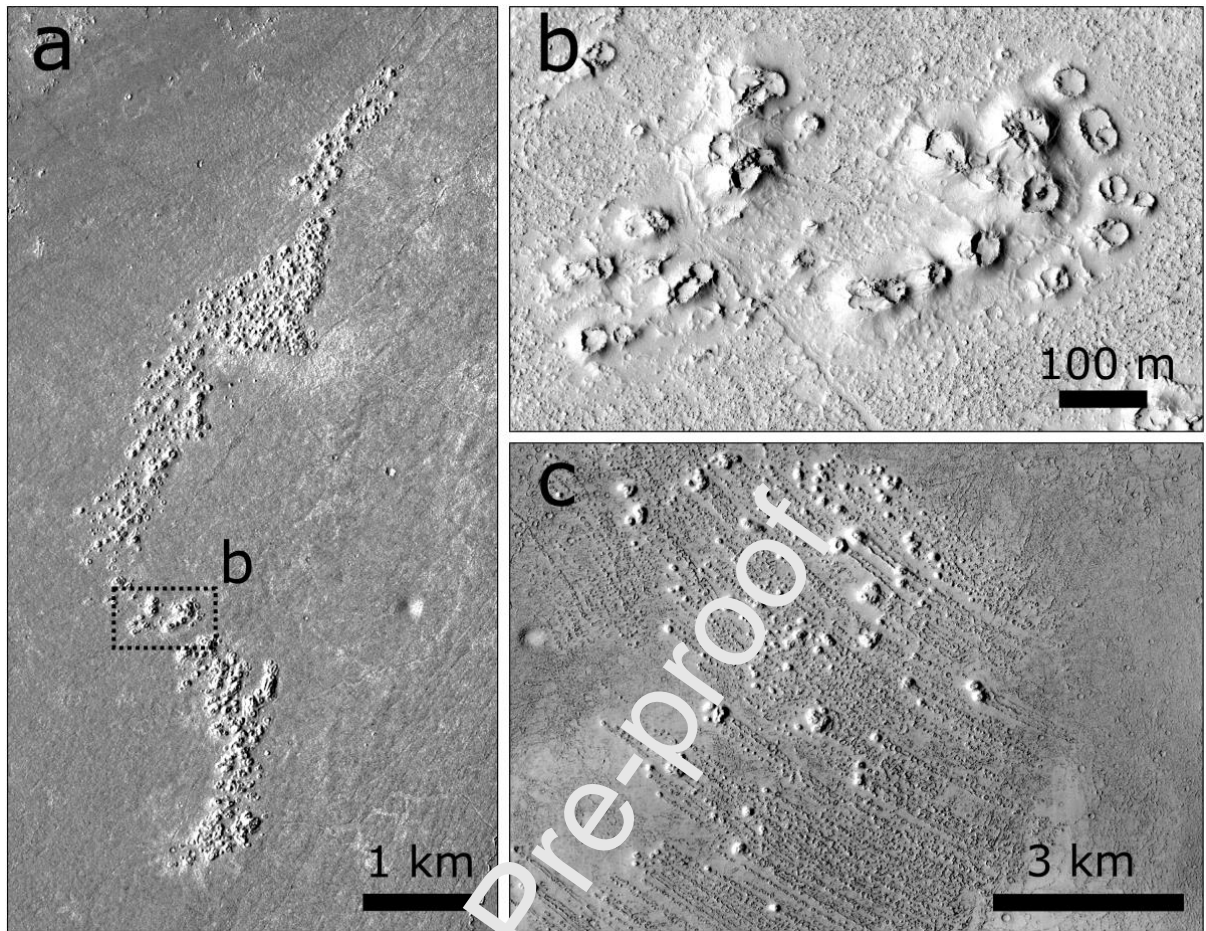


Figure 7: Examples of putative martian rootless cones within Athabasca Valles (a, b) and in the western Tartarus Colles region (c). Proposed rootless cones are characterized by small conical edifices with pitted summit craters. Cones are often clustered (a, b) and associated with chain-like features suggesting the direction of the lava flow movement (c). Panel (a) is based on CTX image P03_002147_1865_XI_06N208W, centred at 7.38°N, 152.81°E, (b) is based on HiRISE image PSP_002292_1875, and (c) is based on CTX image G01_018457_2065_XN_26N186W, centred 26.22°N, 173.57°E. North is up on all panels.

To understand the surface conditions and to assess the formation mechanism of rootless cones on Mars, terrestrial cones are often used as analogues (Noguchi et al., 2016; Hamilton et al., 2010; Fagents and Greeley, 2001). Icelandic rootless cone groups (Noguchi et al., 2016, Greeley and Fagents, 2001) as well as ice mounds in various regions of Alaska (Bruno et al., 2006) have been studied as they show resemblance in their morphological appearance and surface characteristics to

Mars' cones (Noguchi et al., 2016; Greeley and Fagents, 2001). A statistical analysis of randomness of the spatial distribution of cone groups was performed by using Nearest Neighbour Analysis based on skewness and kurtosis features (Bruno et al., 2006) in an attempt to reveal if their spatial distribution can help to resolve the dispute about their origin. According to this analysis, the terrestrial rootless cones are less skewed and leptokurtic distribution in comparison to ice mounds. Similarly for Mars, spatial distribution was used as a tool to distinguish rootless cone groups identified on Cerberus Plains, Acidalia Planitia, Isidis Planitia and Athabasca Valles (Bruno et al., 2006). Putative rootless cone groups in Cerberus Palus and Central Elysium Planitia are aligned in the direction of flow in a linear chain format (Stacey and Kerber, 2017) and densely populated within the center of the Athabasca Valles (Noguchi and Kurita, 2015).

For the formation of rootless cones, a certain amount of surface or near-surface water ice (or other volatiles) is essential. Thus, the spatial and temporal distribution of rootless cones on Mars can lead to a better understanding of the hydrothermal history, climate condition and surface evolution throughout martian history. For example, the distribution, size, and morphological characteristics of putative rootless cones has helped to assert the recent existence (<10-100 Ma) of ground ice in martian regolith in various localities (Fagents et al., 2002). Using numerical modeling, Dundas and Keszthelyi (2013) showed that a lava flow above a certain thickness is not able to produce hydrovolcanic eruption as the steam is trapped inside the thick layer of lava. Based on the outcrop and measured thickness of the lava flow in Tartarus Cones region, it was inferred that lava thicknesses between >30-60 m with an underlying ice layer located at <42 m in depth is optimal to have a rootless cone-producing eruption (Hamilton et al, 2010a).

3.2. Pyroclastic deposits

Identifying pyroclastic deposits, i.e., the fine-grained products of explosive volcanism, on the surface of Mars via remote sensing techniques is non-trivial (e.g., Criswell and Elston, 1981; 1982; de Silva et al., 2010; de Silva and Bailey, 2018) and alternative formation hypotheses should be maintained until more ground truth becomes available. So far, the most common remote observations

motivating interpretations of martian landscapes as pyroclastic deposits are a high friability (based on the observation of erosional landforms, e.g., yardangs and mass wasting [e.g., de Silva et al., 2010]), a small inferred grain-size (based on low thermal inertia), and geographic proximity to potential vent structures (e.g., Grant and Schultz, 1990; Moore, 1990; Tanaka, 2000; Chapman, 2002; Mandt et al., 2008). Layering as indicated by exposed alternating lineations/ridges or terraces, low albedo materials (i.e., potential mafic fines, e.g., as outcrops of low reflectance layers), resurfaced (i.e., less cratered) areas in proximity to volcanic vents/constructs, as well as unconformably draping over (or “mantling”) the landscape are further characteristics used for identifying potential pyroclastic deposits (e.g., Mouginis-Mark et al., 1982; Scott and Tanaka, 1982; Moore, 1990; Tanaka, 2000; Bradley et al., 2002; Chapman, 2002). Adding up the areas of deposits that have previously been interpreted as potentially pyroclastic in origin based on these characteristics (Fig. 1 and Table 1) results in a total of ~12.4 to ~13.4 million km², i.e., 8.6 to 9.3 % of the martian surface. Thickness estimates mostly based on stratigraphic and morphometric observations that are occasionally aided by ground-penetrating radar measurements (Orosei et al., 2017; Whetter et al., 2020) allowed the inference of volumes (Table 1), which potentially total ~8.37 to 9.93 billion km³ across the entire planet. In general, however, our knowledge of the overall volume of pyroclastic deposits on Mars is limited.

On the one hand, none of the indicative observations listed in Table 1 are truly diagnostic and the listed deposits might all have different origins. On the other hand, Mars’ pyroclastic inventory might be even more extensive if, instead of impact brecciation of deposits from effusive volcanism, pyroclastics from explosive volcanism are the source for the mechanically weak, mafic material forming the martian megaregolith (Huang and Xiao, 2014) and therefore large parts of the crust. Moreover, based on globally dispersed, likely fine-grained but somewhat lithified deposits enriched in olivine and pyroxene, the abundance of pyroclastic deposits as well as their alteration products throughout the crust is now believed to be much greater than previously thought (Wilson and Head, 2007; Horgan and Bell, 2012; Rogers et al., 2018; Kremer et al., 2019).

The actual mass contained in the potential pyroclastic deposits on Mars is also not well-constrained, as density is a largely unknown factor that is likely to be extremely variable, even within

a given deposit (e.g., King and Riehle, 1974; Crown and Greeley, 1993; Orosei et al., 2017). This is because large porosity variations can result from welding as determined by flow thickness, emplacement temperature, composition, and residual volatile content (e.g., Sparks et al., 1978; Crown and Greeley, 1993; Wilson and Hildreth, 2003), of which only flow thickness as a factor can be reasonably constrained for at least some locations in several martian deposits. Nevertheless, at least in areas proximal to a vent, where emplacement temperatures would still be high, the apparent erosive resistance as well as layering observed in several potential pyroclastic deposits (Table 1) was suggested to indicate welding, therefore potentially implying welded tuff-like densities of $>2000 \text{ kg/m}^3$ (e.g., Sheridan and Ragan, 1976; Crown and Greeley, 1993). In general, thick ($\sim 100 \text{ m}$ on Earth, i.e., $\sim 300 \text{ m}$ on Mars) pyroclastic deposits as they are suggested across Mars (Table 1) would be expected to cause significant compaction, i.e., welding regardless of temperature, thus potentially resulting in basalt-like densities $>2500 \text{ kg/m}^3$ (e.g., Ragan and Sheridan, 1972; Wilson and Hildreth, 2003; Quane et al., 2009).

Table 1: List of martian deposits larger than 10^5 km^2 (black polygons in Fig. 1), for which a pyroclastic origin has been proposed in previous literature (sorted by minimum volume estimate). Indicative observations are number-coded as follows: 1) Proximity to potential vent(s); 2) Friability; 3) Layering; 4) Small grain-size; 5) Mantle-like draping over topography; 6) Crater deficiency implying resurfacing; 7) Dark outcrops; 8) Potential degassing structures. Representative CTX images of each deposit are shown in Fig. 8.

| Name | Approximate areal extent (10^6 km^2) | Estimated volume (10^6 km^3) | Indicative observations | | | | | | | | Estimated time frame | References |
|--------------------|--|--|-------------------------|-----------------|---------------|-----------------|--------------|------------------|--------------------|--------------------------------|----------------------|--|
| | | | 1 vent s | 2 friability | 3 layering | 4 grain-size | 5 draping | 6 resurfacing | 7 dark outcrops | 8 indications for degassing | | |
| I) Arabia deposits | 5 | <5 | X | X | X | X | | | | | Early to Mid- | Moore, 1990; Tanaka, 2000; Hynes et al., |

| | | | | | | | | | | | | |
|---|-------------|-----------------|---|---|---|---|---|---|---|--|---|--|
| | | | | | | | | | | | Noachia n (?) | 2003; Michalski and Bleacher, 2013 |
| II) Electris deposits | >1.8 | 1 | | X | X | X | | | | | Late Noachia n | Grant and Schulz, 1990; Tanaka, 2000; Wendt et al., 2012 |
| III) Medusae Fossae Formatio n (includin g Gusev crater) | 2.5 | 1 to 1.4 | X | X | X | X | X | X | | | Hesperia n | e.g., Scott and Tanaka, 1982; Bradley et al., 2002; Squyres et al., 2007; Mandt et al., 2008, 2009; Zimbelman and Scheidt, 2012; Ojha and Lewis, 2018 |
| IV) Dorsa Argentea Formatio n (“South Polar Pitted Terrain”) | 1.35 | 0.68 to 1.35 | X | X | X | | X | | X | | Hesperia n | Head and Pratt, 2001; Kress and Head, 2015; Scanlon et al. 2018; Whitten et al., 2020; this study |
| V) Hellas deposits | 0.54 to 1.4 | 0.48 to 1.1 | | X | X | X | | | | | Early Hesperia n | Tanaka and Leonard, 1995; Tanaka et al., 2002; Williams et al., 2009; Bernhardt et al., 2016 |
| VI) Argyre deposits | 0.34 | 0.34 to 0.68 | X | X | X | | | | | | Early Hesperia n to Early Amazon ian | Dohm et al., 2015; Williams et al., 2017 |
| VII) Tyrrhena Patera deposits | 0.2 to 0.32 | 0.18 to 0.22 | X | X | X | | | | | | Hesperia n (?) | Gregg et al., 1998; Gregg and Farley, 2006 |

| | | | | | | | | | | | | |
|-----------------------------|-----|-----------------|---|--|--|--|--|--|--|---|--|--|
| VIII) Isidis deposits | 0.7 | 0.03 to 0.26 | X | | | | | | | X | Late Hesperia n to Early Amazon ian | Ghent et al., 2012; Ivanov et al., 2012 |
|-----------------------------|-----|-----------------|---|--|--|--|--|--|--|---|--|--|

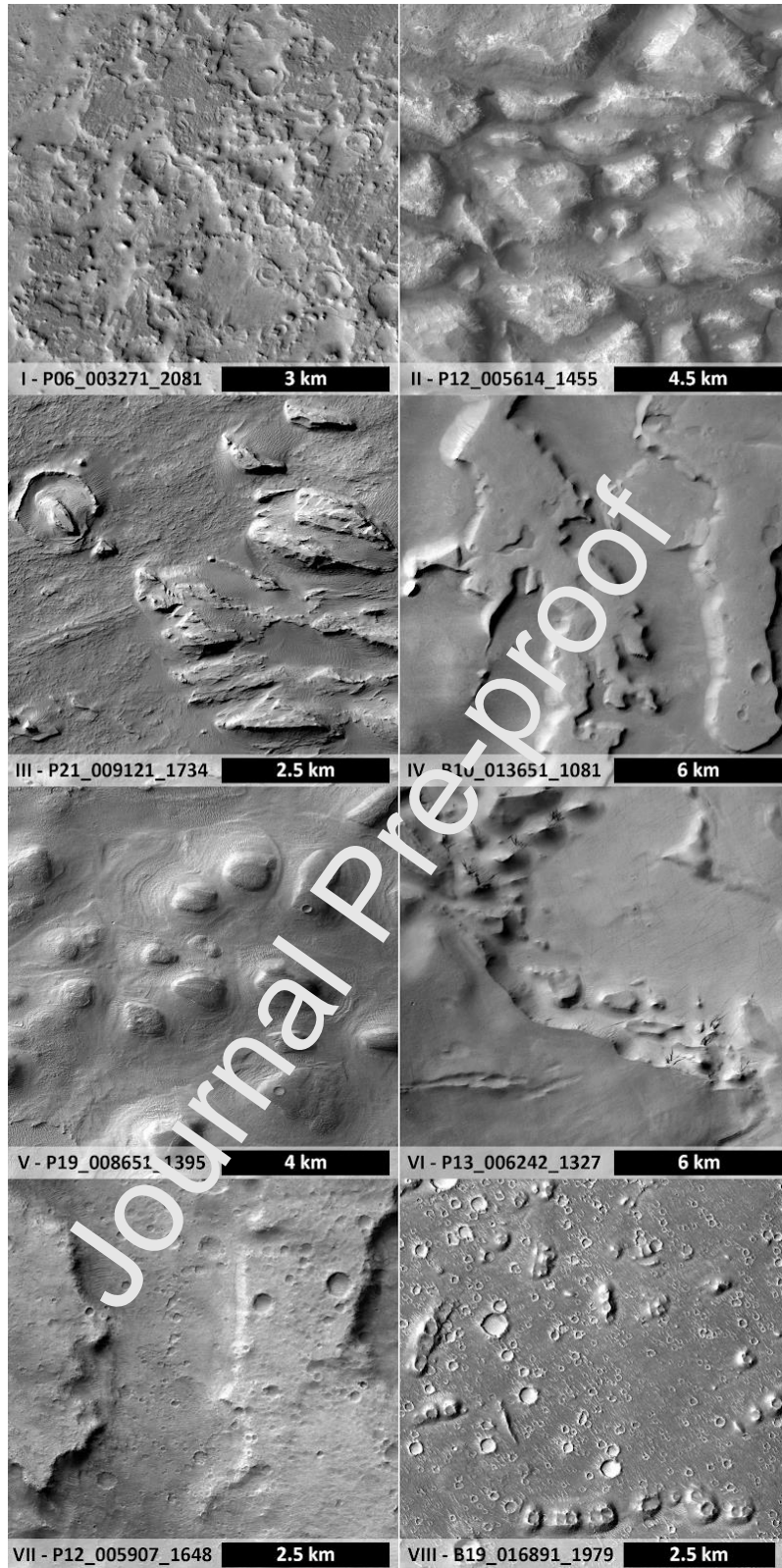


Fig. 8: Cropped CTX images showing example areas of the eight major potential pyroclastic deposits discussed in the Section 3.2. The areas were selected to illustrate morphologic characteristics discussed in the text. Roman numerals correspond to the numbering in Table 1. Black crosses in Fig. 1 indicate image locations.

3.2.1. I - Arabia deposits

At ~5 million km² across a diverse landscape, the proposed Arabia deposits are the most extensive, but also the morphologically most incoherent, of all potential pyroclastic deposits on Mars that have been described to date (e.g., Greeley and Guest, 1987; Tanaka, 2000; Hynek et al., 2003). Based on their apparent friability, horizontal layering, and their mantle-like draping over the landscape (Fig. 8), interpretations as remnants of pyroclastic tuff deposits were proposed for the kilometre-sized knobs and irregularly-shaped mesas in northeastern Arabia Terra (Moore, 1990; Fassett and Head, 2007). Tanaka (2000) proposed to also include mantling deposits mapped by Schultz and Lutz (1988) as well as Edgett and Parker (1997), and extended the deposits throughout northern Arabia Terra, suggesting a ~1 km thick unit sourced by airfall stretching across an elevation spectrum from ~3 km below the datum to ~1 km above it. Although Kerber et al. (2012) could not reproduce airfall dispersal patterns correlated with the Arabia deposits using Global Circulation models, the concept of such widespread and thick deposits emplaced by volcanic airfall was bolstered by the interpretation of five, up to ~55 km wide depressions in northern Arabia Terra as calderas (Fig. 1, letters Q-U) and, thus, potential source vents (Michalski and Bleacher, 2013). Nevertheless, no comprehensive stratigraphic, hyperspectral and geomorphologic studies have been conducted of the Arabia deposits so far and it has not been conclusively demonstrated that they actually represent a stratigraphic unit of singular origin.

3.2.2. II - Electris deposits

The Electris deposits are estimated to cover at least ~1.8 million km² and are comprised of chaotic terrains, i.e., knobs and irregularly-shaped mesas (Fig. 8), including Ariadnes and Simois Colles, as well as Atlantis, Caralis, and Gorgonum Chaoses, which are embayed by the surrounding Hesperian ridged plains and all occur within a closed highland basin called Eridania (e.g., Wendt et al., 2012; Adeli et al., 2015; Michalski et al., 2017). Their interpretation as a single unit of pyroclastic deposits remains debated. Several studies have detected serpentines, carbonates, and other key minerals using hyperspectral data, which suggests that the up to ~400 m high knobs are the remnants of lacustrine, possibly deep-sea hydrothermal, deposits formed in an Eridania lake that might also have

fed Ma'adim Vallis (e.g., Irwin et al., 2004; Adeli et al., 2015; Michalski et al., 2017). Nevertheless, small, potentially volcanic, structures in the area such as Electris Mons and the dike-like fissures of Sirenum Fossae have also motivated tentative interpretations as volcanic airfall deposits (Scott and Tanaka, 1986; Grant and Schulz, 1990; Tanaka, 2000). Furthermore, pyroclastic deposits are generally thought to disperse eastward (Kerber et al., 2012), hence a volcanic centre at Electris Mons, or other potential edifices at the western edge of the Electris deposits, could easily explain the latter's geographic extent. However, while the mineral species identified in the deposits can be explained by differential and/or hydrothermal alteration of volcanic material without a standing body of water (Brown et al., 2010; Ehlmann et al. 2010), many of the potential vent and edifices of the area, e.g., the Sirenum Fossae and most of the edifice-like mounds, postdate the deposits as they are superposed on the ridged plains embaying the Electris knobs (Wilson and Head, 2002; Brož et al., 2015b; Molina et al., 2014). In conclusion, we deem a pyroclastic origin of the Electris deposits to be improbable.

3.2.3. III - Medusae Fossae Formation

The ~2.5 million km² of Medusae Fossae Formation (MFF) overlies both the northern lowlands and the transitional terrain across ~100° of longitude (nearly 6000 km) between the Tharsis and Elysium bulges along the dichotomy boundary and is the most investigated and uncontroversial potential pyroclastic deposit on Mars (Malin 1979; Tanaka 2000 and references therein; Bradley et al., 2002; Squyres et al., 2007; Mandt et al., 2008, 2009; Kerber et al., 2011; Zimbelman and Scheidt, 2012; Ojha and Lewis, 2018; Mougini-Mark and Zimbelmann, 2020). Discontinuous exposures of the MFF are found ~600 km south of the main deposits, suggesting that the original deposit could have been even more voluminous (Harrison et al. 2010). The MFF was recognized already in Viking data as a distinct morphologic and stratigraphic domain characterized by layering, draping over the underlying basement, a high abundance of yardangs (Fig. 8) and other erosional features produced by wind and mass wasting, as well as the retention of steep slopes, all of which are indicative of its friability and some degree of lithification (e.g., Scott and Tanaka, 1982, 1986; Schultz and Lutz, 1988). Topography, gravity, and radar data show that MFF materials range in thickness from tens of meters in western exposures to more than 3 km in eastern exposures, and that it superposes units conformable

with both lowland plains and cratered highlands (Bradley et al., 2002; Zimbelman et al., 2015; Orosei et al., 2017; Ojha and Lewis, 2018). Based on its stratigraphic relations with neighbouring units, the bulk of the MFF was probably emplaced during the Hesperian, while Amazonian model ages from crater size-frequency distributions are most likely attributable to resurfacing by intense deflation and eolian redeposition (Scott and Tanaka, 1986; Werner, 2009; Kerber and Head, 2010; Kerber et al., 2011; El Maarry et al., 2012). While various origins for the MFF, ranging from eolian, paleopolar, and marine sediments to hyaloclastites, i.e., rafted pumice, have been postulated (e.g., Scott and Tanaka, 1986; Schultz and Lutz, 1988; Parker, 1991; Mouginis-Mark and Zimbelmann, 2020), this review focuses on the most robust body of accumulated evidence, which attributes these deposits to explosive volcanism.

In contrast to the morphologic analyses mentioned above, compositional analyses via spectral observations of the MFF are compromised by the ubiquitous dust cover in the region, which shows the typical iron oxide bands in spectral data, but lacks any potentially diagnostic signatures expected in pyroclastic deposits, e.g., feldspars, pyroxenes, or olivines (e.g., Bandfield, 2002; Ruff and Christensen, 2002, 2007; Ody et al., 2012). Nevertheless, other datasets have given information on the composition of the MFF and bolstered its interpretation as volcanic material, possibly welded tuff. The positive magnetic anomaly that is spatially correlated with Lucus Planum, a part of the western MFF near Apollinaris Patera, is indicative of oxidized, Fe-rich basaltic material consistent with a generally volcanic origin (Hood et al., 2010). Moreover, radar-derived bulk dielectric properties of the same area are more consistent with porous pyroclastic materials (Orosei et al., 2017), rather than the alternate explanation of ice-rich material. Radar observations of the eastern parts of the MFF have found them to be very absorbing (“stealth terrain”), which is indicative of pyroclastic deposits (Edgett et al., 1997). Gamma ray spectroscopy has revealed the general region centered around the MFF to have the strongest sulphur and chlorine signatures on Mars (consistent with 2.40–2.88 wt% S and 0.57–0.74 wt% Cl), which can be attributed to volatile phases in pyroclastic materials, although the quality of the spatial correlation is uncertain given the limited data resolution of ~550 km/px (Ojha and Lewis, 2018; Ojha et al., 2018). For Eumenides Dorsum in the eastern MFF, gamma ray spectroscopy also indicates

the material down to a depth of 80 cm to have a higher Si-content than the overlying dust (Peploswski, 2018), which is also consistent with volcanic materials. Lastly, gravity and topography data analyses using standard loading models found a bulk density of the MFF of $\sim 1800 \text{ kg/m}^3$, which is consistent with relatively porous rock, e.g., weakly to moderately compressed pyroclastic deposits (Ojha and Lewis, 2018). Due to its nature as a sulphur- and chlorine-enriched, friable, and fine-grained deposit, the MFF was also proposed to be a main source of Mars' atmospheric dust inventory and thus a significant factor modulating the current climate (Ojha and Lewis, 2018; Ojha et al., 2018).

Based on modelled convective plume dispersal, several extant volcanic vents have been investigated as potential sources for the welded tuff-like deposits of the MFF, including Olympus, Ascraeus, Pavonis, Arsia, and Alba Montes (Fig. 1, letters K-O), as well as Syrtis Major and Apollinaris Patera (Fig. 1, letters E and J, respectively), while only dispersal sourced from the latter can reproduce the MFF's current extent (Kerber et al., 2012). This would place the majority of the MFF far from its potential source vents, thus rendering welding by high emplacement temperatures unlikely. Nevertheless, as was already mentioned at the beginning of this chapter, compaction from pressure in sufficiently thick stacks can also cause welding (e.g., Ragan and Sheridan, 1972; Wilson and Hildreth, 2003; Quane et al., 2005) and might thereby explain the MFF's layered appearance and steep erosional patterns.

3.2.4. IV - Dorsa Argentea Formation

The 1.35 million km^2 and on average $\sim 500 \text{ m}$ thick Dorsa Argentea Formation (DAF, or "South Polar Pitted Terrain") has been dated to the Late Noachian to mid-Hesperian and has been noted for its esker-like ridges and potential hydrovolcanic landforms (e.g., Head and Pratt, 2001; Dickson and Head, 2006; Kress et al., 2010; Kress and Head, 2014). A composition that is at least partially pyroclastic in origin, e.g., as a thick ash blanket covering glacial deposits, was investigated by Kerber et al. (2012), whose deposition models predict that circum-South Pole fallout from convective pyroclastic plumes sourced by the Malea Planum paterae (Fig. 1, letters A-D) would correlate well with today's DAF extent. Alternative or additional potential source vents might be the numerous

isolated buttes of the Sisyphi Montes in the eastern parts of the DAF for which a formation as tuyas, i.e., sub- and englacial volcanoes, has been proposed (Ghatan and Head, 2002; Ackiss et al., 2018, 2019). Furthermore, recent analyses refined previous radar assessments (Plaut et al., 2007; Tanaka et al., 2014b), noting that consolidated basalt or dry, porous deposits, i.e., potentially of pyroclastic origin, would better explain the low dielectric constant and relatively high loss tangent values of the DAF than ice-bearing material (Whitten et al., 2020). Meter-scale blocks as far north as 62°S and dark material eroding out of the flanks of the DAF (Fig. 9) indicate that it contains non-volatile components, possibly including mafic fines. In some locations (Fig. 9B), these blocks appear very angular and are elongate, which is indicative of angular jointing, a common structural characteristic of volcanic deposits including pyroclastics (e.g., Wright et al., 2011; Lim et al., 2015). The various isolated occurrences of the DAF in separate topographic sinks (e.g., Tanaka et al., 2014b; Kress and Head, 2015) could be explained either by a once continuous unit that downwasted due to volatile-loss, e.g., after deposition as mostly glacial moraines (Scahill et al., 2018), and/or by volcanically-sourced airfall deposition. Although esker-like ridges indicate that glacial modification and deposition occurred across the DAF (e.g., Butcher et al., 2016), outcrops of dark fines (Fig. 9) indicate that ashfall might have contributed to its depositional stack, which could include significant amounts of pyroclastics from explosive volcanism pre-dating the proposed glacial activity in the region.

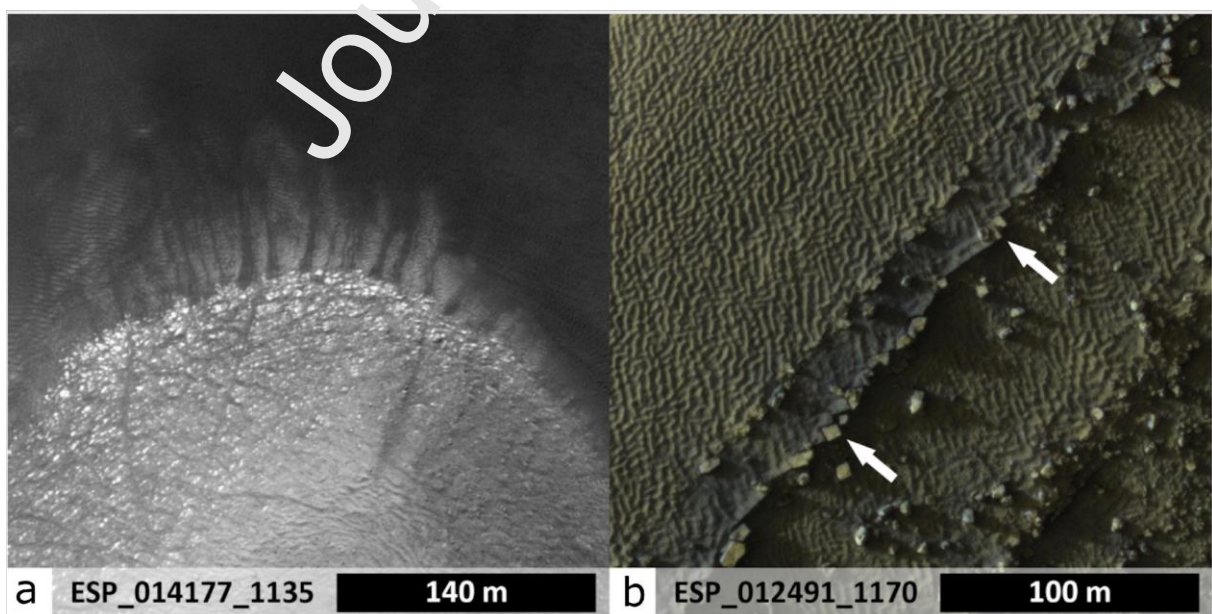


Fig. 9: HiRISE observations of escarpments bounding outlying occurrences of the Dorsa Argentea

Formation (white asterisks in Fig. 1). (a) shows an escarpment wall, where a layer of fine-grained, dark material crops out below more blocky, bright, and apparently jointed material. (b) shows layers of meter-scale blocks. White arrows point to angularly jointed blocks.

3.2.5. V - Hellas deposits

The up to ~1 km thick hummocky, Early Hesperian deposits (Fig. 8) that cover the central ~0.56 million km² of the Hellas basin floor might have been more widespread based on surrounding erosional remnants, possibly once covering the entirety of Hellas Planitia and thus totalling a volume of over a million km³ (Tanaka and Leonard, 1995; Leonard and Tanaka, 2001; Bernhardt et al., 2016). Extensive yardang fields, a high abundance of mass wasting features on shallow to moderate slopes, and a ubiquitous low thermal inertia indicate the deposits to be fine-grained and relatively weakly consolidated (Bernhardt et al., 2016). These observations, in addition to spectral signatures of pyroxenes and hydrous minerals (Williams et al., 2010; Carter et al., 2013), the occurrence of layered knobs, as well as the relative vicinity to potential source vents showing similar, i.e., Early Hesperian, surface model ages (Peneus, Amphitrites, and Hadriaca Paterae; Fig. 1; letters C-F), motivated interpretations as partially altered volcanic materials (Tanaka and Leonard, 1995; Tanaka et al., 2002; Williams et al., 2009; Bernhardt et al., 2016). These might have been (glacio-)fluvially transported from adjacent Malea or Hesperia Plana and/or emplaced directly by pyroclastic flows and/or ashfall (Tanaka et al., 2002; Williams et al., 2009; Bernhardt et al., 2016).

3.2.6. VI - Argyre deposits

The 0.34 million km² inner Argyre basin floor, for which Early to Late Hesperian model ages have been derived, is comprised of up to ~2 km thick, at least partially layered deposits (e.g., Hiesinger and Head, 2002; Bernhardt et al., 2013; Dohm et al., 2015). Due to the etched, yardang-like, i.e., possibly friable, appearance of its northern parts (Fig. 8), and the presence of Argyre Mons (Fig. 1, letter P) on the western basin floor as a potential vent structure resulting from explosive volcanism, an at least partial pyroclastic origin of the basin-fill has been tentatively suggested (Scott and Tanaka, 1986; Tanaka et al., 2014b; Williams et al., 2017). Alternative interpretations include glacial, glacio-

fluvial, and/or lacustrine deposits (e.g., Hiesinger and Head, 2002; Bernhardt et al., 2013; Dohm et al., 2015), and further analyses are needed to fully assess a potential pyroclastic scenario.

3.2.7. VII - Tyrrhena Patera deposits

The plains north of Tyrrhena Patera (Fig. 1, letter G) are covered with notched mesas having rough surfaces, i.e., likely remnants of friable deposits (Fig. 8; Gregg et al., 1998; Greeley and Crown, 1990). Gregg and Farley (2006) found that these deposits cover up to 0.32 million km² and extend at least ~600 km to the north, where they are bounded by up to ~150 m high, often terraced, scarps. Based on the geographic extent and apparent thickness, an emplacement by pyroclastic flows was favoured over ashfall deposition, suggesting that a thin atmosphere, lower gravity, and volatile-rich substrate might have substantially increased the flow runout distance in comparison to that in a similar terrestrial setting (Gregg and Farley, 2006).

3.2.8. VIII - Isidis deposits

The 0.7 million km² Isidis basin floor, for which Late Hesperian to Early Amazonian model ages have been derived, hosts thousands of subkilometre-sized pitted cones forming the largest occurrence of “thumbprint terrain” (Fig. 8) on Mars (e.g., Lockwood et al., 1992; Ghent et al., 2012, Ivanov et al., 2012). Based on buried craters, the underlying unit has been estimated to be up to ~400 m thick and various formation scenarios, usually involving either lacustrine, glacial, or volcanic activity, have been proposed (e.g., Kargel and Strom, 1992; Ghent et al., 2012, Ivanov et al., 2012; Guidat et al., 2015). Partially motivated by its proximity to the potential source vents of Syrtis Major (Fig. 1, letter E) as well as by the fine-grained nature of the basin-fill that impact craters indicate to be interrupted only by some shallow, dark layers, Ghent et al. (2012) proposed a scenario in which the pitted cones formed as degassing structures of pyroclastic deposits. Nevertheless, other formation scenarios for the Isidis basin fill related to wet-based glacial activity or mud volcanism have also been proposed (e.g., Ivanov et al., 2012; Guidat et al., 2015), and its interpretation as pyroclastics remains debatable.

3.2.9. Other deposits potentially sourced by explosive volcanism

In addition to the largest potential pyroclastic deposits on Mars listed in Table 1 and described in the previous sections, other areas have been interpreted as pyroclastic in nature, albeit involving much smaller or unknown volumes.

The western flank of Hecates Tholus (Fig. 1, letter H) appears very smooth and its foot is notched by a 10 km caldera-like embayment (Mouginis-Mark et al., 1982; Hauber et al., 2005). The flank and its embayment lack craters larger than 40 and 200 m respectively, which is why both areas were interpreted to have been resurfaced/formed in the Late Amazonian. Based on the geomorphology of the flank deposits and numerical models, Mouginis-Mark et al. (1982) suggested airfall from a ~70 km high plume by an explosive eruption of Hecates Tholus to have emplaced ~100 m, i.e., potentially 60 km³, of pyroclasts. It might be possible that, instead of the volcano's summit caldera, the flank caldera described by Hauber et al. (2005) sourced this eruption.

On and around the Tharsis bulge, several constructs and deposits probably originating from explosive volcanic activity have been reported. Kilometre-sized, relatively pristine cones on the summit of Pavonis Mons as well as on the plains of western Tharsis are among the most unambiguous landforms indicating Amazonian explosive volcanism in the area, albeit of limited scale (Zimbelman and Edgett, 1992; Brož and Hauber, 2012). Large-scale, probably Hesperian explosive eruptions in the Tharsis province were inferred from theoretical considerations as well as the extensive pyroclast-like deposits and their alteration products within and around the Tharsis rise, including layered deposits within Valles Marineris and Hebes Chasma (e.g., Head and Wilson, 1998a,b; Mouginis-Mark, 2002; Hynek et al., 2003; Squyres et al., 2007; Schmidt et al., 2018). Although the deposits of the eastern MFF and in southern Arabia Terra have also been tentatively associated with explosive activity on Tharsis, various hypotheses about their sources have been brought forward (see Sections 3.2.1. and 3.2.3.). Furthermore, two inactive transverse dune fields likely sourced by local fines, i.e., potentially pyroclastic materials, superpose Late Amazonian terrain in the south-western part of the Tharsis province and have been considered as evidence for recent explosive volcanism in the region (Edgett,

1997). However, the study was limited to Viking data and alternative hypotheses explaining the dunes' location and apparent age, e.g., earlier dune migration from older surfaces and/or atmospheric entrainment of the volcanically-sourced eolian sediments, were not investigated.

In the Nili Fossae region northeast and northwest of Syrtis Major Planum and Isidis Planitia, respectively, outcrops of bright-toned, fine-grained but lithified, layered material has been found (e.g., Hoefen et al., 2003; Kremer et al., 2019). Hyperspectral analyses revealed not only strong olivine-signatures, but also several potential olivine alteration products, motivating an interpretation as ashfall deposits derived from primitive mantle melts that likely erupted on Syrtis Major Planum (Hoefen et al., 2003; Ehlmann and Mustard, 2012; Edwards and Ehlmann, 2015; Kremer et al., 2019). Due to the geographic vicinity, it is possible these deposits are genetically related to the larger Arabia deposits (see section 3.2.1.).

4. Ground truth

Evidence for explosive volcanic activity on Mars has also come from ground observations obtained by rovers. The most notable discoveries were by the Mars Exploration Rover (MER) Spirit, which explored the floor of ~166 km wide Gusev Crater, in which evidence for possible pyroclastic eruptions (Squyres et al., 2006; Picot et al., 2006; 2010; Farrand et al., 2006; Herkenhoff et al., 2006; Schmidt et al., 2006; Ennis et al., 2007; for review see Wilson and Head, 2007; Rice et al., 2016) as well as hydrothermal activity (e.g., Squyres et al., 2008; Ruff et al., 2020 and references therein) was detected from orbit. The in-situ evidence for pyroclastic deposits comes from Spirit's investigation of the Columbia Hills, which rise up to ~100 m above the surrounding plains, with a lateral extent of ~6 km (Fig. 10a). Here, bright deposits, which can be seen from orbit as having a roughly circular outline in plan view, show fine-scale layering and cross-bedding at in-situ scale (Squyres et al., 2007), similar to what is observed in terrestrial volcanic surge deposits formed by hydrovolcanic explosions. One of these areas containing bright deposits is called "Home Plate" (Fig. 10b), an ~90 m wide approximately circular-shaped low plateau about 2–4 m high, in which the presence of inward-dipping layered rocks proposed to be formed by lapilli and volcanic ash later reworked by eolian processes

was found (Rice et al., 2006; Wilson and Head, 2007 and references therein). It was proposed that this material resulted from hydrovolcanic activity that occurred within Gusev crater leading to the formation of a tuff ring or maar, which was later eroded (e.g., Rice et al., 2006; Ennis et al., 2007; Squyres et al., 2007), or that it resulted from materials transported from the Apollinaris Patera, which lies ~280 km north of the Gusev crater rim (Wilson and Head, 2007). A candidate volcanic source vent dubbed “Goddard” was discovered ~170 m south of Home Plate (Fig. 10b; Rice et al., 2010), but not explored by rover because the fine particulate regolith precluded further driving. Also olivine-rich pyroclastic rocks were observed in some outcrops within Gusev crater supporting the theory that volcanism was active inside the crater (Ruff et al., 2014). The observation of a possible volcanic bomb-sag (Fig. 10c) within the Home Plate exposure suggests that the deposits originate from hydrovolcanic activity within the Gusev crater itself (Lewis et al., 2008) as a volcanic bomb would not be expected more than 350 m from its source (Manga et al., 2012).

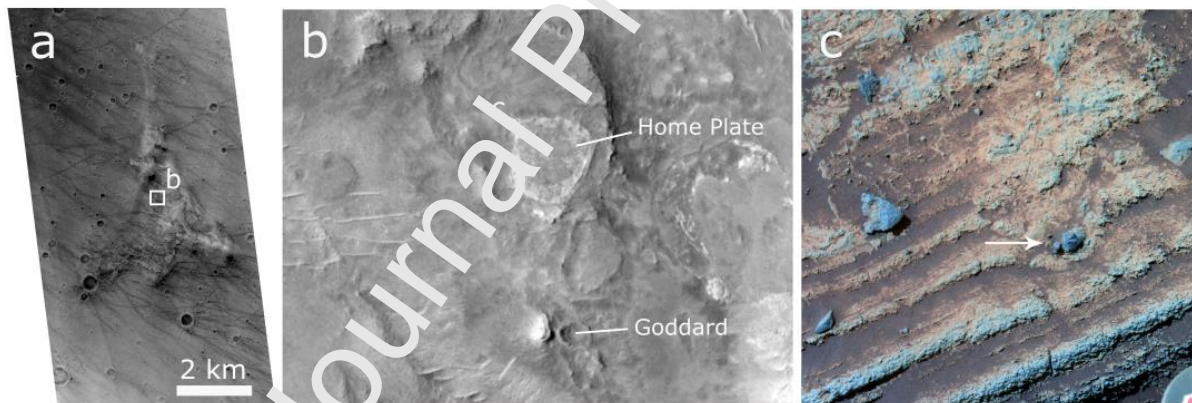


Figure 10: Evidence of explosive volcanism as observed within Gusev crater. Panel (a) shows Columbia Hills on the floor Gusev crater investigated in-situ by the Spirit rover, in which bright semi-circular outcrops were observed. One of these areas containing bright deposits is a Home Plate (b), from which a putative volcanic vent named Goddard was discovered. Panels (a-b) are based on HiRISE image PSP_005456_1650 and north is up. Panel (c) shows a putative volcanic bomb discovered within the Home Plate (for exact position see Lewis et al., 2008), which landed into the strata photographed by the Spirit rover.

Recent studies additionally suggest that the rover-based instruments on board Mars Science Laboratory (MSL) rover Curiosity may be used to distinguish whether altered tuff samples, if detected, were formed by hydro- or glaciovolcanic activity (Farrand et al., 2018), which is giving a prospect to detect direct evidence of hydrovolcanism on Mars in the future.

5. Effects of the environment on explosive volcanism

The specific planetary environment plays a crucial role in determining the type of volcanic activity and hence on resulting shapes of volcanic landforms. Mars is a planet that has a lower gravity and current atmospheric pressure in comparison to Earth, 3.71 m/s^2 vs. 9.81 m/s^2 and $\sim 600\text{--}1000 \text{ Pa}$ vs. $\sim 10^5 \text{ Pa}$, respectively. It was theorized that the reduced atmospheric pressure means that even magmas with very low amounts of gases ($\sim 0.03 \text{ wt\%}$) should be fragmented on Mars, at depths of at least a few meters. The kinetic energy produced by expanding bubbles would be then sufficient to eject pyroclasts from the vent at a speed of at least a few tens of meters per second (Wilson, 1980). Consequently, even magmas, which do not have sufficient dissolved volcanic gases to trigger explosive eruptions on Earth should cause magma fragmentation on Mars and therefore explosive eruptions should be more common on Mars. At the same time, it has been proposed by many researchers that variations in these two parameters strongly affect the entire course of volcanic eruptions and the shape of the resulting volcanic features, which may significantly vary from their terrestrial counterparts (e.g., Wood, 1979b; Dehn and Sheridan, 1990; Wilson and Head, 1994; Glaze and Bologna 2002; Wilson and Head, 2004; Parfitt and Wilson, 2008; Brož et al., 2014). For example, the lower gravity affects the depth at which magma reservoirs form, the crystal settling processes within them and the motion of the melt through the surrounding rocks, and hence allows larger magmatic bodies to be formed in turn affecting the effusion rates of volcanic eruptions as well as the amount of material that can be transported to the surface (for details see Wilson and Head, 1994; Heap et al., 2017). Below, this section contains a summary of how the different environmental properties affect the courses of different types of explosive eruptions.

5.1. Effects on low energetic explosive eruptions

From the point of view of low energetic explosive eruptions, namely Hawaiian and Strombolian eruptions which are causing ejections of particles by ballistic pathways, the lower gravity increases the distance to which the ejected material can travel because particles are subjected to less gravitational force (e.g. Wood, 1979b; Dehn and Sheridan, 1990; Brož et al. 2014). At the same time, the lower atmospheric pressure affects the distance to which the ballistically ejected volcanic material can travel. This is because atmospheric drag is linearly proportional to the air density, which is significantly reduced on Mars, hence allowing ejected particles to travel further (Wood, 1979b; Dehn and Sheridan, 1990; Wilson and Head, 1994; Brož et al. 2014, 2015a). It has also been pointed out that the lower atmospheric pressure leads to more gas expansion causing more energy to be released and therefore the initial ejection velocities of pyroclasts to be higher on Mars than on Earth (Wilson and Head, 1994).

For Strombolian eruptions, it is estimated that on Mars velocities could reach ~2 times typical velocities on Earth, corresponding to a mean particle velocity of ~92 m/s (Brož et al. 2014) so the ejected material can fly further. Also, the dispersion of ejection angles of pyroclasts should be wider than on Earth (Glaze and Baloga, 2000; Wilson and Head, 2007), hence reducing the height and increasing the diameter of resulting volcanic features in comparison with their terrestrial analogues when the same amount of erupted material is considered (Brož et al., 2014, 2015a). In addition, gas nucleation and fragmentation for volatile-bearing magmas is inferred to occur at greater depths on Mars than on Earth and the fragmentation should be far more energetic and hence more effective than on a planet with a denser atmosphere (Wilson and Head, 1994). Brož et al. (2014, 2015a) showed, based on numerical modelling, that the ejected particles forming putative martian scoria cones should be around ~20 times finer than is typical on Earth, corresponding to a mean particle size of ~2 mm on Mars.

The reduced atmospheric pressure and gravity on Mars may also affect the distance to which ballistically emplaced pyroclasts can travel and hence change the shape of kilometre-sized volcanic features by making their slopes more gentle (Brož et al., 2015a; 2018). This can be illustrated using the example of scoria cones. On Earth, flanks of young uneroded scoria cones reach 30° in slope,

which corresponds to the angle of repose for scoria material. Once this value is reached additional scoria is redistributed by avalanching (Riedel et al., 2003). On the other hand, on Mars scoria cones would not easily reach the angle of repose; regardless of whether this angle is gravity dependent or not (Kleinhans et al., 2011; Atwood-Stone and McEwen, 2013), because ejected material should be spread over a much larger area than on Earth. As a result, typical erupted volumes are not large enough for the flank slopes to attain the angle of repose on Mars, in contrast to Earth (Brož et al., 2014, 2015a). Therefore, the redistribution of particles by avalanching should not play an important role in their formation and martian scoria cones should be formed predominantly by ballistic emplacement. As the result, by taking into account the environmental differences between Earth and Mars, the shape of the cones on Mars can be used to inform the type of volcanic eruption, even when this occurred hundreds of millions years ago.

5.2. Effects on high energetic Plinian eruptions

The fact that more energy is released during gas expulsion also suggests that basaltic Plinian eruptions should be very common on Mars (Wilson and Head, 1994), even though they are rare for terrestrial basalts. It was proposed that during such eruptions, ejected particles, which may be up to 100 times smaller than on Earth, should have higher velocities and therefore it was theorized that they could reach significantly higher than those on Earth could due to the reduced gravity and atmospheric drag (Wilson and Head, 1994; Wilson and Head, 2007). Theoretical predictions assumed that this might be up to five times higher on Mars than their terrestrial counterparts resulting from similar eruption rates, hence exceeding 100 kilometres, and that the convective plume should collapse twice higher than terrestrial plumes (Wilson and Head, 1994). However, as the heights of convective plumes on Mars are limited by the amount of air that is available for them to entrain in the thin martian atmosphere, it was later proposed by Glaze and Bologna (2002) that they would not in the end ascend as high as on Earth. Glaze and Bologna (2002) showed that martian convective plumes would not be able to advance above 10-20 km due to the insufficient amount of gases available for entrainment in the martian atmosphere and as the result the distribution of large volumes of pyroclasts over wide

areas should be precluded. In addition, they proposed that the low density of the atmosphere would lead to the significant lateral expansion of the plume. This should cause instabilities of martian convective plumes and hence a frequent formation of pyroclastic density currents (e.g., Crown and Greeley, 1993; Glaze and Bologa, 2002; Glaze et al., 2011); dense mobile mixtures of hot gases and unsorted pyroclasts. However, Glaze et al. (2011) showed that when the geometry of the vent from which the pyroclasts are ejected is changed from circular to linear, the resulting convective plume might have more stability. This is because a higher flux of material would be available and as a result, a pyroclastic fountain capable of injecting a large quantity of pyroclasts and volatiles into the atmosphere might develop.

Later Wilson and Head (2007) showed that even though the martian atmosphere is thin, pyroclastic plumes might still reach tens of kilometres high above the surface. However, they would differ from those on Earth. This is because the current martian atmosphere is so thin that a Plinian eruption could not entrain enough of the surrounding atmosphere to form a convecting pyroclastic plume. Instead the eruption would form an umbrella-like plume, which is a well-known explosive feature on the jovian moon Io (Wilson and Head, 2007; Davies, 2007). Ejected particles in such umbrella-shaped pyroclastic plumes whose sizes are predicted to be in the range of tens of microns to few millimetres, would follow nearly ballistic trajectories until they re-entered the thicker part of the atmosphere and at that point particle movement would be affected by the wind (Wilson and Head, 2007). This means that pyroclasts could be distributed over a wide area (Horth and Weitz, 2001; Hynek et al., 2003; Glaze and Baloga, 2000), however, similarly to on Earth, the distance to which ejected particles can travel should be based on their size. The largest millimetre-sized clasts should be deposited only tens of kilometres from the vent, but the micron-sized grains may be transported tens of thousands of kilometres; in all considered cases further than on Earth (for details see Wilson and Head, 2007). Because of this, explosive eruptions on Mars should form much thinner pyroclastic deposits than are known on Earth (Kerber et al. 2011, 2012, 2013). As such, this may explain the paucity of Plinian eruption deposits recognized so far on Mars (Wilson, 2009). In addition, because the convective plume should contain water vapour, which can condense on the ejected rocky grains,

accretionally lapilli could also be formed causing an increase in the sizes of pyroclasts falling through the atmosphere to the surface (Wilson and Head, 2007).

An umbrella-like pyroclastic plume can be also formed by particularly strong eruptions as the upward velocity of the cloud would be large enough to “punch” the pyroclastic material through the lower, thicker part of the atmosphere (Wilson and Head, 2007; Kerber et al., 2011). Therefore, a transition between these two different transport mechanisms can happen on Mars during the course of the eruption. And finally, as the martian atmospheric pressure varies significantly, which is caused by the elevation difference around of 30 kilometres between the lowest and highest point of Mars it has been predicted that this may have a profound effect on the course of a volcanic eruption depending on where the volcanic activity occurs (Wilson and Head, 2007). For example, if a Plinian eruption occurred at the bottom of Hellas basin or in the northern lowlands, pyroclastic clouds should be formed because the atmosphere would be dense enough. But if a similar volcanic eruption occurred at the summit of Olympus Mons at an elevation of ~20 km, instead an umbrella-like plume would emerge.

5.3. Effects on pyroclastic density currents

It is also predicted that pyroclastic density currents, which are dense mobile mixtures of hot gases and unsorted pyroclasts, should be affected by environmental properties. Pyroclastic density currents have been historically divided into two main types, pyroclastic flows and pyroclastic surges, with two different primary transport mechanisms, but there is increasing evidence that these end members often transition between each other (see e.g. Dufek [2016] for an overview). While the term pyroclastic flows has typically been used to mark concentrated currents with high particle concentrations, the term pyroclastic surges has typically been used to describe dilute and turbulent currents producing deposits with visible bed forms (Dufek, 2016).

Historically, it was proposed that dilute pyroclastic density currents should move faster in the martian atmosphere than on Earth and may be able to overcome topographic obstacles up to 5 km high on Mars (Wilson and Head, 2004) and travel tens of kilometres far away from the source area (Wilson

and Head, 1994). This is because the lower gravity causes weaker particle frictional interactions and the initial velocities might also be higher than on Earth. However, this was later questioned by the work of Brand and Clark (2011) and Clark et al. (2013). Brand and Clarke (2011) showed that, when sedimentation of particles and entrainment of the atmosphere are considered in the model of propagation of dilute pyroclastic density currents, the resulting runout distances of these flows should be six times shorter than previously thought. This corresponds to three times that of equivalent pyroclastic currents on Earth (Clark et al., 2013). Brand and Clark (2011) also found that these lower sedimentation rates should result in density stratification over a greater vertical extent causing longer dune wavelengths on Mars than on Earth. In addition, it was also proposed that the long distances to which the pyroclastic material should travel on Mars would allow for significant cooling during transport. Thus, the deposits resulting from dilute pyroclastic density currents should be poorly consolidated, widely dispersed and composed of fine-grained material (Wilson and Head, 1994). On the other hand, in the case of concentrated pyroclastic density currents it was proposed that they could remain so hot that they would weld without pressure compaction even at large deposition distances, at ranges of hundreds of kilometres from their source (Crown and Greeley, 1993). However, it has to be noted that despite the progress in recent years (Brand and Clarke, 2011; Clark et al., 2013; Florez and Andrews, 2017) the behaviour of pyroclastic density currents on Mars is not fully understood and more work on this subject is warranted.

5.4. The role of evolving Martian atmospheric pressure

However, it is important to note that the atmospheric pressure of Mars has likely changed significantly during the history of the planet. It has been proposed that the early atmosphere on Mars may have had a pressure greater than that of present-day Earth, perhaps ~ 0.3 MPa, and that as much as 99% of the original nitrogen and carbon atmospheric inventory of Mars was lost before about 3.8 Ga, 0.9% has been lost since, and perhaps only 0.1% remains (Catling, 2009). The gases were lost from the atmosphere to space (e.g. Jakosky et al., 2018) due to the weak gravity together with the absence of magnetic field, which was operating prior to ~ 4 Ga ago or they were lost by sequestration into

sedimentary deposits/rocks (e.g. Jakosky, 2019). This would mean that the volcanic eruptions in the distant past of Mars likely occurred under very different circumstances than those taking place more recently, raising the possibility that the ways the pyroclasts were transported together with resulting shapes of volcanic edifices may have varied significantly over time. Therefore it would be misleading to investigate old volcanic features assuming present-day martian atmospheric conditions.

6. Effects of explosive volcanism on the climate of Mars

The injections of both pyroclasts and volatiles into the atmosphere by volcanic activity have been identified as crucial climate forcing mechanisms on Earth (e.g., Kirchner et al., 1999; Petersen et al., 2016; Burgess et al., 2017). Likewise, large volcanic eruptions, as deduced from deposits interpreted to be of such origin, have been invoked to force the ancient martian climate into a significantly warmer state than today (e.g., Sagan et al., 1973; Greeley, 1987; Craddock and Greeley, 2009). Therefore, a better understanding of the nature of past explosive volcanism on Mars is a key to assessing the planet's climate evolution, which includes the amount of water on the surface and in the atmosphere over time (e.g., Carr, 1996; Catling, 2009).

Depending on the assumed H_2O content in magma (e.g., 1-3 wt%) and the total volume of potential volcanic materials on Mars (e.g., $\sim 2 \times 10^7 \text{ km}^3$ by effusive and explosive eruptions), previous estimates for the amount of outgassed water range from few tens of meters to over one hundred metres of global equivalence layer (Greeley, 1987; Carr and Wänke, 1991; Carr and Head, 2015).

On Earth, pyroclastic deposits from explosive volcanism are usually linked with magmas of higher volatile content, usually $< 2 \text{ wt\%}$ mainly consisting of CO_2 and H_2O , but also sulphur-bearing species and H_2 amongst others (e.g., Carroll and Holloway, 1994; Wallace et al., 2015). On Mars, however, only $\sim 0.03 \text{ wt\%}$ of volatiles are needed to result in such eruptions (Wilson and Head, 1981 or see Section 5 for details). Nevertheless, approximately 0.5 to 2.8 wt% of volatiles within the magma have been suggested for the explosion of Hecates Tholus in order to form an eruption plume of sufficient height to deposit the mapped $50 \times 75 \text{ km}$ pyroclastic deposits via airfall (Mouginis-Mark et al., 1982). For the Amazonian volcanic deposits of the southern Elysium region, volatile contents of 1

wt% of CO₂ and H₂O have been proposed, which would imply $\sim 10^{16}$ kg (i.e., almost the entire mass of today's atmosphere) of each species as well as other compounds (10^{15} kg S, 10^{13} kg Cl, 10^{13} kg F) (Plescia, 1993). Combining comparatively low Icelandic CO₂ contents of ~ 0.05 wt% with models of thermal mantle evolution and dike propagation through a 200 km thick lithosphere, it was suggested that martian volcanism declined from a magma eruption rate of $0.17 \text{ km}^3/\text{yr}$ during the Hesperian to around $1 \times 10^{-4} \text{ km}^3/\text{yr}$ at present, which would imply that the CO₂ outgassing rate decreased from 8.8×10^7 to $6.6 \times 10^6 \text{ kg/yr}$ (O'Neil et al., 2007). Although more recent estimates, which take into account the mantle oxygen fugacity, limit the CO₂ content in martian magmas to ~ 0.1 wt%, volatile enrichment during ascent, e.g., H₂O from permafrost, could lead to significantly higher values (Hirschmann and Withers, 2008; Grott et al., 2011). Furthermore, as indicated by martian meteorites, the early martian mantle probably was more volatile-rich than today's, possibly ~ 1 wt% or more, followed by a depletion over time caused by a lack of recycling via plate tectonics (e.g., Johnson et al., 1991; Guest and Smrekar, 2007; Balta and McSween, 2013). Below, we use the currently estimated inventory of all proposed pyroclastic deposits on Mars (Fig. 1; Table 1) to update the estimated total outgassed volatile volumes on Mars, based on a minimum scenario of 0.03 wt% combined volatile content and a conservative Earth-like value of 2 wt% (e.g., Wallace et al., 2015).

Although pyroclastics can be porous, we assume that compaction throughout the majority of the depositional stacks in the extensive and relatively thick deposits listed in Table 1 results in an average, basalt-like density of 3000 kg/m^3 . This would imply ~ 2.55 to $2.98 \times 10^{19} \text{ kg}$ of pyroclastic material, which corresponds to ~ 12 to 14% of the entire martian inventory of extrusive volcanic material (i.e., including the wrinkle-ridged plains) as mapped by remote studies (Craddock and Greeley, 2009). Multiplied by 2 wt%, this results in a mass of $\sim 10^{15}$ to $\sim 10^{17} \text{ kg}$ of volatiles that might have been produced by explosive eruptions. According to Hirschmann and Withers (2008) the upper limit of 0.1 wt% for CO₂ in Martian magmas, this would result in $\sim 10^{16} \text{ kg}$ or $\sim 6\text{-}7 \text{ mbar CO}_2$. An H₂O content of 2 wt% on the other hand, would result in $\sim 10^{17} \text{ kg}$ or $\sim 2 \text{ m}$ global equivalence layer (GEL) of water. Via condensation and runoff, this might explain certain erosional landforms close to potential vents, but could have been only one of many components contributing to the $\sim 500 \text{ m}$ of accumulated

GEL water volume that was proposed to have cycled across the surface over time to form Mars' global inventory of channel networks (Carr, 1996).

The potentially outgassed CO_2 roughly equals the current mass of Mars' atmosphere. Nevertheless, applying a limit of 0.1 wt% on the currently known potential pyroclastic deposits seems to imply that CO_2 from explosive eruptions alone, even if they had occurred at the same time, was an insufficient source for significant climate change as models indicate pressures of several hundreds of millibars of CO_2 are needed for any appreciable temperature increase (e.g., Ramirez et al., 2014; Wordsworth, 2016). Furthermore, the greenhouse properties of sulfur-bearing species such as SO_2 and H_2S that might have been outgassed alongside the CO_2 were probably more than counteracted by the formation of sulfate aerosols causing a cooling effect on the atmosphere (Mischna et al., 2013; Halevy and Head, 2014; Kerber et al., 2015; Wordsworth, 2016). Despite this, the effect by collision-induced absorptions between CO_2 , $\text{H}_2/\text{H}_2\text{O}$ and CH_4 outgassed by volcanism might have caused significant warming (Ramirez et al., 2014; Wordsworth et al., 2016). Even assuming that H_2 was not outgassed directly (which would require a reducing mantle), but only came from photo-dissociation of outgassed H_2O , the above-mentioned 2 wt% H_2O would result in ~9 mbar of H_2 . This would be a significant contribution to the 10s of mbar partial H_2 pressure that models require to raise the average surface temperature above 273 K in a 1.5 bar atmosphere (Wordsworth et al., 2016). For an ideal endmember scenario, in which all the potential deposits listed in Table 1 are indeed pyroclastic in nature, we therefore submit that the CO_2 outgassing from explosive volcanism alone was probably not a significant climate-altering parameter by itself, but that its combination with outgassed $\text{H}_2/\text{H}_2\text{O}$ and CH_4 might have significantly contributed to atmospheric warming. However, even in this endmember scenario, the large time span over which the potential pyroclastic deposits of Mars were emplaced (crater-based model ages span ~1 Gyr from the Early Noachian to the Early Amazonian, see Table 1) would result in a very low average production rate of ~0.9 mbar H_2 per 100 million years. Nevertheless, if the largest potential pyroclastic deposits of Arabia Terra formed within a relatively short time span, its up to ~5 mbar of outgassed H_2 might have caused an appreciable effect.

Various studies have also highlighted the significant effects and implications of other volcanic gases and their isotopes on the martian atmosphere and surface throughout the planet's history (Smith et al., 2014; Craddock et al., 2009; Gilman et al., 2011). Via 1-dimensional photochemical reactions, volcanic outgassing can form sulfate, nitrate, and perchlorate salts in the atmosphere, which then accumulate in the regolith (Smith et al., 2014). Assuming Mars' volcanic flux throughout the Amazonian was 0.1% of Earth's current flux and an otherwise unchanged background atmosphere during that time, the amount of expected sulfates (1.0 to 1.7 wt%) and nitrogen (0.2 to 0.4 wt% in the form of pernitrate) in the surface is consistent with in-situ observations of soils from rovers and landers, while the perchlorate exceeds the expected value (Smith et al., 2014). Outgassed sulphur-species could also have combined with H_2O , CO_2 , and other compounds to form H_2SO_4 , HCl , or HF , which might have significantly contributed to the chemical alteration of the martian surface (Craddock et al., 2009). Furthermore, volcanic eruptions inject noble gases and their isotopes into the atmosphere, which are then partially lost to space at different rates via sputtering at high altitudes and can therefore be used to assess atmospheric loss over time (e.g., Hutchins and Jakosky, 1996; Slipski and Jakosky, 2016). The deviation of measured argon and neon isotope volume relations in today's martian atmosphere from the expected values indicates that the atmosphere once contained at least 10 times the volume of these species, hence implying substantial sources that cannot be explained by volcanic outgassing alone, but require processes like passive outgassing (Hutchins and Jakosky, 1996).

However, it has to be noted that we considered only the effect of gases released by magma degassing, but neglected those potentially released by hydrovolcanism. Released hydrovolcanic volatiles might increase the atmospheric pressure, but the extent and timing of hydrovolcanism is not well-constrained and beyond the scope of this review.

7. Conclusions

As has been laid out in this review, a broad spectrum of observations indicates that signs of explosive volcanism are present on the surface of Mars. The most spectacular are hundreds of kilometres wide paterae on the summits of edifices, which suggest that explosive volcanism must have

been commonplace in the early history of Mars (e.g., Williams et al., 2009; Michalski and Bleacher, 2013; Bandfield et al., 2013; Huang and Xiao, 2014). On the other hand, the presence of tholi, fresh-looking scoria cones together with several types of hydrovolcanic edifices (including tuff rings, tuff cones and rootless cones) and widespread pyroclastic deposits of various ages reveal that explosive volcanism has been present on Mars even more recently, most likely over the entire history of the planet. However, the sparsity of Amazonian explosive edifices compared with effusive edifices implies that explosive volcanism was unevenly distributed in time and space. The more common occurrence of ancient martian explosive constructs hints that magma ascending to the surface of Mars in the Noachian had a higher volatile content than Amazonian magmas and/or that the interaction between magma/lava and water was much more frequent in the Noachian than later in Mars' history. This points to a transition from early frequent explosive volcanism to younger, dominantly effusive volcanism (e.g., Greeley and Crown, 1990; Crown and Greeley, 1993; Robinson et al., 1993; Robbins et al., 2011; Kremer et al., 2019). It remains unclear, even after decades of martian exploration, why such a transition in volcanic styles occurred. It is possible that the martian mantle may have contained a large quantity of volatiles in the past, but got drier as volatiles were released by volcanism and never recycled due to the absence of plate tectonics (Balta and McSween, 2013).

The decline in explosive activity with time had a direct effect on the evolution of the martian atmosphere as less volcanic gases were released. This limited the ability of the planet to replace the gases lost from the atmosphere to space (e.g. Jakosky et al., 2018) or those buried underneath the martian surface (e.g. Jakosky, 2019). The declining input of volcanic gases into the atmosphere affected the climate at a global scale as both surface atmospheric pressure and temperature have decreased into the Amazonian (Kite, 2019). The present-day climate precludes the prolonged presence of liquid water on the surface of Mars, but a wide range of geological and mineralogical evidence shows that this was not the case in the Noachian period (Kite, 2019; Lasue et al., 2019) when Mars is thought to have been much wetter than today. It remains a matter of debate as to how the wetter climate was achieved (see Wordsworth, 2016 for details), however, there is a consensus that input of gases from explosive volcanism was a key factor. Therefore, the evolution of the entire planet has been affected by the decline of explosive volcanism.

As shown by this review, the presence of explosive volcanism cannot be confirmed unequivocally by remote sensing data alone as alternative explanations often arise. This limits our ability to understand fully the role that explosive volcanism has played in the evolution of Mars. We argue that to overcome this uncertainty there is a need for a dedicated future rover mission to study signs of explosive volcanism on the surface of Mars and therefore allow us to gain a better constraint on this important aspect of martian history.

8. References

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Declaration of interests

- ☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.
- ☐ The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

- We present an overview of explosive volcanism on Mars
- Evidence for explosive volcanism is less common than for effusive activity
- Still indications of explosive volcanism have been identified at various sites
- Explosive edifices are often different in shapes from their terrestrial analogues
- Explosive eruptions on Mars would behave differently from those on Earth

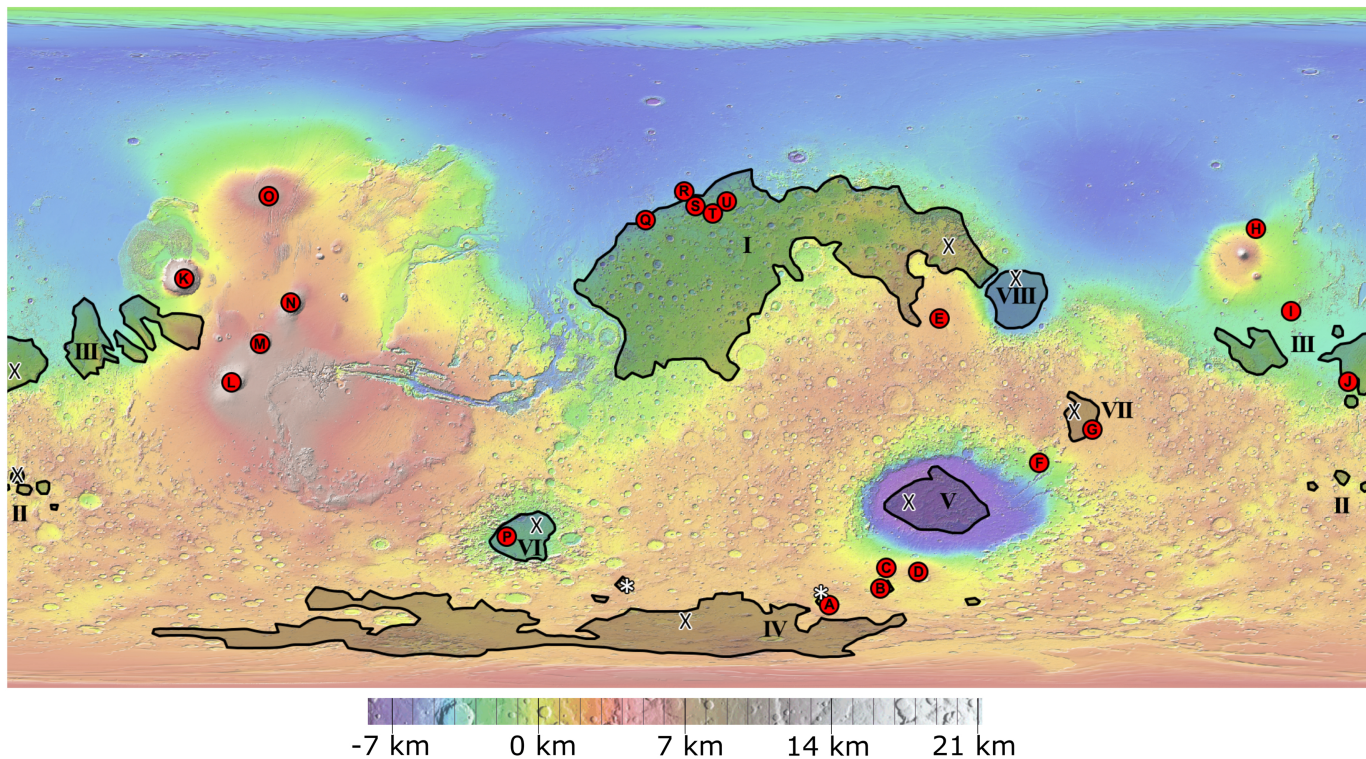


Figure 1

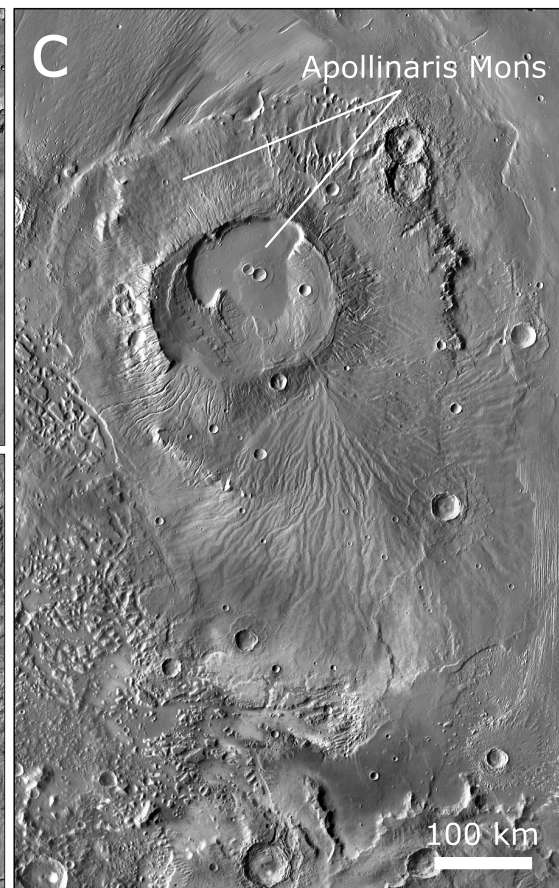
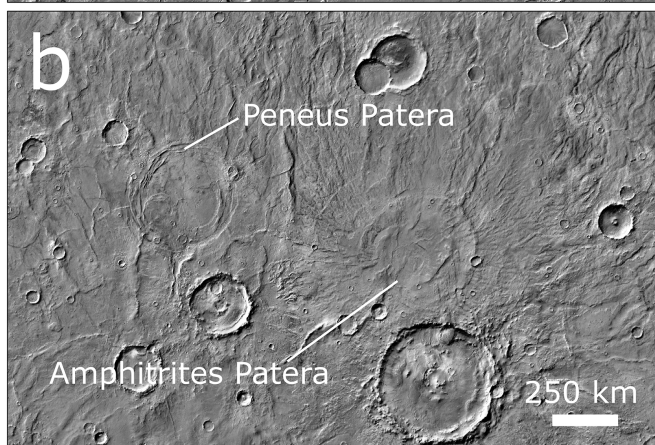
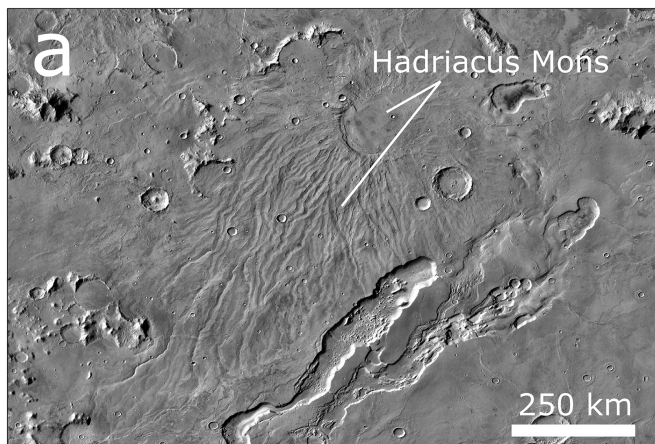


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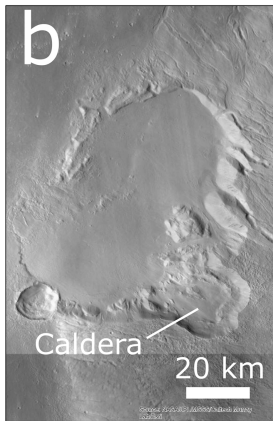
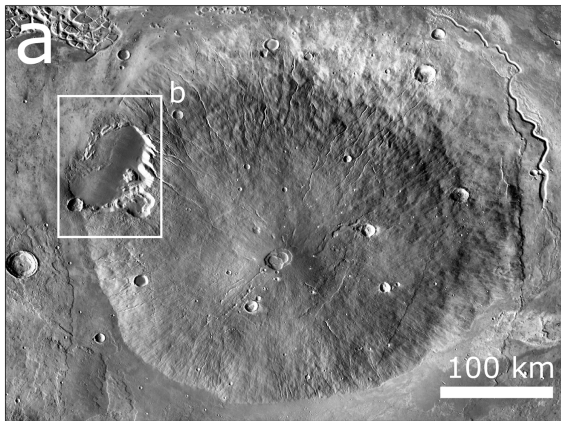


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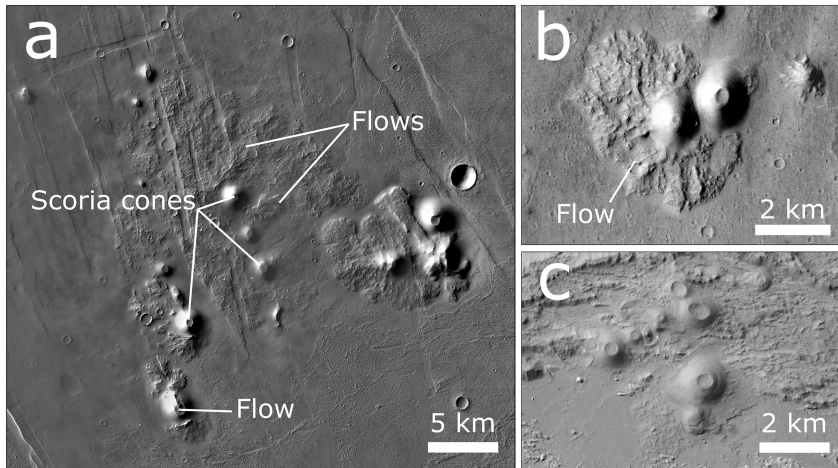


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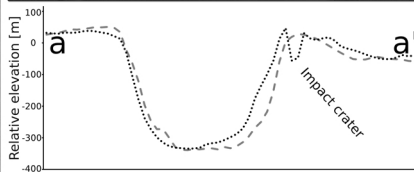
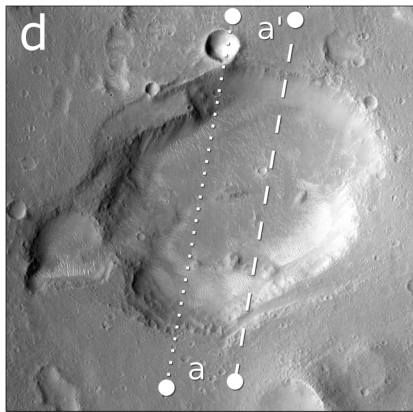
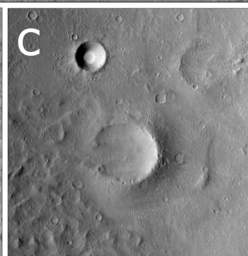
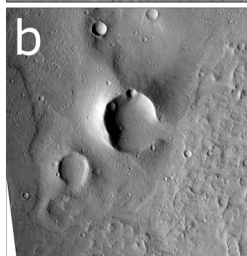
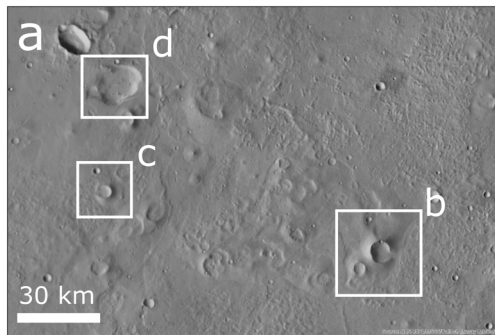


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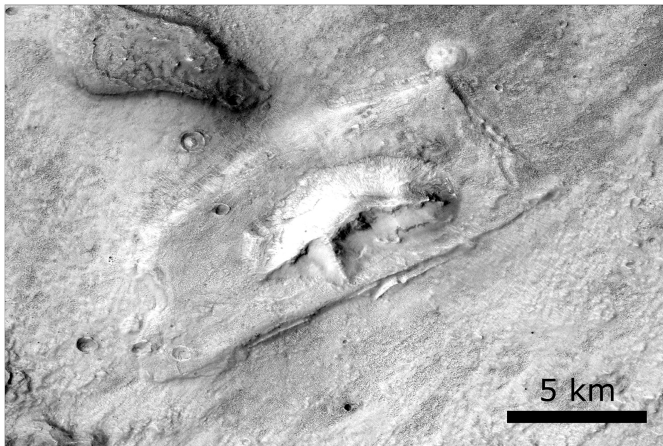


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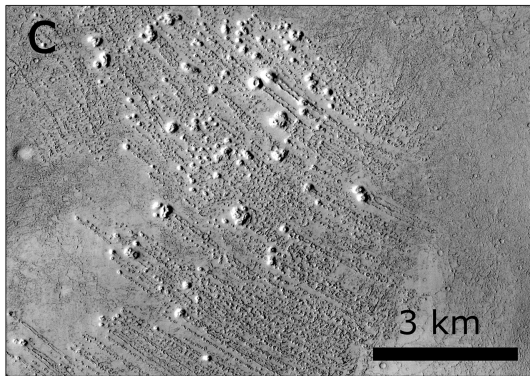
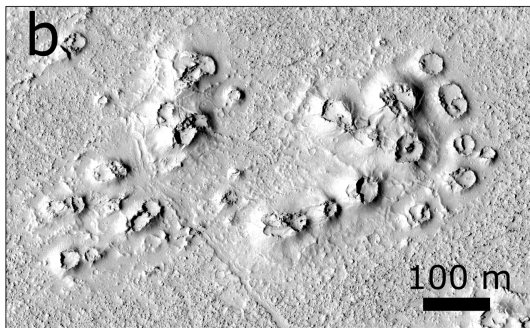
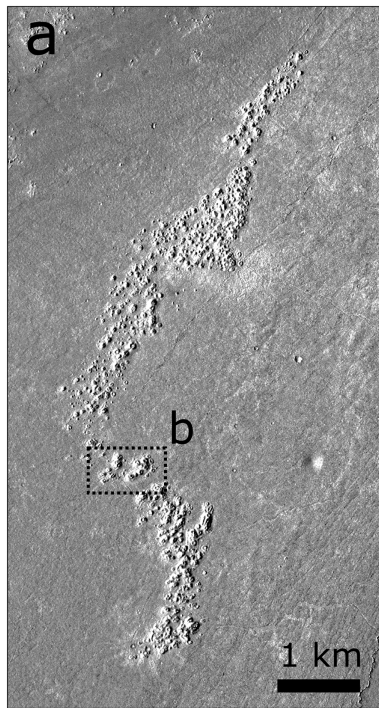


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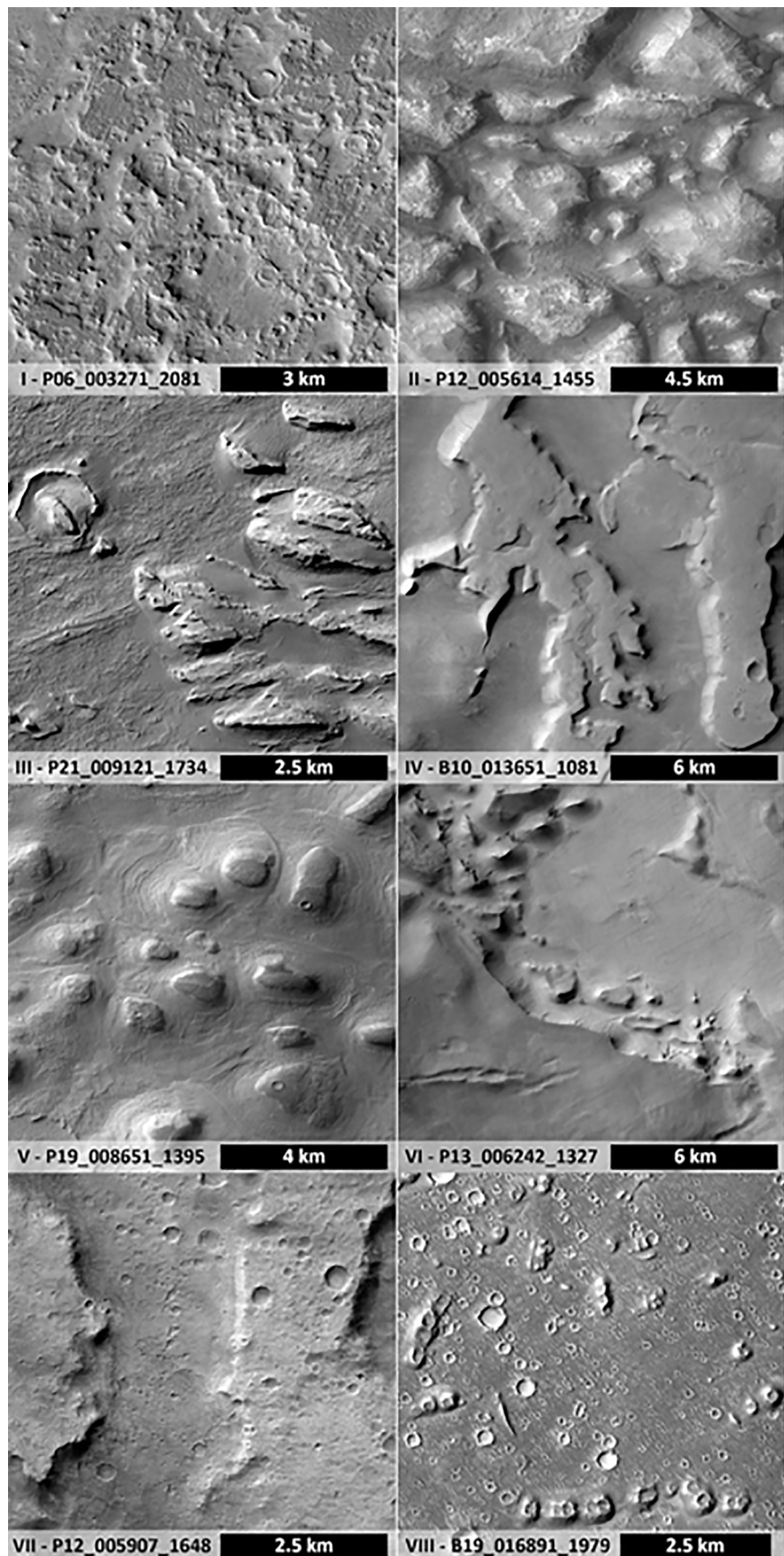


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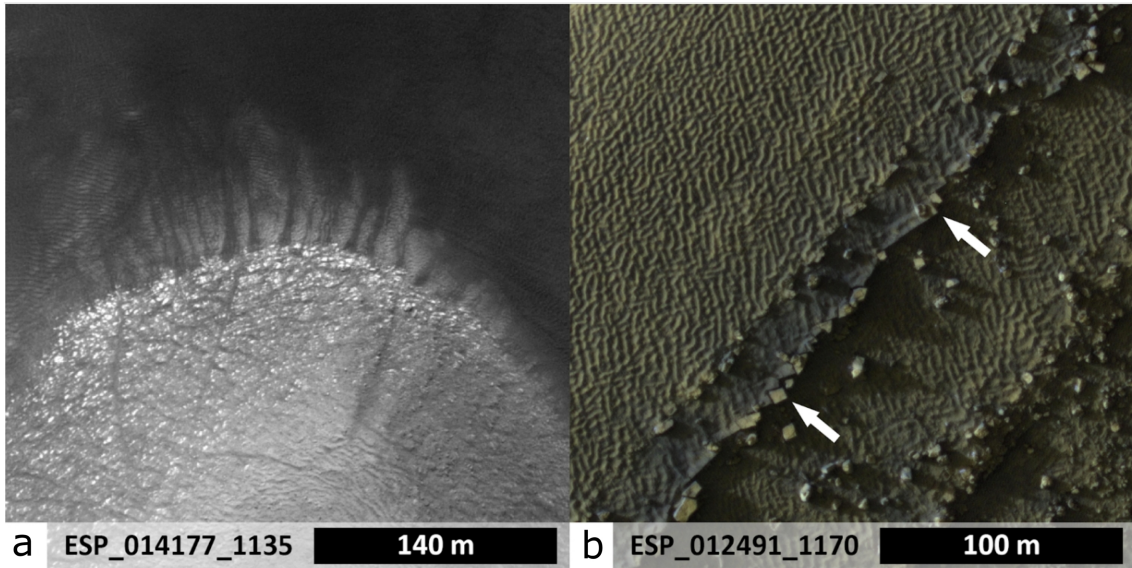


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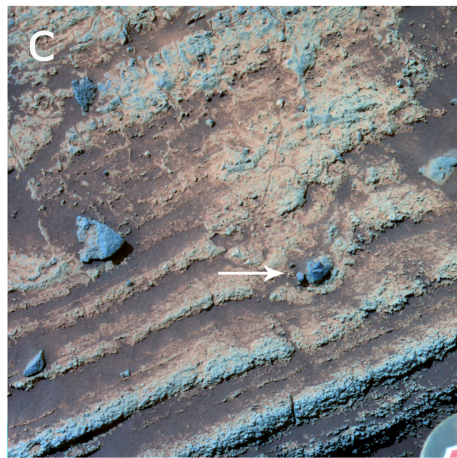
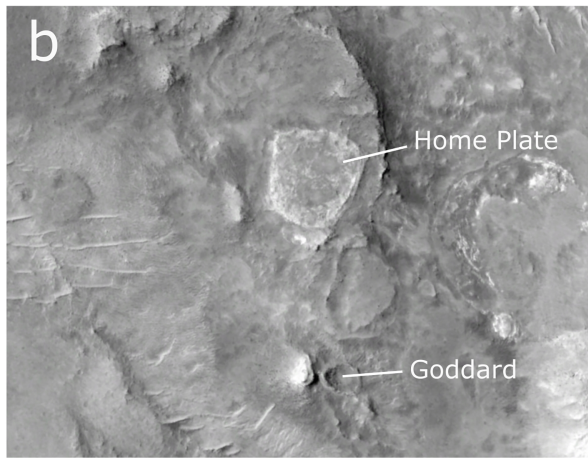
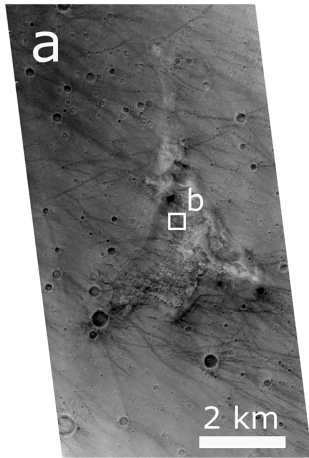


Figure 10