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1 **Experimental evidence for lava-like mud flows under Martian surface conditions**

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23 **Large outflow channels on ancient terrains of Mars have been interpreted as the**
24 **products of catastrophic flood events. The rapid burial of water-rich sediments**
25 **following such flooding could have led to sedimentary volcanism, in which mixtures of**
26 **sediment and water (mud) erupt to the surface. Tens of thousands of volcano-like**
27 **landforms populate the northern lowlands and other local sedimentary depocenters on**
28 **Mars. However, it is difficult to determine whether the edifices are related to igneous**
29 **or mud extrusions, partly because the behaviour of extruded mud under martian**
30 **surface conditions is poorly constrained. Here, we investigate the mechanisms of mud**
31 **propagation on Mars using experiments performed inside a low-pressure chamber at**
32 **cold temperatures. We find that low viscosity mud under martian conditions**
33 **propagates differently from on Earth, because of rapid freezing and the formation of**
34 **an icy crust. Instead, the experimental mud flows propagate like terrestrial pahoehoe**
35 **lava flows, with liquid mud spilling from ruptures in the frozen crust, then refreezing**
36 **to form a new flow lobe. We suggest that mud volcanism can explain the formation of**
37 **some lava-like flow morphologies on Mars, and that similar processes may apply to**
38 **cryovolcanic extrusions on icy bodies in the Solar System.**

39 The physics behind igneous volcanism on Mars is better understood [e.g., 1-4] than that
40 of sedimentary volcanism in which mixtures of water and sediment, subsequently referred
41 to as mud, are extruded onto the surface. On Earth, sedimentary volcanism manifests at the
42 surface as eruptions of fluids (water, gas, occasionally oil), fine grained sediments (e.g.
43 clays) and clasts from the country-rock. These geological phenomena are the result of fluid
44 (on Earth typically associated with methane) overpressure [5], generated at several hundred
45 to several thousand metres depth, combined with gravitational instability of buoyant
46 sedimentary units buried at deeper stratigraphic levels [6]. The viscosity of ascending mud
47 varies, and affects the shapes, sizes and thicknesses of resulting flows. The higher the water

48 content, the lower the viscosity and vice versa (Figure 1). The focus of our experimental
49 investigation are water-dominated mud flows that propagate over shallow slopes via
50 centimetre-thick flows (Fig. 1a,b), as opposed to clay-dominated flows that can be meter(s)
51 thick (Fig. 1c).

52 Although the propagation of water at low atmospheric pressure has been previously
53 studied [7-11], there is a lack of theoretical and empirical knowledge about the behaviour of
54 mud at low atmospheric pressure, temperature and gravity, despite an initial study by [12].
55 This knowledge gap represents an obstacle in the study of landforms interpreted to be the
56 result of mud extrusion on Mars [13-24] and other terrestrial or icy solar system bodies.
57 Currently, the low martian atmospheric pressure inhibits the sustained presence of liquid
58 water on the surface [e.g., 7-11,25,26], so evaporation and ice-formation cause the rheology
59 of the extruded mud to change rapidly; hence mud flows could propagate differently from
60 on Earth [17].

61 We used analogue experiments performed in a low pressure chamber to examine how low
62 viscosity, water-dominated mud with a solid fraction of less than ~6.5 wt.% (12.7 mPa.s at
63 276 K and 10.7 mPa.s at 296 K) propagates over a cold surface (244 K to 265 K) under
64 terrestrial and martian (7 mbar) atmospheric pressures. These experiments enabled
65 comparing flow mechanisms at different pressures to be compared and reveal a unique
66 propagation behaviours under martian conditions. Based on these observations we propose
67 that evaporation and surficial freezing would dominate the morphology for relatively thin
68 mud flows (< 1 m) and may influence thicker mud flows hypothesized to be present on the
69 martian surface [e.g. 12,15,17,24,25].

70 **Mud flow experiments**

71 We performed 21 experiments (Table S1 and Fig. S1 in Supplementary Information)
72 using the Open University (UK) Mars Chamber. During each experiment, 500 ml of mud
73 was poured over a 0.9×0.4 m aluminium tray containing either (a) a ~2 cm deep sand bed
74 (~63–200 μm grain diameter; 14 experiments) representing a sedimentary surface, or (b) a
75 plastic plate (7 experiments) representing an impermeable icy surface. Fifteen experiments
76 were performed at 7 ± 0.5 mbar and six experiments at ~1 bar (Table S1). The mud was
77 released onto the surface from a tilting container situated inside the chamber. This design
78 was chosen for its simplicity and reproducibility, although it represents a simplification of
79 the natural setting. At the beginning of the experiment the mud was above the freezing point
80 of water. The temperatures of the sand bed or plastic plate ranged from ~244 K to 265 K and
81 gradually increased with time as no active cooling of the experiment was performed. The
82 aluminium tray was inclined at 5° (18 experiments) or 10° (3 experiments) to force the mud
83 to move in a preferred direction. Each experimental run was performed at least in triplicate
84 and was recorded with four cameras. The experiments did not account for the effect of the
85 lower gravity on Mars as compared to Earth.

86 At the beginning of each experiment the atmospheric pressure was gradually reduced,
87 triggering the boiling of the water in the mud [25,27,28]. The mud within the container
88 cooled by evaporation [25] to almost its freezing point before being poured onto the surface
89 at a pressure of 7 ± 0.5 mbar (Table S1). Once in contact with the cold surface, the mud
90 rapidly began to freeze at the bottom and margins of the flow, and at its upper surface (Fig. 2a
91 and Supplementary Information). The freezing resulted in the formation of an ice-mud crust
92 which modified flow propagation and decreased lateral spreading (Fig. 3).

93 Mud propagation occurred through an intricate system of narrow flow lobes (Fig. 3b) or
94 several lobate flows (Fig. 3c and 3d). Their formation was controlled by the development of
95 frozen marginal ridges that confined the flow of liquid mud inside a central channel. As

96 freezing continued, ice crystals floated to the surface and started to merge. However, mud
97 still propagated within the crust via a network of “mud tubes”, in an analogous way to flow
98 within lava tubes (Fig. 2c). When new pulses of mud arrived, they caused breakouts and the
99 formation of further lobes (Fig. 2b). The newly extruded material rapidly developed a frozen
100 crust.

101 The presence of internal mud tubes was confirmed by sectioning the frozen mud flows
102 after the re-pressurization of the chamber. A liquid mud core was present even in the
103 experiments where mud was exposed to low pressure for several tens of minutes (Fig. 2d,e).
104 Vesicles ranging in size from 1 to 10 mm were observed within the crusts (Fig. 2d), produced
105 by vapour bubbles that did not escape. The vesicular nature of the crust inhibited the
106 conduction of heat [29] from the interior of the flow, increasing the depth to which vapour
107 bubble nucleation occurred.

108 **Comparison with terrestrial lava flows**

109 During flow formation, newly supplied mud was observed to increase the thickness of
110 lobes up to several centimetres (Fig. 3d) via lifting of the protective crust. This occurred
111 when the terminal part of the flow was frozen and the mud release was blocked, but newly
112 supplied mud was still intruding the lobe via mud tubes. This created overpressure within
113 the mud which was able to lift the crust of the lobes. Once a sufficient volume of mud had
114 accumulated within the lobe, the overpressure was able to break the crust and a new lobe
115 formed at the terminus. The mud flow inflated in a manner directly analogous to that of
116 pahoehoe lava flows [30]. This inflation was observed on 5° and 10° slopes, and for
117 impermeable and permeable substrates. Under terrestrial atmospheric conditions (i.e. room
118 pressure and temperature) a mud flow moving over a cold surface did not form lobes, did
119 not inflate, and had no icy crust (Fig. 3d) regardless of the temperature of the mud (~274,
120 ~290, and ~293 K were tested). Instead, the mud spread out over the surface in a broad sheet

121 only a few millimetres thick, and was in the liquid phase over the entire length of the flow
122 (Fig. 3a,d). Only minor freezing was observed in the form of ice crystals on the margins of
123 the flow after several minutes.

124 Like basaltic lavas on Earth, low viscosity mud flows produce laterally extensive
125 structures with lower relief than those resulting from high viscosity flows [6]. Because the
126 mud used in our experiments is water-dominated, it initially behaves as a Newtonian fluid.
127 In the low-pressure environment, evaporative cooling leads to the formation of ice crystals,
128 which increase the solid content forming a protective crust. The mud evolves into a non-
129 Newtonian fluid with non-zero yield strength as the total solid volume fraction increased
130 beyond ~15% [31]. This behaviour is similar to that of low viscosity basaltic lavas whose
131 movement is affected by the formation of an external crust, formed by solidification of the
132 lava due to cooling [32]. In both cases the strength of the visco-elastic part of the crust,
133 between the brittle outer part and the more fluid interior [30], is able to inhibit lateral
134 spreading and allows fluid accumulation and vertical inflation.

135 Over time, the water in the flow (i.e. not lost by evaporation) freezes to form ice in the
136 crust, and hence, the thin crust develops an ever-increasing yield strength due to the ice
137 crystal network becoming increasingly interconnected between the clay particles (see
138 rheology references in [33]). This creates a solid with mechanical strength similar to how
139 mineral crystal frameworks in a cooling silicate rock magma replace the rheological yield
140 strength of the lava. Thus the strength of the crust increases with time, and more inflation
141 should occur before breakouts occur, in agreement with our observation of more inflation in
142 the distal parts of the experimental flows. There is no reason to assume that the mechanical
143 strength of the crust would be different on Mars for mud with the same clay content. Since
144 the rate, as well as the mode, of its formation are controlled by the non-equilibrium
145 thermodynamics of vapour loss, specifically the transfer of sensible heat to latent heat in the

146 liquid surrounding each nucleating vapour bubble [25], the time-scale of crust development
147 should also be the same on Mars. Calculations describing these processes are reported in the
148 Supplementary Information.

149 The above arguments assume that, as in the experiments, the motion of the mud is laminar.
150 However, it is likely that in large flows, that the fluid motion is turbulent. For steady, uniform
151 laminar flows of mud with viscosities of 0.01, 0.1, 1 and 10 Pa.s on 0.6° slopes under martian
152 gravity, the transition to turbulence will occur for mud flows with thicknesses of 17 mm,
153 8 cm, 36 cm and 1.7 m, respectively (see equations in [12]). As the solid content (silicate
154 particles and ice crystals) of the mud increases, its viscosity increases, but even for a total
155 solids content of 60% [e.g. 34] the viscosity does not exceed 1 Pa.s. Thus any low viscosity
156 mud flow thicker than a few tens of centimetres on Mars will probably be turbulent and we
157 discuss the applicability of our experiments to this case below.

158 **Implications for mud flows on planetary surfaces**

159 In a model of mud flow dynamics on Mars, [12] considered the effect of high solid content
160 on the non-Newtonian rheology of mud, assuming the rheological properties were constant
161 everywhere along the flow. Our experiments underline the importance of considering the
162 thermodynamics of the processes that occur when mud is exposed to the martian
163 environment. In a turbulent mud flow, mud from all depths will be exposed to the low
164 atmospheric pressure and will boil, lose vapour, and cool. Thus, it will cool rapidly,
165 increasing the ice crystal content and evolving a yield strength. As the yield strength
166 increases, the critical Reynolds number required to sustain turbulence increases [35], but the
167 associated increase in viscosity causes the actual Reynolds number to decrease. When the
168 two become equal this forces a transition to laminar flow, encouraging the formation of lobes
169 and breakouts, as observed in our experiments.

170 Although our small-scale experimental mudflows show characteristics similar to those of
171 low-viscosity lava flows, they have a different heat loss mechanism. For lava flows simulated
172 in the laboratory using wax [e.g., 36-37] a modified Péclet number has been used to
173 distinguish flows that do or do not readily form crusts. This is not appropriate for our flows,
174 where the heat loss mechanism is decompressional boiling, not conduction and convection
175 into the environment (for details see Supplementary Information).

176 On Earth, low effusion rates and volumes can lead to the formation of mud flows
177 (Fig. 1a,b) with similar magnitudes to those in our experiments. Therefore, such terrestrial
178 flows exported to Mars would be strongly affected by the mechanisms observed in our
179 experiments. We expect these mechanisms to also be important at the margins of kilometre-
180 scale flows where their thicknesses decrease and numerous overlapping centimetre-thick
181 flows occur. Once mud is extruded into the martian surface pressure environment, it would
182 cool due to evaporative cooling (Fig. 4a) and freezing of the flow surface would eventually
183 cause a change in the flow regime. From that point, the mud would no longer propagate via
184 open channel(s) (Fig. 4b), but instead via mud tube(s) (Fig. 4c) and lobes (Fig. 4d). The
185 distance such a transition occurs from the source would depend on the effusion rates,
186 volumes, viscosity and temperature of the extruded mud.

187 Our study shows that inflation of decimetre-thick and meter-sized lobes could occur at
188 the margins of both mud flows and lava flows, thus both igneous and mud volcanism surface
189 flows could have similar morphological characteristics [e.g., 12,17]. Our calculations show
190 that the morphologies of mud flows at scales larger than those covered by our experiments
191 could be affected by the same processes, particularly at their margins. Hence, as their
192 mechanisms of mud propagation would differ, martian mud volcanoes may be substantially
193 different in shape from terrestrial ones [17].

194 Sedimentary volcanism has also been proposed for the dwarf planet Ceres [38,39], which
195 may have a water-muddy ocean beneath a crust made of clays, salts, clathrates and ice
196 [39,40]. The process of evaporative cooling and associated freezing should also occur there,
197 affecting the morphologies of resulting effusive landforms; even more so on Ceres than on
198 Mars as Ceres lacks an atmosphere. The same principles would apply to other Solar System
199 bodies and icy moons, and so our experimental results should be considered when
200 interpreting effusive cryovolcanic surface features on these bodies [e.g., 41,42]. Our results
201 show that it is vital to consider the effects of the differing environmental conditions on other
202 planetary surfaces when comparing analogue landforms observed on Earth with apparently
203 similar effusive morphologies on other bodies.

204 **References**

- 205 [1] Wilson, L., & Head, J. W. Mars: Review and analysis of volcanic eruption theory and
206 relationships to observed landforms. *Reviews of Geophysics* 32(3), 221–263 (1994).
207 <https://doi.org/10.1029/94RG01113>
- 208 [2] Brož, P., Čadek, O., Hauber, E., & Rossi, A. P. Scoria cones on Mars: Detailed investigation
209 of morphometry based on high-resolution digital elevation models. *Journal of*
210 *Geophysical Research: Planets* 120, 1512–1527 (2015).
211 <https://doi.org/10.1002/2015JE004873>
- 212 [3] Parfitt, E. A., & Wilson L. *Fundamentals of Physical Volcanology* (Blackwell, Oxford, U.
213 K., 2008). ISBN: 978-0-632-05443-5
- 214 [4] Fagents, S. A, Gregg, T. K. P., & Lopes, R. M. C. *Modelling volcanic processes: the physics*
215 *and mathematics of volcanism* (Cambridge University Press, Cambridge, 2013). ISBN:
216 9781139021562

- 217 [5] Dimitrov, L. I. Mud volcanoes—the most important pathway for degassing deeply buried
218 sediments. *Earth-Science Reviews* 59, 1–4, 49–76 (2002). <https://doi.org/10.1016/S0012->
219 8252(02)00069-7
- 220 [6] Mazzini, A., & Etiope, G. Mud volcanism: An updated review. *Earth-Science Reviews* 168,
221 81–112 (2017). <https://doi.org/10.1016/j.earscirev.2017.03.001>
- 222 [7] Wallace, D., & Sagan, C. Evaporation of ice in planetary atmospheres: Ice-covered rivers
223 on Mars. *Icarus* 39, 385–400 (1979). [https://doi.org/10.1016/0019-1035\(79\)90148-9](https://doi.org/10.1016/0019-1035(79)90148-9)
- 224 [8] Carr, M. H. Stability of streams and lakes on Mars. *Icarus* 56, 476–495 (1983).
225 [https://doi.org/10.1016/0019-1035\(83\)90168-9](https://doi.org/10.1016/0019-1035(83)90168-9)
- 226 [9] Baker, V. R. Erosional processes in channelized water flows on Mars. *Journal of*
227 *Geophysical Research* 84, 7985–7993 (1979). <https://doi.org/10.1029/JB084iB14p07985>
- 228 [10] Brass, G.W. The stability of brines on Mars. *Icarus* 42, 20–28 (1980).
229 [https://doi.org/10.1016/0019-1035\(80\)90237-7](https://doi.org/10.1016/0019-1035(80)90237-7)
- 230 [11] Kossacki, K. J., Markiewicz, W. J., Smith, M. D., Page, D., & Murray, J. Possible remnants
231 of a frozen mud lake in southern Elysium, Mars. *Icarus* 181, 363–374 (2006).
232 <https://doi.org/10.1016/j.icarus.2005.11.018>
- 233 [12] Wilson, L., & Mougini-Mark, P. J. Dynamics of a fluid flow on Mars: Lava or mud?
234 *Icarus* 233, 268–280 (2014). <https://doi.org/10.1016/j.icarus.2014.01.041>
- 235 [13] Oehler, D.Z. & Allen, C.C. Evidence for pervasive mud volcanism in Acidalia Planitia,
236 Mars. *Icarus* 208, 636–657 (2010). <https://doi.org/10.1016/j.icarus.2010.03.031>
- 237 [14] Allen, C.C. et al. Fluid expulsion in terrestrial sedimentary basins: a process providing
238 potential analogs for giant polygons and mounds in the martian lowlands. *Icarus* 224,
239 424–432 (2013). <https://doi.org/10.1016/j.icarus.2012.09.018>

- 240 [15] Komatsu, G. et al. Small edifice features in Chryse Planitia, Mars: Assessment of a mud
241 volcano hypothesis. *Icarus* 268, 56–75 (2016).
242 <https://doi.org/10.1016/j.icarus.2015.12.032>
- 243 [16] Hemmi, R., & Miyamoto, H. High-resolution topographic analyses of mounds in southern
244 Acidalia Planitia, Mars: Implications for possible mud volcanism in submarine and
245 subaerial environments. *Geosciences*, 8(5), 152, 1–19 (2018).
246 <https://doi.org/10.3390/geosciences8050152>
- 247 [17] Brož, P., Hauber, E., van de Burgt, I., Špillar, V., & Michael, G. Subsurface sediment
248 mobilization in the southern Chryse Planitia on Mars. *Journal of Geophysical Research:*
249 *Planets* 124, 703–720 (2019). <https://doi.org/10.1029/2018JE005868>
- 250 [18] Okubo, C. H. Morphologic evidence of subsurface sediment mobilization and mud
251 volcanism in Candor and Coprates Chasmata, Valles Marineris, Mars. *Icarus* 269, 23–37
252 (2016). <https://doi.org/10.1016/j.icarus.2015.12.051>
- 253 [19] Skinner, J. A., & Mazzini, A. Martian mud volcanism: Terrestrial analogs and implications
254 for formational scenarios. *Marine and Petroleum Geology* 26(9), 1866–1878 (2009).
255 <https://doi.org/10.1016/j.marpetgeo.2009.02.006>
- 256 [20] Hemmi, R. & Miyamoto, H. Distribution, morphology, and morphometry of circular
257 mounds in the elongated basin of northern Terra Sirenum, Mars. *Progress in Earth and*
258 *Planetary Science* 4, 26, 1–15 (2017). <https://doi.org/10.1186/s40645-017-0141-x>
- 259 [21] Kumar, P. S. et al. Recent seismicity in Valles Marineris, Mars: Insights from young faults,
260 landslides, boulder falls and possible mud volcanoes, *Earth and Planetary Science Letters*
261 505, 51–64 (2019). <https://doi.org/10.1016/j.epsl.2018.10.008>

- 262 [22] Brož, P., & Hauber, E. Hydrovolcanic tuff rings and cones as indicators for
263 phreatomagmatic explosive eruptions on Mars. *Journal of Geophysical Research: Planets*
264 118, 1656–1675 (2013). <https://doi.org/10.1002/jgre.20120>
- 265 [23] Brož, P., Hauber, E., Wray, J. J., & Michael, G. Amazonian volcanism inside Valles
266 Marineris on Mars. *Earth and Planetary Science Letters* 473, 122–130 (2017).
267 <https://doi.org/10.1016/j.epsl.2017.06.003>
- 268 [24] Skinner, J. A., & Tanaka, K. L. Evidence for and implications of sedimentary diapirism
269 and mud volcanism in the southern Utopia highland-lowland boundary plain, Mars. *Icarus*
270 186(1), 41–59 (2007). <https://doi.org/10.1016/j.icarus.2006.08.013>
- 271 [25] Bargery, A. S., Lane, S. J., Barrett, A., Wilson, L., & Gilbert, J. S. The initial responses of
272 hot liquid water released under low atmospheric pressures: Experimental insights. *Icarus*
273 210(1), 488–506 (2010). <https://doi.org/10.1016/j.icarus.2010.06.019>
- 274 [26] Hecht, M. H. Metastability of liquid water on Mars. *Icarus* 156(2), 373–386 (2002).
275 <https://doi.org/10.1006/icar.2001.6794>
- 276 [27] Raack, J. et al. Water-induced sediment levitation enhances downslope transport on Mars.
277 *Nature Communications* 8, 1–10 (2017). <https://doi.org/10.1038/s41467-017-01213-z>
- 278 [28] Herny, C. et al. Downslope sediment transport by boiling liquid water under Mars-like
279 conditions: Experiments and potential implications for Martian gullies. *Geological*
280 *Society Special Publication* 467, 373–410 (2019). <https://doi.org/10.1144/SP467.10>
- 281 [29] Smith, D. et al. Thermal conductivity of porous materials. *Journal of Materials Research*
282 28(17), 2260–2272 (2013). <https://doi.org/10.1557/jmr.2013.179>
- 283 [30] Hon, K., Kauahikaua, J., Denlinger, R., & Mackay, K. Emplacement and inflation of
284 pahoehoe sheet flows: observation and measurements of active lava flows on Kilauea

- 285 Volcano, Hawaii. *Geological Society of America Bulletin* 106, 351–370 (1994).
286 [https://doi.org/10.1130/0016-7606\(1994\)106<0351:EAIOPS>2.3.CO;2](https://doi.org/10.1130/0016-7606(1994)106<0351:EAIOPS>2.3.CO;2)
- 287 [31] Ayel, V., Lottin, O., & Peerhossaini, H. Rheology, flow behaviour and heat transfer of ice
288 slurries: a review of the state of the art. *International Journal of Refrigeration* 26, 95–107
289 (2003). [https://doi.org/10.1016/S0140-7007\(02\)00016-6](https://doi.org/10.1016/S0140-7007(02)00016-6)
- 290 [32] Cashman, K. V., Kerr, R. C., & Griffiths, R. W. A laboratory model of surface crust
291 formation and disruption on lava flows through non-uniform channels. *Bulletin of*
292 *Volcanology* 68(7-8), 753–770 (2006). <https://doi.org/10.1007/s00445-005-0048-z>
- 293 [33] Chevrel M. O., Baratoux, D., Hess, K.-U., & Dingwell, D. B. Viscous flow behavior of
294 tholeiitic and alkaline Fe-rich martian basalts. *Geochimica et Cosmochimica Acta* 124,
295 348–365 (2014). <https://doi.org/10.1016/j.gca.2013.08.026>
- 296 [34] Kaitna, R., Rickenmann, D. & Schatzmann, M., Experimental study on rheologic
297 behaviour of debris flow material. *Acta Geotechnica* 2, 71–85 (2007).
298 <https://doi.org/10.1007/s11440-007-0026-z>
- 299 [35] Skelland, A.H.P. *Non-Newtonian flow and heat transfer*. (John Wiley and Sons, New
300 York, 1967).
- 301 [36] Rader, E., Vanderkluyesen, L., & Clarke, A. The role of unsteady effusion rates on
302 inflation in long-lived lava flow fields. *Earth and Planetary Science Letters* 477, 73–
303 83 (2017). <https://doi.org/10.1016/j.epsl.2017.08.016>
- 304 [37] Fink, J. H., & Griffiths, R. W. A laboratory analog study of the morphology of lava
305 flows extruded from point and line sources. *Journal of Volcanology and Geothermal*
306 *Research* 54, 19–32 (1992). [https://doi.org/10.1016/0377-0273\(92\)90112-Q](https://doi.org/10.1016/0377-0273(92)90112-Q)

- 307 [38] Sori, M. M. et al. The vanishing cryovolcanoes of Ceres. *Geophysical Research Letters* 44,
308 1243–1250 (2017). <https://doi.org/10.1002/2016GL072319>
- 309 [39] Ruesch, O. et al. Slurry extrusion on Ceres from a convective mud-bearing mantle. *Nature*
310 *Geoscience* 12, 505–509 (2019). <https://doi.org/10.1038/s41561-019-0378-7>.
- 311 [40] Marchi, S. et al. An aqueously altered carbon-rich Ceres. *Nature Astronomy* 3, 140–145
312 (2018). <https://doi.org/10.1038/s41550-018-0656-0>.
- 313 [41] Allison, M. L. & Clifford, S. M. Ice-covered water volcanism on Ganymede. *Journal of*
314 *Geophysical Research* 92, 7865–7876 (1987). <https://doi.org/10.1029/JB092iB08p07865>
- 315 [42] Fagents, S.A. Considerations for effusive cryovolcanism on Europa: The post-Galileo
316 perspective. *Journal of Geophysical Research: Planets* 108 (E12), 5139 (2003).
317 <https://doi.org/10.1029/2003JE002128>

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331 **Author contribution**

332 The experimental set-up and the methodology were conceived and designed by P.B. and O.K.
333 with the help and advice of S.J.C., J.R., M.R.P., M.R.B., A.M., and E.H. The technical support
334 was provided by M.R.S. The data analysis was done by P.B. with significant feedback from
335 O.K., L.W., S.J.C., E.H. and A.M. The DEM production was done by O.K. and the theoretical
336 considerations associated with scaling were done by L.W. All authors contributed to discussion,
337 interpretation and writing of the manuscript.

338 **Competing interests**

339 The authors declare no competing interests.

340 **Figure captions**

341 **Fig. 1. Examples of surface expressions of terrestrial sedimentary volcanism caused by**
342 **muds of various viscosity.** (a) A water-dominated mud flowing from the crater of Bakhar mud
343 volcano in Azerbaijan (39°59'55.7"N, 49°28'29.9"E). (b) An individual mud flow from a
344 'gryphon' on top of Dashgil volcano in Azerbaijan (39°59'48"N, 49°24'11"E). (b). (c) A
345 kilometre-sized highly-viscous mud flow outgoing from Koturdag mud volcano in Azerbaijan
346 (39°58'30"N, 49°21'36"E). Note minibus for scale.

347 **Fig. 2. Examples of morphologies and interior structures of mud flows formed in a low-**
348 **pressure environment.** Panel **a** shows three frames from video taken by Cam #2 covering
349 experiment #16 in which the formation of a narrow, thick mud flow occurred. Panel **b** shows in
350 detail the formation of the icy crust and outbreaks of new mud pulses from beneath the icy-
351 muddy crust. When the resulting mud flow features were sectioned, large cavities filled with
352 liquid mud (**c**), or voids (**d**) in the ice were observed. A liquid mud core was commonly
353 observed in the flow interior (**e**).

354 Fig. 3. **Timeline maps of modelled mud flows derived from the videos and final**
355 **topographic cross sections.** Flows formed in terrestrial (a) and low-pressure environments
356 when mud was poured by high (b) or low (c) release rates. The numbered bold lines represent
357 the margins of the flows at 5 s intervals, the other lines are 1 s intervals. (d) The position of
358 each topographic profile is marked in panels a, b and c. Topography was measured after the
359 chamber was re-pressurized.

360 Fig. 4. **Hypothesised development of a low viscosity mud flow on Mars.** (a) The mud
361 boils and self-cools through evaporative cooling. Once a mud flow develops, the mud is
362 transported via an open channel (b) evolving into a mud tube due to freezing of the flow surface
363 (c). The mud propagates via mud tubes to the flow front. As the crust prevents free movement
364 of the mud, it spreads via lobe breakout (d) which occurs when the pressure inside the frozen
365 flow is large enough to break the crust or to lift it up; exposing the mud again to the martian
366 environment.

367 **Methods**

368 **Experimental setup.** The low viscosity mud used in our experiments was a mixture of water
369 that contains 0.5 % w/w dissolved magnesium sulphate (MgSO_4) salts corresponding to the
370 average river water salinity and clay obtained from the claystone named after the Rokle locality
371 operated by the company Keramost, which is situated near the town of Kadaň in the Czech
372 Republic. The clay was a bentonite composed of 76 % montmorillonite, 23 % illite, and 1 %
373 kaolinite which has been formed by alteration of pyroclastic rocks. As there is no direct in-situ
374 knowledge of which types of clays could be present on Mars during subsurface sediment
375 mobilization and as explosive volcanism was once present on Mars [42], to the first
376 approximation this material seems to be a suitable analogue. To exclude the presence of
377 potentially more lithified clayey aggregates, the clay was mixed with water and salt and
378 homogenized in a blender for 3 minutes. Adding the small amount of the salt was necessary to

379 allow submillimetre particles to get into suspension within the mixture [43] and also realistic
380 for the martian surface [e.g., 44-46]. The resulting viscosity of the mud was 12.7 mPa.s at ~276
381 K and 10.7 mPa.s at ~296 K and the average density of the mixture was 1037.5 kg.m⁻³. The
382 viscosity was measured with Haake Rotovisco RV 20 and Viscotester VT 550 rheometers
383 (Institute of Hydrodynamics of the Czech Academy of Sciences) with ledges on the MV2
384 cylinder to prevent slip of the measured material on its walls.

385 Each experimental run (for details see Table S1 in the Supplementary Information) started
386 by inserting the 0.9 × 0.4 m aluminium tray filled with (a) a ~2 cm deep substrate bed (natural
387 sand, ~200 µm), or b) a plastic plate, and similarly sized copper plate inside the freezer to
388 pre-cool the tray and the plate to temperatures around 238 K. Once the required temperature
389 was reached, the plate and the tray were inserted inside the vacuum chamber. At the same
390 time the 500 ml of liquid mud was poured inside the tilting container equipped with one
391 thermocouple to record the temperature of the mud, and the container was installed inside
392 the chamber. The temperature of the mud varied from 274 to 297 K before the pressure drop.
393 The temperature within the chamber was also monitored by another thermocouple.
394 Additionally, five thermocouples were set in a grid (see Fig. 1 in the Supplementary
395 Information for details about the positioning of thermocouples) within the tray in order to
396 monitor the temperature of the surface over which the mud propagated.

397 Once the tray was in place inside the chamber, a series of images were taken by a single-
398 lens reflex camera from different angles to obtain the digital elevation model of the pristine
399 surface before the experimental run. Subsequently the chamber was closed and the process
400 of depressurization started. To achieve the pressure drop from ambient terrestrial pressure to
401 7 mbar took usually around 6 minutes. Once the pressure started to drop, the decrease in the
402 temperature of the mud within the container was measured. Every time the mud self-cooled
403 close to 273 K during the pressure drop, but it remained liquid. When the pressure of ~7 mbar

404 was reached, the container was manually flipped by the operator and hence mud was poured
405 from the height of ~5 cm to the surface. The mud flux was not directly measured, we recorded
406 only the time for how long the mud was extruded (Table S1), as the intense boiling occurred
407 within the container and hence caused irregularities within the flux.

408 The movement of the mud over the surface was recorded by four video cameras. Once the
409 mud propagation stopped, the resulting mud flow feature was left in the low pressure
410 environment for various lengths of time ranging from several minutes to about one hour. After
411 that the process of re-pressurization of the chamber to terrestrial values started, typically
412 before temperature of the tray surface rose above ~273 K. Once the pressure inside the chamber
413 reached atmospheric pressure, the chamber was opened and the resulting flow features were
414 documented by taking images from different angles to acquire data for subsequent DEM
415 production. Ultimately the mud flows produced were sectioned and their inner structure was
416 investigated and documented.

417 **DEM production.** To compare the elevation profiles along and perpendicular to the flow
418 directions of the mud flows we calculated a series of digital elevation models (DEM). The
419 sedimentary bed was photographed after each experimental procedure ~30-70 times from
420 multiple viewpoints. The reconstruction of a 3D model surface was produced by using the
421 ‘Structure-from-Motion’ [47] commercial software Agisoft PhotoScan. For image orientation
422 correlation and scaling of the 3D models we used twelve fixed black-on-white printed markers
423 which were affixed onto flat topped cylindrical posts. The posts had two different elevations
424 (4.6 cm and 9.6 cm) and the markers were ~2.67 cm in diameter. Typical discrepancies between
425 actual and calculated marker positions were ~0.8 - 1.6 mm. Exported DEMs and orthophotos
426 (TIFF format) were imported to QGIS for further analysis and production of the elevation
427 profiles.

428 **Cooling and freezing of mud when the external pressure is less than the saturation**
 429 **vapour pressure of the water.** Both the silicate component and the water that has not yet
 430 vaporized will cool from the initial temperature until the freezing point is reached. After the
 431 freezing point is reached, the temperature remains constant and vapour continues to be lost until
 432 all of the remaining water has been converted to ice. If the external pressure is less than the
 433 saturation vapour pressure of the ice, evaporation continues, the frozen mud surface cools, and
 434 a wave of cooling propagates into the frozen mud. The bulk density of the mud changes
 435 continuously throughout these stages as a function of the initial mass fractions of silicate and
 436 water in the mud. Let the initial masses of water and silicate in a given sample of the mud be
 437 m_w and m_s , respectively, the volumes of water and solid be v_w and v_s , respectively, and the
 438 corresponding densities be ρ_w and ρ_s . Let the bulk density of the mud be β . Then

439

$$440 \quad \beta = (m_w + m_s) / (v_w + v_s) = (m_w + m_s) / [(m_w / \rho_w) + (m_s / \rho_s)] \quad (1).$$

441

442 Expanding, collecting terms and simplifying:

443

$$444 \quad m_w / m_s = (\rho_w / \rho_s) [(\rho_s - \beta) / (\beta - \rho_w)] \quad (2).$$

445

446 We next find the mass of vapour, M_{vc} , that must be lost from the water to cool the
 447 remaining water and silicate to any given lower temperature. Let the initial mud temperature be
 448 θ_i and the final temperature be θ_f . The heat H_v removed from the water by the formation of the
 449 vapour is

$$451 \quad H_v = M_{vc} L_v \quad (3)$$

452

453 where L_v is the latent heat of vapourization, $2.46 \times 10^6 \text{ J kg}^{-1}$. During this cooling process, the
454 liquid water mass decreases from its initial value m_w to a smaller final value m_f , where by
455 definition

456

$$457 \quad m_f = m_w - M_{vc} \quad (4).$$

458

459 Thus, the average mass of water, m_a , during the cooling process is

460

$$461 \quad m_a = 0.5 (m_w + m_f) = 0.5 (m_w + m_w - M_{vc}) = (m_w - 0.5 M_{vc}) \quad (5).$$

462

463 As long as the specific heat of the water can be approximated as a constant, the heat the water
464 loses while cooling is H_c where

465

$$466 \quad H_c = m_a C_w (\theta_i - \theta_f) \quad (6)$$

467

468 and C_w is the specific heat of water, $4186 \text{ J kg}^{-1} \text{ K}^{-1}$. The silicate mass m_s also cools, and loses
469 an amount of heat equal to H_s where

470

$$471 \quad H_s = m_s C_s (\theta_i - \theta_f) \quad (7).$$

472

473 Here C_s is the specific heat of the silicate, say $1000 \text{ J kg}^{-1} \text{ K}^{-1}$. Equating the sum of H_c and H_s
474 to H_v ,

475

$$476 \quad (m_w - 0.5 M_{vc}) C_w (\theta_i - \theta_f) + m_s C_s (\theta_i - \theta_f) = M_{vc} H_v \quad (8)$$

477

478 and regrouping,

479

$$480 \quad M_{vc} [H_v + 0.5 C_w (\theta_i - \theta_f)] = (m_w C_w + m_s C_s) (\theta_i - \theta_f) \quad (9)$$

481

482 or

483

$$484 \quad M_{vc} / m_s = \{[(m_w / m_s) C_w + C_s] (\theta_i - \theta_f)\} / [H_v + 0.5 C_w (\theta_i - \theta_f)] \quad (10).$$

485

486 Using equation (2) for (m_w / m_s) we can also find (M_{vc} / m_s) as a function of the assumed value

487 of ρ_s . Finally, the ratio, R, of the mass of water converted to vapour to the initial water mass,

488 i.e. $R = (M_{vc} / m_w)$, is equal to $[(M_{vc} / m_s) (m_s / m_w)]$ or more conveniently

489

$$490 \quad R = [(M_{vc} / m_s) / (m_w / m_s)] \quad (11).$$

491

492 The mass of water remaining in the mud after the cooling phase, m_f , is therefore

493

$$494 \quad m_f = (1 - R) m_w \quad (12).$$

495

496 The above equations apply between any pair of temperatures θ_i and θ_f until θ_f becomes

497 equal to the freezing point, θ_{fr} . After the freezing point is reached, the temperature remains

498 constant while liquid water continues to evaporate and the latent heat of vaporization is

499 extracted from the remaining water, progressively freezing into ice all of the water that is not

500 lost as vapour. The latent heat of vaporization in the 283-293 Kelvin range is $2.46 \times 10^6 \text{ J kg}^{-1}$

501 and the latent heat of solidification is $3.34 \times 10^5 \text{ J kg}^{-1}$. The ratio of these is $Q = (3.34 \times 10^5 /$

502 $2.46 \times 10^6) = 0.13577$. Thus, to produce 1 kg of ice we would have to evaporate 0.13577 kg of

503 water into vapour from an initial total mass of 1.13577 kg of water. A fraction $[Q/(1 + Q)]$ of
 504 the water mass remaining after cooling must become vapour and a fraction $[1/(1 + Q)]$ of the
 505 water mass remaining after cooling becomes ice. The final ice mass, m_i , is

$$506$$

$$507 \quad m_i = [1/(1 + Q)] [(1 - R) m_w] \quad (13)$$

508

509 and the mass of water converted to vapour during the freezing phase is M_{vf} where

$$510$$

$$511 \quad M_{vf} = [Q/(1 + Q)] (1 - R) m_w \quad (14).$$

512

513 The total mass of vapour generated by the whole process is $M_v = M_{vc} + M_{vf}$.

514

515 As a result of the loss of vapour, the bulk density of the frozen mud will be different from
 516 the density of the initial mixture. The ice has a density ρ_i of 916.8 kg m^{-3} so the mass m_i of ice
 517 has a volume of $v_i = (m_i / \rho_i)$. The silicate volume is still v_s and so the final bulk density is β_f
 518 where

$$519$$

$$520 \quad \beta_f = (m_i + m_s) / (v_i + v_s) = (m_i + m_s) / [(m_i / \rho_i) + (m_s / \rho_s)] \quad (15).$$

521

522 Equation (12) gives (m_i / m_w) and equation (2) gives (m_w / m_s) so in terms of these,

$$523$$

$$524 \quad \beta_f = [(m_i / m_w) + (m_w / m_s)^{-1}] / \{[(m_i / m_w) / \rho_i] + [\rho_s (m_w / m_s)]^{-1}\} \quad (16).$$

525

526 The density, ρ_s , of the clay minerals in the experimental mud was $\sim 2500 \text{ kg m}^{-3}$ and the bulk
 527 density of the mud was $\sim 1040 \text{ kg m}^{-3}$, implying that the clay component formed $\sim 6.5 \%$ of

528 the mud mass. The mud was released into ambient experimental chamber pressures in the
529 range 650-700 Pa, and initial mud temperatures were up to ~275.5 K. Taking account of the
530 weight of the overlying mud and the experimental chamber pressure, the saturation vapour
531 pressure of water, with values up to ~730 Pa [48], would have been reached at depths up to
532 5-6 mm in the experiments. The relative values of the specific heat and the latent heat of
533 evaporation of water are such that while the mud was cooling from ~275.5 K to its freezing
534 point, ~0.34 % of its water would have been lost, having a trivially small effect on its
535 essentially Newtonian rheology. By the time subsequent vapour loss had frozen the
536 remaining water, ~88 % of the initial mass of water would have been converted to ice at the
537 expense of losing ~12 % of the initial water mass as vapour, leaving solid mud with a density
538 of ~961 kg m⁻³, slightly less dense than the original liquid mud. This should have produced
539 a frozen crust, again of thickness 5-6 mm, beneath which the mud would have been partially
540 liquid, as observed in the experiments. In similar scale flows under martian gravity, the
541 thermodynamics of this process would have been the same, as it involves only heat transfer
542 by conduction, but the thickness of the frozen outer crust would have been greater, 14-17
543 mm, because the pressure in the mud depends on the acceleration due to gravity [49].

544 **Data availability**

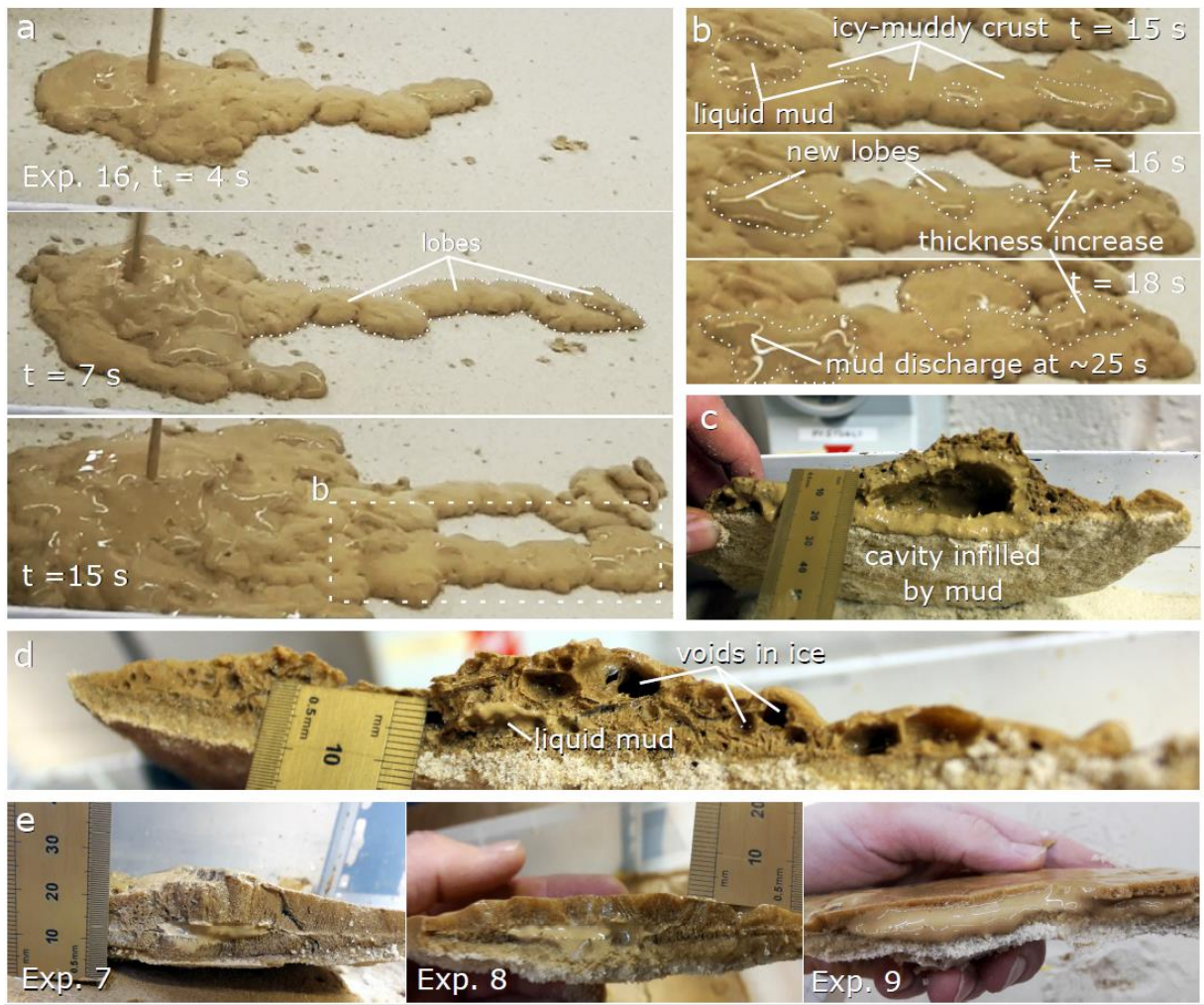
545 The movies, photos, pressure and temperature logs generated during and analysed during the
546 current study that support our findings are available in the Zenodo repository with the identifier
547 DOI: 10.5281/zenodo.3457148 (<https://doi.org/10.5281/zenodo.3457148>).

548 **References**

549 [42] Platz, T., Byrne, P. K., Massironi, M. & Hiesinger, H. Volcanism and tectonism across the
550 inner solar system: an overview. Geological Society, London, Special Publications 401,
551 1–56 (2014).

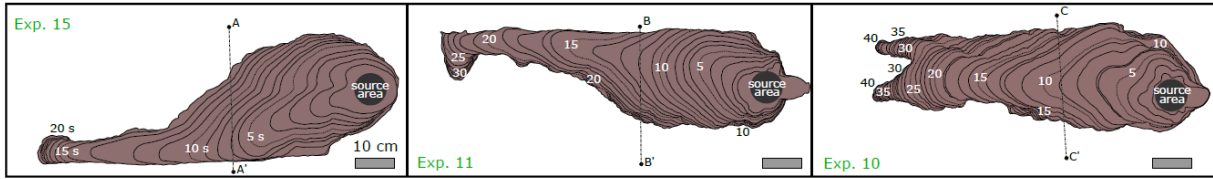
- 552 [43] Corradi, A. B., Manfredini T., Pellacani, G. C. & Pozzi, P. Deflocculation of Concentrated
553 Aqueous Clay Suspensions with Sodium Polymethacrylates. *Journal of the American*
554 *Ceramic Society* 77, 2, 509–513 (1994). [https://doi.org/10.1111/j.1151-](https://doi.org/10.1111/j.1151-2916.1994.tb07022.x)
555 [2916.1994.tb07022.x](https://doi.org/10.1111/j.1151-2916.1994.tb07022.x)
- 556 [44] Clark, B. C. Implications of abundant hygroscopic minerals in the martian regolith. *Icarus*
557 34, 645–665 (1978). [https://doi.org/10.1016/0019-1035\(78\)90052-0](https://doi.org/10.1016/0019-1035(78)90052-0)
- 558 [45] Vaniman, D. T. et al. Magnesium sulphate salts and the history of water on Mars. *Nature*
559 431, 663–665 (2004). <https://doi.org/10.1038/nature02973>
- 560 [46] Hecht, M. H., et al. Detection of perchlorate and the soluble chemistry of martian soil at
561 the Phoenix lander site. *Science* 325, 64–67 (2009).
562 <https://doi.org/10.1126/science.1172466>
- 563 [47] Westoby, M. J., Brasington, J., Glasser, N. F., Hambrey, M. J. & Reynolds, J. M.
564 ‘Structure-from-motion’ photogrammetry: a low-cost, effective tool for geoscience
565 applications. *Geomorphology* 179, 300–314 (2012).
566 <https://doi.org/10.1016/j.geomorph.2012.08.021>
- 567 [48] Lide, D. R. (ed.). *CRC Handbook of Chemistry and Physics* (85th edition, CRC Press,
568 Boca Raton, Florida, 2004).
- 569 [49] Pedersen, G. B. M. Frozen Martian lahars? Evaluation of morphology, degradation and
570 geologic development in the Utopia-Elysium transition zone. *Planetary and Space*
571 *Science* 85, 59–77 (2013). <https://doi.org/10.1016/j.pss.2013.05.020>

572

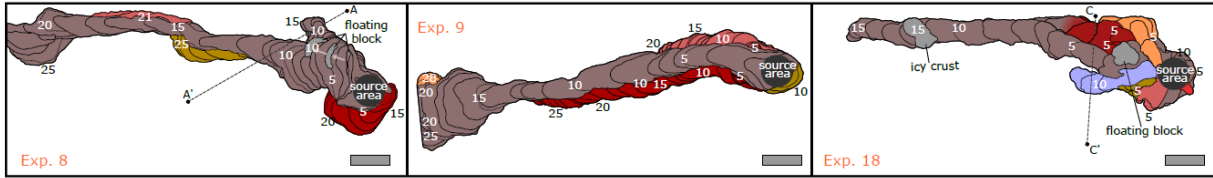


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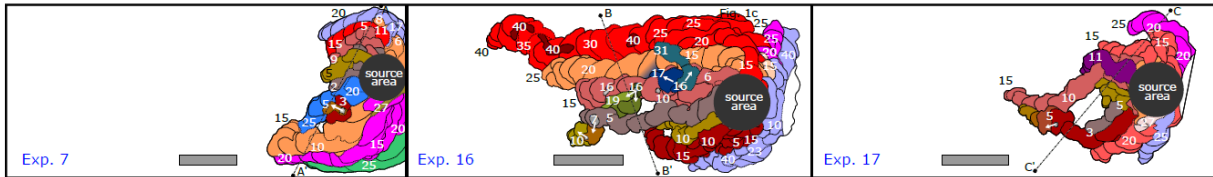
a) ambient pressure, variable release rates



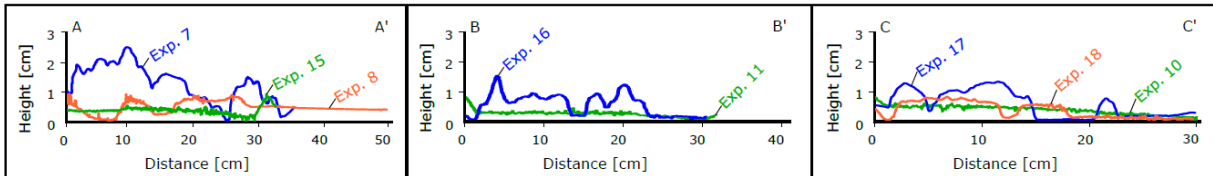
b) pressure ~7 mbar, 'high' release rates



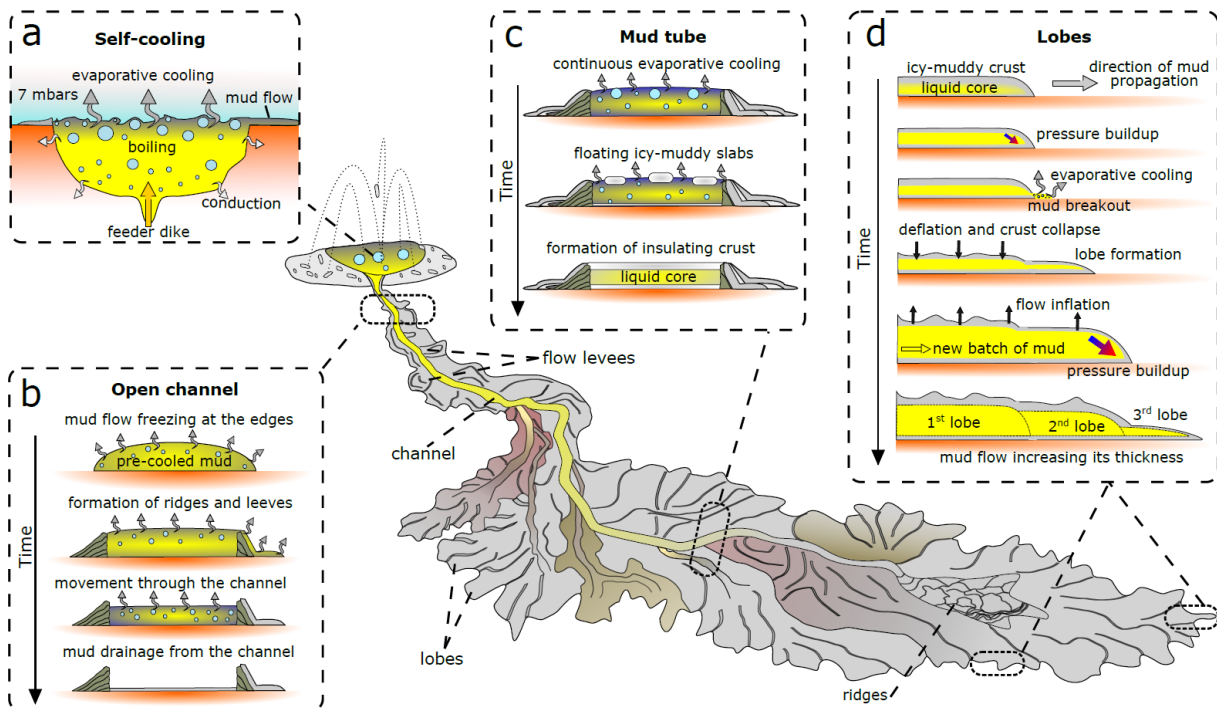
c) pressure ~7 mbar, 'low' release rates



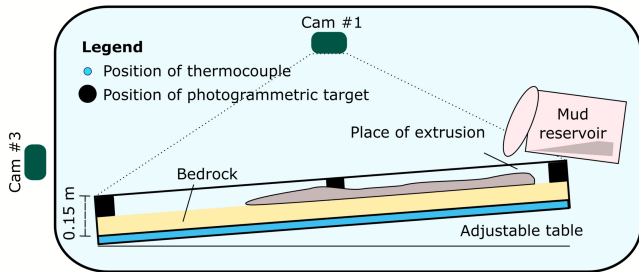
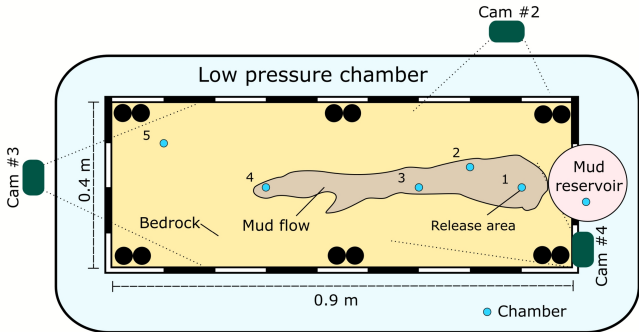
d) topography



574



575



Exp #	Pressure range [mbar]	Inclination [°]	Release time* [s]	Surface T** [°K]	Mud T*** [°K]	Type of surface	Salinity****
5	6.55 - 7.33	5	15	265	294	sand	D.I. water
6	7.16 - 7.66	5	26	264	292	sand	tap water
7	6.43 - 7.02	5	27	258	296	sand	saline water
8	7.11 - 7.37	5	25	261	294	sand	saline water
9	7.09 - 7.63	5	27	262	294	sand	saline water
10	1000	5	21	258	290	sand	saline water
11	1000	5	19	264	274	sand	saline water
15	1000	5	15	258	293	sand	saline water
16	6.32 - 6.54	5	46	256	290	sand	saline water
17	6.57 - 6.94	5	34	252	278	sand	saline water
18	6.37 - 6.58	5	15	260	290	sand	saline water
19	6.06 - 6.81	10	35	258	286	sand	saline water
21	6.55 - 6.77	10	34	253	293	sand	saline water
22	7.08 - 7.36	10	37	254	294	sand	saline water
23	6.43 - 6.97	5	40	265	293	plastic plate	saline water
24	6.66 - 7.11	5	28	254	295	plastic plate	saline water
29	6.55 - 7.08	5	40	259	295	plastic plate	saline water
34	5.29 - 6.55	5	failed exp.	259	297	plastic plate	saline water
41	1000	5	42	244	295	plastic plate	saline water
49	1000	5	34	247	290	plastic plate	saline water
54	1000	5	34	249	277	plastic plate	saline water

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

** Temperature of the surface before the release of the mud from the flipping container

*** Temperature of the mud within the container before the pressure drop

**** Saline water refers to a mixture of water that contains 0.5% w/w dissolved magnesium sulphate (MgSO₄)