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Mud Flow Levitation on Mars: Insights from Laboratory Simulations

P. Brož^{1*}, O. Krýza¹, S. J. Conway², N. T. Mueller³, E. Hauber³, A. Mazzini⁴, J. Raack⁵, M.R. Balme⁶,
M.E. Sylvest⁶ and M. R. Patel^{6,7}

¹Institute of Geophysics of the Czech Academy of Sciences, Boční II/1401, 141 31, Prague, Czech
Republic

²CNRS UMR-6112 Laboratoire de Planétologie et Géodynamique, Université de Nantes, France

³Institute of Planetary Research, DLR, Rutherfordstr. 2, 12489, Berlin, Germany

⁴Centre for Earth Evolution and Dynamics (CEED), University of Oslo, Norway

⁵Institut für Planetologie, Westfälische Wilhelms-Universität Münster, Germany

⁶School of Physical Science, STEM, The Open University, Milton Keynes, UK Open University, Milton
Keynes, United Kingdom

⁷Space Science and Technology Department, STFC Rutherford Apple-ton Laboratory, Oxford, UK

Corresponding Author

Petr Brož

Institute of Geophysics of Czech Academy of Sciences (CAS) v.v.i

Boční II/1401

14131 Prague 4

Czech Republic

Petr.broz@ig.cas.cz

+420267103063

Abstract

Sediment mobilisation occurring at depth and ultimately manifesting at the surface, is a process which may have operated on Mars. However, the propagation behaviour of this mixture of water and sediments (hereafter simply referred to as mud) over the martian surface, remains uncertain. Although most of the martian surface is below freezing today, locally warmer surface temperatures do occur, and our current knowledge suggests that similar conditions prevailed in the recent past. Here, we present the results of experiments performed inside a low pressure chamber to investigate mud propagation over a warm (~ 295 K) unconsolidated sand surface under martian atmospheric pressure conditions (~ 7 mbar). Results show that the mud boils while flowing over the warm surface. The gas released during this process can displace the underlying sand particles and hence erode part of the substrate. This “entrenched” flow can act as a platform for further mud propagation over the surface. The escaping gas causes intermittent levitation of the mud resulting in enhanced flow rates. The mud flow morphologies produced by these phenomena differ from those produced when mud flows over a frozen martian surface as well as from their terrestrial counterparts. The intense boiling removes the latent heat both from the mud and the subsurface, meaning that the mud flow would eventually start to freeze and hence changing again the way it propagates. The diverse morphology expressed by our experimental mudflows implies that caution should be exercised when interpreting flow features on the surface of Mars and other celestial bodies.

1. Introduction

The surface of Mars is characterized by features with various shapes (Fig. 1a,b) ranging from decimetres to kilometres in scale whose origin has been attributed to the action of sedimentary volcanism (e.g., Allen et al., 2013; Hemmi and Miyamoto, 2018; Oehler and Allen, 2012; Pondrelli et al., 2011; Salvatore and Christensen, 2015; Skinner and Mazzini, 2009; Skinner and Tanaka, 2007; Okubo, 2016; Komatsu et al., 2016; Rubin et al., 2017; Brož et al., 2019; Wheatley et al., 2019; Kumar et al., 2019). However, this interpretation is not unanimous, as igneous volcanism has also been proposed as a formation mechanism for several of these features (e.g. Brož and Hauber, 2013; Brož et al., 2017). Regardless of this ambiguity, if sedimentary volcanism (Oehler and Etiope, 2017) has ever been present

on Mars, the mud would propagate in an environment significantly different from the terrestrial one, in particular with respect to atmospheric pressure and gravity.

Until recently it remained unclear as to whether such different environmental conditions would even allow mud to propagate over the martian surface. This uncertainty was related to the fact that the martian low-pressure environment inhibits the sustained presence of liquid water on the planet's surface (e.g., Bargery et al., 2010; Hecht, 2002). Previous studies have shown that water would boil and this could significantly affect sediment transportation (Conway et al., 2011; Massé et al., 2016; Raack et al., 2017; Herny et al., 2018). These results suggest that martian environmental conditions may have a profound effect on mud rheology and hence its propagation may vary from our terrestrial experience.

On Earth, the effusion rates and volumes of ascending mud vary, and affect the sizes, thicknesses and shapes of mud flows. Low effusion rates lead to the formation of flows a few centimetres-thick capable of propagating up to several meters (Fig. 1c,d,e). In contrast, high effusion rates and volumes may lead to flows metres thick capable of propagating over kilometres. However, the sizes and thicknesses of mud flows strongly depends on its viscosity which is mainly linked to the water content. Water-dominated mud flows have thicknesses varying from metres to centimetres. This is because the flows preferentially spread laterally and therefore their thicknesses decreases towards the margins forming overlapping lobes.

While the behaviour of mud during emplacement and its rheology under terrestrial conditions is well studied and understood (e.g., O'Brien and Julien, 1988; Laigle and Coussot, 1997 and references therein), this is not the case for Mars, nor for other planetary bodies within the Solar System, where sedimentary volcanism has also been proposed (e.g., Ruesch et al., 2019). An initial study examining the general behaviour of kilometre-sized mud flows in a low pressure environment was performed by Wilson and Mouginis-Mark (2014), where some aspects of mud propagation over the martian surface were discussed from a theoretical point of view. The authors proposed that the water present in the mud would be unstable and hence evaporate from the mud flow, ultimately removing the latent heat from the mixture. As a consequence, the residual water within mud should freeze in a relatively short period of time, hours to days. Recent insights came from experimental work of Brož et al. (in

press) in which the behaviour of low viscosity mud was experimentally studied in a low pressure chamber that partly simulated the environment of Mars. Their work demonstrated that low viscosity mud flows can propagate over a cold (<273 K) surface under martian atmospheric pressure, however, the mechanism of such propagation would be very different from that observed on Earth. On Mars, mud would rapidly freeze due to evaporative cooling (Bargery et al., 2010) forming an icy-crust leading to the propagation of the decimetre thick mud flows resembling pahoehoe lava flows on Earth (Brož et al., in press).

The average temperature of the martian surface today is far below the freezing point of water, but thermal infrared observations (e.g. Sinton and Strong, 1960; Kieffer et al., 1977; Christensen et al., 2001) show locally higher temperatures can occur at certain locations at favourable seasonal times (Hecht, 2002). This is mainly because the redistribution of solar heat is impeded by the low atmospheric pressure, which limits advective heat transport by the atmosphere, and by the low thermal conductivity of the regolith covering most of the surface (e.g. Presley and Christensen, 1997). Numerical thermal models of the surface have been developed to derive regolith thermophysical properties such as albedo and thermal inertia from the observed temperatures (e.g. Kieffer et al., 1977, Putzig and Mellon, 2007). Conversely, we can use such models and the thermophysical properties to calculate the subsurface temperature structure that is consistent with observed temperatures. The results are shown in Fig. 2 and the details of calculations are described below in Section 2.2. Our data show that conditions above freezing are not rare on Mars and that some significant amount of heat can be released to potentially keep a mud flow from freezing.

The goal of this manuscript is therefore to investigate the behaviour of low viscosity mud flowing under martian pressure conditions over a warm (from ~ 292 K to ~ 296 K) surface. More specifically, we examine how the instability of water within the mud would change its flow behaviour during the phase change from liquid to gaseous. Hence, we aim to reveal which transport processes would be reasonable to expect during sedimentary volcanism on Mars and how they may affect the final morphology of the resulting mud flows.

2. Methods

2.1. Experimental setup

We performed a set of experiments (see Table 1 for details) using the Mars Simulation Chamber at the Open University (UK). The chamber was equipped with a 0.9×0.4 m aluminium tray filled with a ~ 2 cm deep sediment bed (natural sand, ~ 200 μm) together with a reservoir containing 500 ml of low viscosity, and hence water-dominated, mud (12.7 mPa*s at ~ 276 K and 10.7 mPa*s at ~ 296 K). The container with the mud hung ~ 5 cm above the tray (Figure 3). Before depressurisation, the mud and sand were at room temperature (from ~ 292 K to ~ 296 K). The tray was inclined by 5° and 10° to force the mud to move under gravity once poured onto the surface. The mud was released from the container under reduced pressure (~ 7 mbar) and the movement of the mixture was observed and recorded by four cameras from different angles. In some experiments the sand bed was replaced by a smooth plastic plate inclined by 5° to assess if the observed processes were dependent on the presence of unconsolidated and porous material. We also performed comparative experiments under terrestrial pressure to get a reference set of experiments. In an additional experimental setup, a plastic box (0.6×0.4 m) filled by a 33 cm thick layer of sand was used to investigate the depth to which the mud is able to propagate vertically. Each experiment was performed in triplicate to confirm the reproducibility of the results.

The mud used for the experiments was a mixture of deionised water with 0.5% w/w of dissolved magnesium sulphate salts (MgSO_4) and clay at ~ 6.5 % of the mud mass (the mass ratio between deionised water and clay was 4:1). The magnesium sulphate salt, which has previously been detected on the martian surface (Clark, 1978; Vaniman et al., 2004; Hecht et al., 2009), was added to achieve the average river water salinity which is necessary to suspend submillimetre clay particles (Corradi et al., 1994). As there is no direct in-situ knowledge of which types of clays could be involved in the subsurface sediment mobilisation on Mars, we decided to use the clay obtained from the claystone named after the Rokle locality situated near the town Kadaň in the Czech Republic and operated by the private company Keramost. This clay is a bentonite composed of 76% montmorillonite, 23% illite, and 1% kaolinite and formed by alteration of pyroclastic rocks. As explosive volcanism was likely common on Mars (e.g., Wilson and Head, 1994), to a first approximation this material is a suitable analogue. The mud mixture

was obtained using a blender for 3 minutes to reduce the presence of more lithified clayey aggregates. The average density of the resulting mixture was 1037.5 kg/m^3 . The viscosity was measured at the Institute of Hydrodynamics of the Czech Academy of Science by using Haake Rotovisco RV 20 and Viscotester VT 550 rheometers with ledges on the MV2 cylinder to prevent slip of the measured material on its walls.

2.2. Modelling of Mars surface temperatures

To model the subsurface temperature on Mars we used the albedo and thermal inertia maps derived by Putzig and Mellon (2007) from Thermal Emission Spectrometer data (Christensen et al., 2001). Thermal inertia is the square root of the product of thermal conductivity, bulk density, and specific heat capacity. Since the latter two exhibit much less variability in regolith than the former, we assume they are constant. Density is assumed to be 1300 kg/m^3 and heat capacity is assumed to be 630 J/kg/K from the numbers recommended for the InSight landing site (Morgan et al., 2018).

The variables governing the boundary conditions are Mars' orbit and spin axis obliquity, as well as atmospheric pressure and dust optical thickness. For our modelling we use the timeseries output of the local surface atmospheric pressure and local dust optical thickness at the 610 Pa pressure level provided by the average climatology scenario of the Mars Climate Database (MCD, Forget et al. 1999, Millour et al. 2017). The database covers the average of TES observations over martian years 24 – 31, with planet encircling dust events excluded (Montabone et al., 2015). We scale the dust opacity to the local atmospheric column by multiplying with the ratio of local pressure to 610 Pa (Montabone et al., 2015). For the calculations of subsurface temperature, we use a 1-D version of the Mars Climate Database MCD, which calculates the downwelling visible and infrared fluxes and solves the heat conduction equation in the subsurface, assuming a zero heat flux lower boundary condition. The calculations are done for 10×10 degree tiles, with each tile having the average surface albedo and thermal inertia, and the average from the surface pressure and dust opacity timelines. The output is generated at intervals of 15° of solar longitude and 0.5 h of local solar time for the duration of one Mars year. The results are validated by comparison with version 5.3 of the MCD and found to be consistent within a few Kelvin. This dataset is searched to find the maximum surface temperatures, presented in Fig. 2a. The percentage

of the total time of a Mars year when the surface temperatures are above freezing is presented in Fig. 2b. The maximum depth of subsurface layer above freezing is shown in Fig. 2c and the maximum heat that can be released from that layer before everything is below freezing in Fig. 2d.

3. Observations

Once the atmospheric pressure was reduced inside the chamber, the mud in the container started to boil and to cool due to evaporative cooling. The boiling intensified as the pressure was decreased to 12-14 mbar and continued to 7 mbar. When a pressure of ~7 mbar was reached, the mud had a temperature slightly above freezing and was manually released by tipping the container, letting it flow over the 'warm' (from ~292 K to ~296 K) sand surface inclined at 5°. The contact of the mud with the warm surface triggered boiling, which caused ejection of sand grains to a height of several centimetres. The particles landed both on the mud and on the surrounding sand. The deposition of the sand grains formed a small raised rim around the contact area resulting in a crater-like depression (Fig. 4, t=10 s). The explosive activity decreased with time. At the beginning, the mud was not visible inside the crater area, as it was covered by a layer of loose sand grains, which was repeatedly disturbed by bubbling (Fig. 4, t=10 s). Within seconds, mud could be observed on the surface – not necessarily at the site where it was directly poured from the container – propagating inside the crater (Fig. 4, t=22 s). At the boundary between the mud and the sand layer, a large number of millimetre-scale explosion pits formed, from which gas continued to eject particles for several minutes. This caused a progressive expansion of the rim.

Continued mud supply caused the flow to breach the sandy rim and a lobe of mud advanced over the warm sand (Fig. 4, t=22 s). This flow front triggered new explosions as the mud propagated and infiltrated into the surface of the sand. The escape of gas at the base of the mud flow caused the lobe to vibrate vertically and to quickly levitate over the first few centimetres of the sand surface (Fig. 4, t=26 s; Fig. 5a). When boiling was insufficient to lift the entire weight of the mud flow, the flow lobes slid/crept over the surface entraining the sand particles. Then the lobe stalled and small millimetre-scale explosions occurred around its edge forming small ridges. Simultaneously fresh mud outpouring from the crater started to propagate over the lobe's surface and accumulate at the

front of the flow (Fig. 4, $t=30$ s). Once enough material had accumulated to overcome the small ridges at the edges, a new lobe formed (Fig. 4, $t=34$ s to $t=48$ s) and the process repeated until the supply of mud was exhausted (Fig. 4, $t=49$ s and $t=105$ s). The movement of mud through the lobes created an interior trough with a curvy and irregular shape (Fig. 4, $t=928$ s).

When the same experiment was repeated with the surface inclined at 10° , a similar behaviour was observed (Fig. 6). However, the flow lobes travelled faster and further than lobes at 5° inclination. The resulting deposited lobes were also longer and narrower than in the previous set of experiments (Fig. 6, $t=30$ s). In some cases, the flow lobe became separated from the main flow and/or from the source crater by a layer of sand grains ejected by escaping gases from the sand layer (see the evolution of the left flow lobe on Fig. 6 from $t=25$ s to $t=35$ s), giving the flow a discontinuous appearance.

The boiling within the bulk volume of the mud also caused the nucleation of variously sized bubbles which moved upwards to the surface of the mud. Once they reached the surface, they dramatically expanded and as a result some portions of the mud were fragmented into small droplets (Fig. 5b). These droplets were then ejected along ballistic pathways and deposited close to the active mud flow (e.g. Fig. 4, note the presence of small dark droplets whose frequency increased around the flow between times $t=30$ s and $t=105$ s). The influence of the mud fragmentation on the total transported mud volume was only minor, however, it occurred over the entire length of the mud flow wherever the bubbles were able to form and move through the mud. This mechanism was not active during the experiments performed under terrestrial pressure conditions. Part of the ejected sand and mud was also deposited on the mud flow where it formed a sandy-muddy crust (Fig. 4b) which partly hindered the mud flow. The crust was either later destroyed by a new batch of mud or by a gas release, or it survived until the end of the experiment.

We also observed the release of gases in the form of centimetre-sized bubbles for dozens of minutes in the pouring area and in several other areas along the entire length of the flow providing evidence that the mud remained liquid in the subsurface (Fig. 5b). The presence of liquid mud covered by sand was later confirmed once the chamber was recompressed and the interior of the

resulting features were studied by breaking them apart from their edges to the source area (Fig. 7). By using an additional experimental setup, a plastic box infilled by a 33 cm thick layer of sand, we found that the mud flow was capable of eroding into the sand layer to a depth of several centimetres as mud was observed beneath the original sand surface. This setup also revealed that the mud flow was surrounded by a layer of sand saturated by water (Figs. 7b-d).

The inspection of the final morphology of the mud flows once the chamber was recompressed revealed that the internal structures of the flows were supported by a hardened mixture of mud and sand allowing the formation of vertical cliffs or overhangs forming crusts which fully or partially covered the flow channels. These crusts enclosed cavities through which the flows had propagated (Fig. 7f). We also observed that the bottoms of the troughs were covered by fine-grained clay. Repeated gas explosions had formed holes which were located above small subsurface pockets infilled by mud.

The experiments performed under terrestrial pressure did not show the same behaviour or morphologies as those at martian pressure (Fig. 8). The mud within the container before pouring did not boil and its internal temperature did not drop. Once the mud was poured from the container (Fig. 8, $t=5$ s), it started to flow relatively uniformly over the sand surface (Fig. 7f) and, as a consequence, a broad and thin mud flow formed (Fig. 8, $t=20$ s). In some cases, centimetre-sized flow lobes formed on the edge of the propagating flow (Fig. 8, $t=25$ s and $t=85$ s) as a result of topographic irregularities in the sand layer. The flow did not significantly erode into the sand surface and vertical cliffs or protective muddy-sandy crust were not observed to form.

The experiments performed under reduced atmospheric pressure over the 5° inclined plastic plate showed that boiling still occurred, but it was not able to lift the propagating mud flow. Once the mud was poured from the container, it started to flow over the surface forming a several centimetres wide and a dozens of centimetres long flow. As no unconsolidated material was available, the flow did not erode to the subsurface and hence the direction of the propagation of the mud flow was only controlled by gravity. Analogue experiments performed under terrestrial pressure lead to the formation of similarly looking mud flows, however, no boiling was observed.

4. Discussion

4.1. Application to Mars

Our experiments revealed that a low pressure environment has a significant effect on the propagation of low viscosity mud over a warm unconsolidated surface (Fig. 9). The reason for this different behaviour is the instability of water (Bargery et al., 2010) that leads to boiling and the formation of water vapour bubbles which escape violently from the mud to the surrounding atmosphere and partly also into the subsurface. The strongest influence of the boiling on the mud movement occurs at the contact between the mud and the warm material over which the mud propagates. Here the boiling is most intense as the heat stored within the substrate is available to support boiling. As a result, a large quantity of water vapour is produced at this interface (Fig. 5a). The release of water vapour can then modify the way the mud moves leading to levitation and sliding/creeping of the advancing mud lobes.

This is similar to the transport mechanism described by Raack et al. (2017) and Herny et al. (2018), who studied the downslope movement of water-saturated sediments in a low pressure environment, albeit with one major difference. Whereas Raack et al. (2017) and Herny et al. (2018) observed that the water-saturated sediments propagated in the form of individual small pellets, the levitating mud observed herein moves as a coherent fluid (Figs. 2, 3 and 6a). The entire flow lobe is therefore affected by the levitation. Once the gas production in the advancing lobe decreases, the mud seems to propagate over the inclined unconsolidated surface by sliding and/or creeping (Figs. 2, 3 and 6a) as the gas release can only partly lift the lobe. These mechanisms cause an increase in the speed of mud propagation over the unconsolidated surface under martian ambient pressure (Figs. 2 and 3) as compared to terrestrial ambient pressure (Fig. 8).

The boiling of water has also another effect on mud flow propagation in regard to its stability in the martian environment. This is because the ejection of sandy grains by expanding gas and their re-deposition on top of the mud flow can form a sandy-muddy crust by coalescing sand and finer-grained clay particles together. The crust can partly isolate the liquid mud from the atmosphere and prolongs its lifetime on the surface. The mud can then propagate under the crust giving rise to the

formation of an interconnected network of mud channels beneath the crust (e.g. experiment #26 and #31).

The crust which developed in our experiments extended along the flow over several centimetres of the mud flow channel (e.g., Figs. 4, 5, 6 and 7), but real mudflows are likely to be longer and wider, so it is difficult to estimate the efficacy of this process at field-scale. We hypothesize that the limited coverage of the flow in our experiments was caused by the fact that these experiments were volume-limited and the processes associated with the development of the mud flow operated for only a relatively short period of time. If the mud volume were larger, the flow would last for longer and more mud-sand interaction would occur and perhaps a crust may be formed over a larger surface area of the mud channel. If a larger section of the channel were covered by the crust, this may fully isolate the flow from the low atmospheric pressure, causing boiling to cease. Under such scenario, boiling would only occur at the active edges of the mud flow where the mud would be exposed to the atmosphere. Even if the crust could only cover part of the channel over its entire length, our observations suggest that mud propagating over a warm surface – similarly to the mud propagating over a cold surface (Brož et al., in press) – is capable of developing a protective crust prolonging the lifetime of the flow on the surface of Mars. This is an aspect of mud propagation that has not previously considered and should be investigated further. However, it should be noted that this process may only be important for martian mud flows whose width is within the range of centimetres to decimetres, and less significant for wider ones. This is because even under the reduced martian gravity the ballistic trajectory of the particles is limited. Furthermore, a larger width of the flow would mean that the volume of ballistically transported particles, necessary to form a sufficiently thick crust, would significantly increase. As a consequence it would be difficult to form a crust over a flow more than several tens of centimetres wide. Therefore, decimetre-wide flows would show variations on their surfaces as the flow edges (unlike the central part) would be affected by the presence of a newly formed crust. Hence, the central part would display patterns associated with flow processes. It should be highlighted that we predict crust formation only on martian mud flows propagating over mobile substrates, such as, unconsolidated sandy regolith. As shown by the experiment setup using the plastic plate, if the mud propagated over impermeable rocky or crusty

surfaces, even though boiling would still occur, the crust would not be formed because of the absence of small particles that could be ejected onto the mud. The properties of the substrate could significantly influence the way the mud propagates, and the final morphology observed at the surface.

We also observed that mixing of the mud and sand at the flow boundaries produced vertical cliffs or even overhangs which were relatively stable as shown by their capacity to survive the processes of chamber repressurisation (Figs. 7a,d and 8). We expect that within the range of centimetres to decimetres wide mud flows such cliffs may result in atypical morphologies compared to those found in terrestrial mud flows suggesting that comparison to terrestrial analogues could be misleading. However, we do not expect that shapes of metre- or kilometre-sized mud flows would be affected as the mechanical strength of such cliffs would be limited. Centimetre-sized cliffs can be attained, whereas metre-sized cliffs would collapse. Small mud flows can therefore share some similarities with terrestrial low viscosity lava flows for which vertical cliffs or overhangs are common (e.g., Kilburn, 2000) as lava solidification is capable of forming such landforms.

4.2. Temperature drop within the mud and substrate

In addition, our experiments show that there is another crucial factor which deserves further investigation when studying surface expressions of possible martian mud flows. The instability of the water causes a temperature drop in the mud due to evaporative cooling which removes latent heat from the mud and from the subsurface (Bargery et al., 2010; Brož et al., in press). So, once the ascending mud reaches the martian surface, the mixture would rapidly self-cool close to the freezing point.

Specifically, when the mud initially propagates over a warm surface, and the flow continues over the same area for a certain amount of time, evaporative cooling removes the available heat both from the mud and the substrate. This results in gradual freezing of the mud and the subsequent formation of an icy-muddy crust (Brož et al., in press), altering the mechanisms of mud propagation. Ultimately the mud would propagate via frozen mud tubes in similar fashion as pahoehoe lava (Brož et al., in press). This leads us to conclude that mud movement over today's martian surface would involve a combined process governed both by levitation and/or sliding at the front and by freezing as the flow

progresses. Both mechanisms may occur during the formation of a single mud flow and this suggests that the resulting morphologies may record these variations of mud transport.

To obtain a first order of understanding regarding what happens in case of a mud flow over a warm surface, we use the subsurface temperature profiles corresponding to the maximum ground temperature case in Fig. 2a as the initial temperature state. We force the top of that temperature profile to be at 273 K, representing the coldest possible temperature in contact with boiling water. We then solve the heat conduction equation starting from the above described condition, using the explicit Euler finite difference method with 1 mm depth resolution and 0.1 s time resolution. The model is stopped when the entire profile is below 273 K, the time until at that moment is presented in Fig. 2e. For comparison we also conduct this cooling model with a profile resembling the laboratory setup: i.e. assuming a thermal conductivity of $0.04 \text{ Wm}^{-1}\text{K}^{-1}$ for the 0.2 mm particle diameter sand at 700 Pa (Presley and Christensen, 1997) and an initial constant temperature of 273 K. The timescale of cooling below freezing everywhere is not meaningful in this context, but we can calculate the heat flow conducted through the surface as a function of time. This heat flow q with unit Wm^{-2} can be translated into a vapour column production rate v with unit of m s^{-1} , which is the thickness of the vapour layer produced per second:

$$v = q R T / (L_v m_{H_2O} P)$$

where $R = 8.314 \text{ J K}^{-1} \text{ mol}^{-1}$ is the ideal gas constant, $T = 273 \text{ K}$ is the assumed gas temperature, $L_v = 2.46 \times 10^6 \text{ Jkg}^{-1}$ is the latent heat of water vaporization, $m_{H_2O} = 0.018 \text{ kg}$ is the weight of 1 mol of water, and $P = 611 \text{ Pa}$ is the saturation pressure of water at temperature T . Temperature and pressure below a levitating flow will likely be higher, and the volume lower, than these assumed values, since it must compensate for hydrostatic and ambient pressure. It is likely that for levitation a critical production rate, proportional to the heat flow out of the top of the substrate must be sustained. In the laboratory experiments the levitating is intermittent, with a typical duration on the order of 10 s after the emplacement of fresh mud flow lobes (Fig. 4). In the thermal model of the laboratory setup with the top of the substrate forced to 273 K, the heat flow conducted from the surface is $> 550 \text{ Wm}^{-2}$ in the first 10 seconds. This corresponds to column of vapour produced per second of $v = 46 \text{ cm s}^{-1}$. It should be noted

that not all of the heat from the subsurface will produce water vapour from the mud flow, some heat will be transported away by the vapour.

Nevertheless, taking 550 Wm^{-2} as requirement for mud levitation, we can apply it to the model of the forced cooling of Mars surface to estimate how long flows could levitate on Mars, Fig. 2f shows the duration of heat flow $> 550 \text{ Wm}^{-2}$ conducted through the substrate surface forced to a temperature of 273 K starting from the maximum modelled temperature conditions (Fig. 2a). This map shows that over a band of latitudes of 60°S to the equator, levitation is possible if a mud flow occurs close to noon during the warmest season. In the northern hemisphere this also applies to parts of Syrtis Major Planum and Chryse Planitia, due to their relatively high albedo.

This model neglects the observed ejection of substrate particles from below the flow, which would initially increase the heat flow from the subsurface by exposing deeper layers. At some point however the erosion of the substrate would expose layers below zero, so that this process would stop at most at a few centimetres of depth (Fig. 2c). Also neglected is the possibility of water penetrating into the subsurface, either in vapour or in liquid form, which would likely speed up cooling of the substrate.

We are also aware that the environmental properties on Mars vary over time, due to the chaotic variation in Mars' orbital parameters, which affects the distribution of incoming insolation (Laskar et al., 2004). For example, during periods of higher obliquity (Ward, 1973) areas where surface temperatures could be above the freezing point of water would vary in magnitude and position. Also, the depth to which the freezing point would descend would vary (Costard et al., 2002) and hence the amount of available heat necessary to keep the process of the violent boiling active would be different. The time required by the evaporative cooling to remove the available heat from the subsurface increases proportionally with the depth of the freezing point. Variations in surface temperatures caused by variations in Mars' orbital parameters may therefore prolong or foreshorten the period over which boiling would dominate the behaviour of the mud. This means that the mechanisms dominating in mud propagation should change through time.

Additionally, the complex magmatic history of Mars (e.g. Grott et al., 2013) suggests that the heat flux within the martian crust has not been steady. In fact, it has varied through time at various

localities producing geothermal anomalies (e.g. Plesa et al., 2016). These sites may represent ideal places where sedimentary eruptions may result due to the melting of subsurface water ice or hydrothermal activity. Higher heat flux may also imply that the surface can be sufficiently warm to prevent the freezing of the extruded mud for extended periods of time, or even completely inhibited. Under such a scenario, it may be possible that the described processes, which operated in our experiments at the centimetre-scale only, would be able to affect even wider surfaces and therefore produce larger scale mud features. It should therefore be taken into account that some morphologies which are present at the surface of Mars today, and which may have an origin associated with mud eruptions, may be linked to the intense boiling and the above described transport mechanisms.

4.3. Scaling to Mars and experimental limitations

The presented experimental approach has several limitations as it cannot fully attempt to simulate all the conditions of sedimentary volcanism on Mars or even on Earth. Firstly, our experiments were performed with a fixed volume of mud (500 ml). This limits the size of the resulting flows to a length of several decimetres, the width to several centimetres and the thickness to less than 1-2 centimetres. On Earth, sedimentary volcanism can extrude mud volumes several orders of magnitudes larger than those in our experiments (e.g. Kopf, 2002; Mazzini and Etiope, 2017) resulting in metre- to kilometre-long mud flows. However, low volume eruptions are more common than voluminous ones. Obviously, we were not able to perform experiments at the metre- to kilometre-scale due to the size limitation of the available low pressure chamber. Nevertheless, our work can still provide valuable insights into mud propagation in a low pressure environment due to the chosen mud viscosity. This is because low viscosity mud flows, naturally occurring on Earth, propagate over flat surfaces preferentially on their edges forming centimetres-thick mud lobes (Mazzini and Etiope, 2017; Miller and Mazzini, 2018) similar to those observed in our experimental setup. Therefore, we are capable to directly address, some of the differences between those terrestrial flows and putative martian analogues. Our work therefore suggests that environmental differences between Earth and Mars would affect the propagation of both small and large martian mud flows, but with differing intensity. However, our experiments do not provide appropriate insights about the propagation of thicker low viscosity flows, for example, when the mud becomes channelized or a large quantity of mud is

extruded over a short period of time. This is because in thick mud flows the flow regime may change. This should significantly change the way the mud propagates in the low pressure environment as heat will be transported differently within the flow and there is a need for further experimental or theoretical investigation of this problem.

An additional limitation is associated with the inability to perform experiments in a reduced gravity environment, mimicking the gravity of Mars (3.7 m s^{-2}). Our experiments were performed under terrestrial gravity (9.8 m s^{-2}), which has an influence on the way the mud propagates. Terrestrial gravity limited the height to which the vibrating mud could levitate as well as the total volume of mud which could be lifted. A similar limitation was encountered in the study of Raack et al. (2017) in which the different gravitational force affects the levitation of the sedimentary pellets propagating over the hot surface. These authors showed that the reduced gravity would allow pellets to levitate for up to 48 times longer on Mars than on Earth and that the lifted pellets could be up to seven times heavier compared to those lifted in their experiments. These calculations suggest that, on Mars, mud would levitate for a longer period of time and to a greater height than observed in our experiments. The lower martian gravity would also affect the distance to which the sand grains and small mud droplets would be ejected. Brož et al. (2014) showed that ballistically emplaced scoria particles can travel about 20 times further on Mars than on Earth due to the lower gravity and lower atmospheric pressure. Therefore, the ejected material would be spread over much wider area than observed in our experiments. This means that on Mars more transported material would be necessary to build a steep-sided rim which in part affects the flow of material over the flat surface. As a consequence, the role of marginal rims to, at least partly, guide the flow direction may be limited on Mars as the rims may be much broader but shallower than those observed in our experiments. Ultimately gravity may also affect the way the flow erodes the subsurface as sand and other particles would be more easily shifted by the gases released from the advancing mud flow. To overcome these limitations, it would be necessary to develop a dedicated numerical model and perform additional analogue experiments which would allow better understanding of the behaviour and rheology of mud under reduced atmospheric pressure. This approach would allow direct comparison of modelled features with real martian landforms. This goal is beyond the scope of our present work.

5. Conclusions

Our experiments show that a warm (from ~292 K to ~296 K) and unconsolidated surface has a profound effect on the behaviour of flowing low viscosity mud in a low-pressure environment (Fig. 9). The resulting boiling occurring during the mass flow, causes transient levitation of the mud above the warm surface as well as the erosion of the unconsolidated sandy substrate. Both mechanisms alter mud propagation in a low pressure environment. Moreover, we expect that these processes would be even more effective on Mars which has a lower gravitational acceleration than on Earth. On Mars gravity does not change the boiling rate, but the sediments can be more easily entrained (Raack et al., 2017; Herny et al., 2018). The gas release should levitate mud for a longer period of time than observed in our experiments, as also similarly suggested for wet sand (Raack et al., 2017; Herny et al., 2018), hence allowing the mud to propagate over larger distances than modelled within the low pressure chamber. Additionally, as the process of evaporative cooling would remove the latent heat from the mud and from the surface over which it is propagating, at a certain point, the mud would start to freeze. This would cause the formation of a protective icy-muddy crust affecting the way the mud moves (Brož et al., in press) switching from levitation and sliding/creeping to the propagation via mud tubes. The mechanisms of mud propagation on Earth at different pressure-temperature conditions are well studied. Mud propagation, during this relatively simple process, is controlled by mass gravity flow. In contrast, very little is known about the modes controlling the same events on Mars and other bodies with or without atmospheres where mud eruptions may be present (e.g., Ruesch et al., 2019). Our new data demonstrate that the behaviour of mud and its propagation in a low-pressure environment, is strongly controlled by the surface temperature since freezing or rapid boiling give rise to different transport mechanisms than simple liquid flow. We conclude that mud eruption activity on other celestial bodies may produce profoundly different morphologies compared to those commonly observed on Earth.

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6. Figures and tables

463 Figure 1. Examples of putative martian low viscosity mud flows and their meter-sized terrestrial
464 counterparts. Panels a) and b) show two examples of putative kilometre-scaled martian water-dominated
465 mud flows within the Chryse Planitia (Komatsu et al., 2016, Brož et al., 2019). c) and d) show meter-
466 sized low viscosity mud flows within the crater of Bakhar mud volcano in Azerbaijan which, during
467 prolonged activity, are capable of building dozens of meter-sized morphological features. e) shows
468 mud volcanos within the area of Salse di Nirano in Italy and f) shows the scale and context of this feature
469 where other similar features are present. Panel a) (centred at 19.16°N, 322.73°E) and panel b) (centred
470 at 20.24°N, 324.01°E) are based on CTX images F05_037598_1988_XN_18N037W and
471 B19_016856_1990_XI_19N035W respectively, image credit NASA/JPL/MSSS. Panel f) based on
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Figure 2. Results of thermal model calculations described in the Section 2.2. Panel a) is the map of the maximum temperature the surface can be expected to attain in an average current martian climate. b) is the total percentage of a Mars Year that the surface is above 273.16 K. c) shows the maximum depth of the layer that experiences temperatures above 273.16 K. d) shows the maximum energy that can be released before the entire surface layer is cooled below 273.16 K. e) shows the time needed to cool the subsurface from the maximum temperature case (a) to below 273.16 K when the substrate top is forced to be at 273.16 K. f) shows the time needed until the heat flow conducted through the surface drops below 550 W/m² under the same conditions as in e).

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483 Figure 3: Schematic illustration showing the experimental setup with the position of thermocouples,
 484 photogrammetric targets and four cameras marked. Data from thermocouples are not discussed within
 485 this study, however, they are provided in the Supplementary materials uploaded on Zenodo.org for those
 486 who are interested.

Figure 4: A sequence of images from different time steps (in seconds) capturing the propagation of low viscosity mud over a warm surface inclined by 5° under low pressure mimicking conditions on Mars. The mud was poured on the surface from a hanging container. Once the mud touched the surface, intense boiling occurred, and a central crater-like depression started to form ($t=10$ s). Soon a surface mud flow developed ($t=22$ s) and started to propagate downslope a few centimetres at a time in the form of a narrow lobe which was levitating and sliding/creeping over the sand surface ($t=26$ s). After a while, the flow stopped its propagation ($t=30$ s) and a new lobe developed ($t=34$ s). This process repeated as long as mud was being poured onto the surface ($t=37$ s; $t=40$ s). The boiling also caused the formation of rims surrounding the lobes. The images were obtained from the video recorded by camera #1 observing experiment #31 from above.

Figure 5: Two-time sequences of images showing (a) the levitating and sliding mud over an unconsolidated surface and (b) the continuous explosions indicating the presence of liquid mud in the subsurface. Note the irregular shape of the propagating active mud flow in the vertical direction in panel (a) revealing intense boiling within the flow and associated release of water vapour from the flow. Such releases caused levitation of the material and rapid propagation of the flow over the surface. The images were obtained from the video recorded by camera #3 observing experiment #28 from front.

Figure 6: A time series of images (time steps in seconds) capturing the propagation of low viscosity mud over a warm surface inclined by 10° under low pressure, mimicking conditions on Mars. The mud was poured to the surface from a hanging container. Similarly to the results captured in Figure 3, intense boiling occurred ($t=8$ s) and after a while a narrow mud lobe developed ($t=9$ s). Due to the higher inclination of the surface the mud lobe travelled faster and further ($t=11$ s) than in those experiments performed at a slope of 5° . The narrow, long lobes were partly or fully covered by ejecting sandy grains ($t=25$ s) which caused a seemingly discontinuous appearance of the final morphology. The images were obtained from the video recorded by camera #1 observing experiment #26 from above.

Figure 7: An example of the resulting morphology of a low viscosity mud flow and its inner structure formed by the movement over “warm” sand in a low pressure environment. (a) The edge of the flow is surrounded by set of sandy ridges and several central troughs through which the mud propagated. The dashed line at b marks the position at which the mud flow was exposed by removing the sand (shown in panel b) and the dashed lines c to e mark where the flow was sectioned to reveal the inner structure of the flow (shown in the corresponding panels). The flow was composed of a layer in which clay-particles dominated and by a layer in which the sand was saturated by water (marked on panel d). In some cases, the liquid mud was still present in the subsurface (e). f) Detail of the resulting low viscosity mud flow morphology, which was characterised by a network of open central channels of varying depths surrounded by rims composed of sandy particles ejected from multiple small explosion sites. These explosion sites mark the boundary between the liquid mud and surrounding dry sand. In some places a protective crust developed by gluing together clay and sand particles. The width of the main channel is around 2 cm. Note the cliffs which can be vertical or overhangs can be formed.

Figure 8: A sequence of images taken at different time steps (t) capturing the propagation of low viscosity mud over a warm surface inclined by 5° under terrestrial ambient pressure conditions. The mud propagates as a tens of centimetres wide and a few mm thick sheet-like mud flow. Images obtained from the clip recorded by camera #1 observing the experiment #14 from above.

Figure 9: Schematic illustration showing the development of a low viscosity mud flow during the movement over the 'warm' inclined surface under martian conditions with insets showing the main

537 processes operating during its emplacement. (a) The ascending mud would move downhill from the
538 source area via partly overlapping flow lobes. The instability of water within the mud would cause
539 redeposition of unconsolidated sandy surface as well as fast propagation of the fronts of the active lobes.
540 The different colours mark different states of the mud; brown indicates immobile mud, while orange
541 indicates liquid, and hence mobile, mud. The black dashed rectangles mark the position of small insets
542 in panel **b** (b) **Stage 1** shows mud propagation by levitation over the inclined sand surface. The levitation
543 is caused by boiling water releasing gases from the base of the mixture. Escaping gases are additionally
544 able to trigger ejections of sand grains and hence cause self-burying of the mud flow under the surface.
545 **Stage 2** shows the situation when the mud flow lobe stops propagating by levitation and when the water
546 from the mud starts to infiltrate into the subsurface. **Stage 3** then shows the input of new liquid mud
547 flows over the older mud flow. Finally, at **Stage 4** the new batch of mud overcomes the margins of the
548 older mud flow and hence the mud flow would be again be exposed to a warm surface. This then causes
549 the levitation of the mud flow and repetition of the whole process.

Exp #	Pressure range* [mbar]	Inclination [°]	Release time** [s]	Type of surface
exp_31	7.1-8.1	5	52	~2 cm sand layer
exp_32	6.9-7.6	5	29	~2 cm sand layer
exp_33	6.9-7.4	5	21	~2 cm sand layer
exp_51	7.0-7.7	5	45	~33 cm sand layer
exp_52	6.6-7.1	5	34	~33 cm sand layer
exp_53	6.6-8.5	5	22	~33 cm sand layer
exp_12	1000	5	19	~2 cm sand layer
exp_13	1000	5	20	~2 cm sand layer
exp_14	1000	5	18	~2 cm sand layer
exp_26	6.7-7.4	10	36	~2 cm sand layer
exp_27	7.1-8.1	10	23	~2 cm sand layer
exp_28	7.1-7.8	10	43	~2 cm sand layer
exp_46	6.7-7.6	5	34	plastic plate
exp_47	6.9-7.3	5	50	plastic plate
exp_48	6.8-7.4	5	60	plastic plate
exp_43	1000	5	45	plastic plate
exp_44	1000	5	40	plastic plate
exp_45	1000	5	37	plastic plate

* Pressure range during the first two minutes of the experimental run

** Time period over which the mud was poured from the container

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551 Table 1: Summary of measured and controlled variables for each experimental run.

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