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1 Transition of eruptive style: Pumice raft to dome-forming
2 eruption at the Havre submarine volcano, South Pacific
3 Ocean

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11 **ABSTRACT**

12 Transitions in eruptive style are common at volcanoes. Understanding how and
13 why these transitions occur remain open questions. The 2012 eruption of the submarine
14 Havre volcano in the Kermadec arc (South Pacific Ocean) produced a raft of floating
15 pumice followed by a pair of domes from the same vent. Here, we used measurements on
16 erupted magmas and constraints on the eruption rate, combined with a model for magma
17 ascent, to identify the dominant controls on the transition in eruption style. During the
18 raft-forming stage, magma ascent was fast enough that little gas was lost. Magma reached
19 the seafloor with great enough vesicularity to be buoyant and produce clasts that could
20 float. As the eruption waned, the eruption rate decreased and the conduit narrowed.
21 Sufficient gas was then lost to the surrounding country rocks during ascent such that the
22 erupted magma was no longer buoyant relative to seawater. Most of the original

23 dissolved water in the magma was lost to the crust surrounding the conduit during the
24 dome-forming stage.

25 **INTRODUCTION**

26 Volcanic eruptions commonly transition between different styles, for example
27 between explosive and effusive eruption. Understanding how and why these transitions
28 occur remain key outstanding questions (National Academies of Sciences, Engineering,
29 and Medicine, 2017) that provide insight into both ascent processes and hazards
30 (Cassidy et al., 2018). The 2012 silicic submarine eruption of Havre volcano in the
31 Kermadec arc, South Pacific Ocean (Fig. 1A) provides a new opportunity to understand
32 transitions in eruptive style. Initially, it created a gigantic raft of floating pumice (Fig.
33 1B; Jutzeler et al., 2014) and then it extruded a dome on the seafloor from the same vent,
34 900 m below sea level (Fig. 1C; Carey et al., 2018).

35 Here we used a model for magma ascent in a conduit, constrained by measured
36 magma properties, seafloor observations, and eruption constraints, to elucidate the
37 processes governing eruption style. We propose that as the eruption rate decreased during
38 the course of the eruption, sufficient gas loss during ascent eventually led to magma
39 erupting on the seafloor with vesicularities low enough to be denser than seawater and
40 hence to form a dome.

41 **THE 2012 HAVRE ERUPTION**

42 The 2012 Havre eruption was the largest deep silicic submarine eruption recorded
43 since A.D. 1650 (Jutzeler et al., 2014). On July 18, 2012, more than 1.2 km³ of pumice
44 (bulk volume) reached the ocean surface (Carey et al., 2018), creating a raft of pumice

45 that floated for years and distributed pumice around the Pacific and Southern Ocean
46 basins (Jutzeler et al., 2014).

47 In March 2015, to better understand this eruption, the Mapping Exploration and
48 Sampling of Havre (MESH) expedition made a high-resolution (1-m resolution)
49 bathymetric map (Fig. 1C) and collected 290 samples from different locations on the
50 submarine edifice. Submarine exploration of the volcano revealed three clastic
51 pumiceous units, and 15 domes and lavas (Carey et al., 2018). Mapping of stratigraphic
52 relationships and sampling demonstrated that the vent responsible for the pumice raft is
53 overlain by a 250-m-high, 0.11 km³ pair of domes also erupted in 2012, which we refer to
54 as the OP dome (Fig. 1C). The OP Dome is unusual in that it is offset from the structural
55 lineament parallel to the southern caldera margin that focused magma in seven other
56 locations to form smaller domes (Fig. 1).

57 The creation of pumice clasts in subaerial settings is generally attributed to the
58 fragmentation processes that lead to an explosive eruption. Manga et al. (2018) showed
59 that the high hydrostatic pressure at the vent allowed sufficient water to remain dissolved
60 in the melt such that the magma viscosity was too low to permit brittle fragmentation in
61 the conduit, and the resulting pumice raft-forming eruption was effusive. Furthermore,
62 Manga et al. (2018) proposed that buoyant magma was extruded into the ocean where it
63 fragmented upon quenching (van Otterloo et al., 2015) and was then able to float to the
64 ocean surface to supply the pumice raft (Fauria and Manga, 2018).

65 There remains a key open question: why did the extruded magma change from
66 being less to more dense than ocean water? The compositions of dome OP and raft
67 pumice are essentially identical (Table DR1 in the Supplementary Material). The main
68 obvious differences are the vesicularity and texture (Fig. 2): Raft pumice has a mean

69 vesicularity of 78% (Rotella et al., 2015; Carey et al., 2018) and the average of 36
70 samples from the dome carapace and talus is 38.9% (Table DR2). While the vesicles are
71 filled with gas, pumice and dome clast densities are less and greater than that of seawater,
72 respectively. The irregular-shaped vesicles in the dome samples (Fig. 2B) suggest gas
73 loss and collapse.

74 **ASCENT MODEL**

75 We consider a one-dimensional isothermal and quasi-steady model for magma
76 ascent through a cylindrical conduit of constant radius r following Kozono and
77 Koyaguchi (2012). Two-dimensional models (e.g., Chevalier et al., 2017) permit lateral
78 variations in properties but show qualitatively similar results. Because the phenocryst
79 volume fraction is low, $\sim 5\%$ (Carey et al., 2018), we consider two phases, melt and
80 exsolved water with volume fraction ϕ , and use subscripts l and g to denote these two
81 phases. We ignore crystallization during ascent, which would act to increase magma
82 viscosity. The mass concentration of dissolved volatiles is c . We allow the melt velocity
83 u_l and gas velocity u_g to differ, and we permit lateral gas loss through the conduit walls
84 with flux Q_w . Conservation of mass for the melt and gas are, respectively,

$$85 \quad \frac{d}{dz} [\rho_l(1-c)(1-\phi)u_l] = 0 \quad (1)$$

$$86 \quad \frac{d}{dz} [\rho_l c(1-\phi)u_l + \rho_g \phi u_g] = -Q_w \quad (2)$$

87 where z is depth. Conservation of momentum, with inertial terms neglected owing to the
88 low Reynolds number, is

$$89 \quad 0 = (1-\phi) \frac{dP}{dz} + \rho_l(1-\phi)g + F_{lw} \quad (3)$$

$$90 \quad 0 = \phi \frac{dP}{dz} + \rho_g \phi g + F_{lg} \quad (4)$$

91 where g is gravity and F_{lw} and F_{lg} describe the drag forces between magma and the
 92 conduit walls and between gas and liquid, respectively. The pressure P is assumed to be
 93 the same in the gas and melt.

94 We assume equilibrium outgassing with solubility given by

$$95 \quad c = s\sqrt{P}. \quad (5)$$

96 We assume Poiseuille flow of the magma and thus

$$97 \quad F_{lw} = \frac{8\mu}{r^2} u_l, \quad (6)$$

98 and vertical gas loss described by Darcy's law

$$99 \quad F_{lg} = \frac{\mu_g}{k} \phi^2 (u_g - u_l), \quad (7)$$

100 with permeability $k = 10^{-11} \phi^3 \text{ m}^2$ (Mueller et al., 2005). Lateral gas loss through the
 101 conduit walls is driven by the pressure difference between magma in the conduit P and
 102 lithostatic pressure P_l in the surrounding crust (e.g., Jaupart and Allègre, 1991):

$$103 \quad Q_w = \frac{2\rho_g \phi k_w}{\mu_g r^2} [P - P_l], \quad (8)$$

104 and it is 0 otherwise, with k_w being the country rock permeability. These models for
 105 vertical and lateral volatile loss ignore thermal, multiphase (e.g., liquid vs vapor), and
 106 turbulent effects. The viscosity of the magma μ_m varies with dissolved water content,
 107 which affects the melt viscosity μ_l (Manga et al., 2018), and ϕ (Llewellyn and Manga,
 108 2005), such that

$$109 \quad \mu_m = \left(1 - \phi^{\frac{5}{3}}\right) \mu_l. \quad (9)$$

110 As boundary conditions, we specify the pressure at the vent (equal to the
 111 hydrostatic value at the seafloor depth of 0.9 km) and the mass inflow rate q at the bottom
 112 of the conduit. We use $c_0 = 4.9 \text{ wt \%}$ based on melt inclusions from seafloor and raft

113 pumice (mean of 38 inclusions, standard deviation of 0.4 wt.%; summarized in Table
114 DR3), temperature $T = 850$ °C (Manga et al., 2018), and $\rho_l = 2400$ kg/m³, which is also
115 assumed equal to the crust density, and $\mu_m \cdot \mu_l$ is a function of c , calculated using data
116 from Giordano et al. (2008) for the composition and temperature reported in Manga et al.
117 (2018). We solve for four depth-dependent variables, $P(z)$, $\phi(z)$, $u_l(z)$, and $u_g(z)$ in
118 addition to the “chamber” pressure P_{ch} at the bottom of the conduit. Those variables also
119 determine magma properties such as $\rho_g = P/RT$. We assumed a 5-km-long conduit and
120 solve the coupled differential equations on a regular grid with 5 m spacing (parameters
121 are summarized in Tables DR4).

122 **RESULTS**

123 Figure 3 shows the relationship between the mass eruption rate and vesicularity at
124 the vent. We chose these two variables because they are measured (vesicularity) or
125 bounded by observations (eruption rate) for the raft- and dome-forming stages (Carey et
126 al., 2018). We considered two different conduit radii $r = 30$ m and 12 m, and three
127 different wall-rock permeabilities, $k_w = 10^{-14}$, 10^{-13} and 10^{-12} m², to cover the range
128 typical of upper crustal rocks (Manning and Ingebritsen, 1999) and oceanic crust (Fisher,
129 1998).

130 As the mass eruption rate increases, less gas is lost to the country rock, illustrating
131 “the essential result that the fraction of gas lost is inversely proportional to the eruption
132 rate because the flow of gas occurs at a given rate through the immobile country rock
133 whilst magma rises” (Jaupart and Allègre, 1991, p. 416). At the lowest mass eruption
134 rates shown, vertical gas loss can also reduce vesicularity even when the crust has a low
135 permeability. However, to achieve vesicularities similar to those of the dome without

136 lateral gas loss, eruption rates are required that are a couple orders of magnitude lower
137 than those calculated at Havre or recorded elsewhere, demonstrating that gas loss to the
138 country rock must have occurred during ascent, and that lateral gas loss (controlled by
139 country rock permeability) likely dominated over vertical gas loss (controlled by magma
140 permeability).

141 As the conduit radius decreases, the amount of gas lost from the conduit
142 increases. This occurs for two reasons. First, gas flux is inversely proportional to the
143 square of conduit radius (Eq. 8). Second, as conduit size decreases, for the same mass
144 flux, the resistance to ascent (Eq. 6) increases, leading to greater chamber and conduit
145 pressures (colors in Fig. 3) and hence larger pressure differences driving lateral gas loss
146 (Fig. DR5). Vesicularity can increase rapidly as magma approaches the vent owing to
147 both a reduction in the pressure difference between the magma and its surroundings and
148 the increasing ascent speed, which limits the time available for gas loss.

149 **DISCUSSION AND CONCLUSIONS**

150 During the course of an eruption, we expect the overpressure in the magma source
151 to progressively decrease as magma is evacuated (e.g., Woods and Koyaguchi, 1994),
152 leading to a decreasing mass eruption rate. Conduit size can also evolve: Conduit erosion
153 acts to widen ascent paths, but decreasing pressure allows conduits to narrow (e.g., Costa
154 et al., 2007) and cooling and/or crystallization of ascending magma near conduit walls
155 may further decrease the effective conduit size. We ascribe the transition in eruption style
156 at Havre volcano to both evolving magma pressure and decreasing conduit radius. The
157 conduit size was largest during the pumice raft-forming stage of the 2012 Havre eruption,
158 and minimal gas loss occurred during magma ascent because the ascent speed was too

159 high. As the eruption waned, the conduit narrowed, and vertical and lateral gas loss was
160 enhanced.

161 There are a number of idealizations in the models and uncertainties in the eruption
162 rate data used as inputs. Approximations in our model include a constant permeability for
163 the crust, a cylindrical conduit, and neglect of crystallization. The conduit during the
164 earliest phase of the eruption may well have been more elongate or dike-like, with a
165 shape that evolved over the course of the eruption (e.g., Aravena et al., 2018), but there
166 are no observations to better constrain vent and conduit geometry. The pumice raft
167 samples have a very low abundance of microlites whereas dome samples have abundant
168 microlites (Fig. 2) that nucleated and grew at some point during ascent or upon
169 emplacement. The eruption rates plotted in Figure 3 are estimates from Carey et al.
170 (2018) based on the mass erupted, constraints on the duration of eruption for the raft, and
171 a lower bound for Dome OP based on 90 d between the raft-forming stage, 18 July 2012,
172 and a comparison of bathymetric surveys on 17 October 2012 and March 2015 that
173 revealed no further growth. This lower bound is within an order of magnitude of the mass
174 eruption rates of recent small-volume rhyolite eruptions at Chaitén and Cordón Caulle
175 (Pallister et al., 2013; Schipper et al., 2013; Tuffen et al., 2013) but considerably less than
176 inferred rates for large-volume rhyolite flows (Befus et al., 2015). The eruption rate
177 might also have decreased monotonically between these estimates for the raft and dome.
178 Nevertheless, the general conclusion that increased gas loss occurs as eruption rate
179 decreases should be robust. A decrease in radius of a factor of ~ 2 combined with a
180 reasonable wall-rock permeability of $\sim 10^{-13} \text{ m}^2$ capture the observed vesicularities and
181 estimates of mass eruption rates. Alternatively, an increasing permeability from 10^{-13} to

182 10^{-12} m^2 , via fractures in the country rock or volcanoclastic layers, would also explain the
183 changes in vesicularity. We also note that modest vesiculation may have continued as
184 clasts rose in the water column above the vent, increasing raft vesicularity relative to the
185 values at the vent (Mitchell et al., 2018), which are plotted in Figure 3.

186 Our explanation for the transition in eruption style requires large volatile fluxes
187 through the magma, particularly during the dome-forming stage. Lateral volatile loss
188 from the conduit to the surrounding rocks is a substantial fraction of the total magmatic
189 volatile budget, ~25% and 70% of the initial water during the raft- and dome-forming
190 stages, respectively (Fig. 3A). Further evidence for high exsolved volatile flux includes
191 the presence of cristobalite in the dome samples (Fig. 2), which likely resulted from
192 vapor-phase crystallization (e.g., Schipper et al., 2017). Given the initial water content of
193 4.9 wt% and erupted mass of the pumice raft and Dome OP, these values correspond to
194 $5.4 \times 10^9 \text{ kg}$ and $7.8 \times 10^9 \text{ kg}$, respectively, of high-temperature supercritical water and
195 vapor supplied to the crust surrounding the conduit. These fluids in hydrothermal systems
196 have the potential to form veins and disseminated mineral deposits within highly altered
197 zones of wall rock surrounding conduits. Syneruptive inputs of magmatic volatiles, fluids
198 and metals into the shallow (~500 m) seafloor around conduits in deep submarine
199 settings cannot be assigned into classical epithermal or porphyry-style mineralization
200 models (e.g., Large, 1992; Sillitoe and Hedenquist, 2003). Hybrid-styles of epithermal-
201 volcanic-hosted massive sulfide-porphyry deposition have been proposed for both active
202 modern and ancient ore bodies, e.g., Mount Lyell (Yosemite, California; Huston and
203 Kamprad, 2001) and Brothers volcano (South Pacific Ocean; Keith et al., 2018). Greater

204 understanding of these hybrid mineral systems could be attained by geothermal, chemical
205 and hydrological modeling constrained by quantitative information from Havre volcano.

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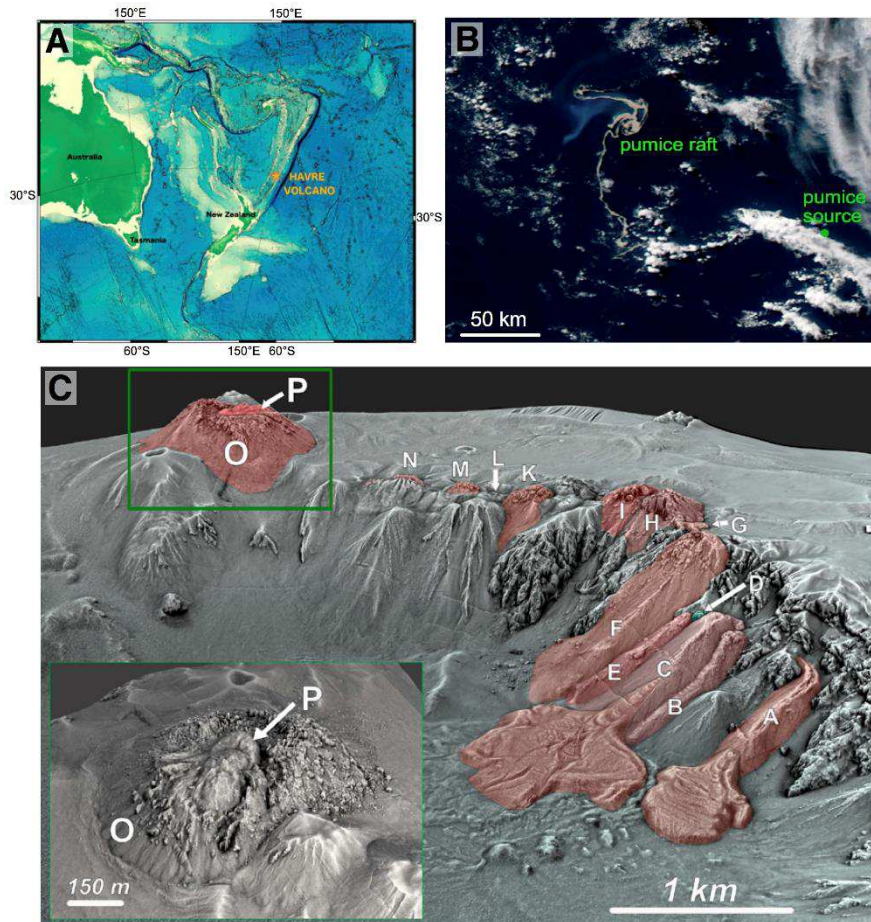
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321 **FIGURES**



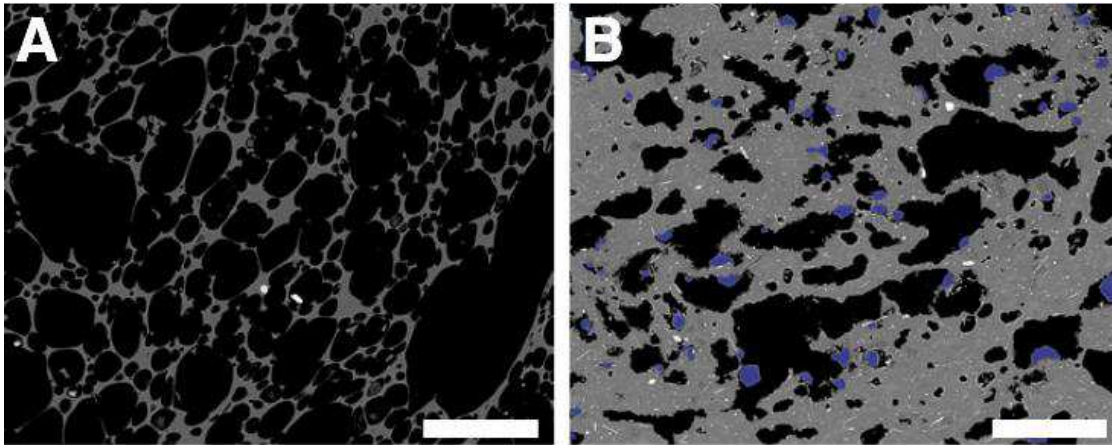
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323 Figure 1. A: Location of the Havre volcano in the Kermadec arc, South Pacific Ocean. B:
324 Image of pumice raft, 5 d after eruption. C: Map of caldera identifying newly erupted
325 lava flows and domes in red (lettered).

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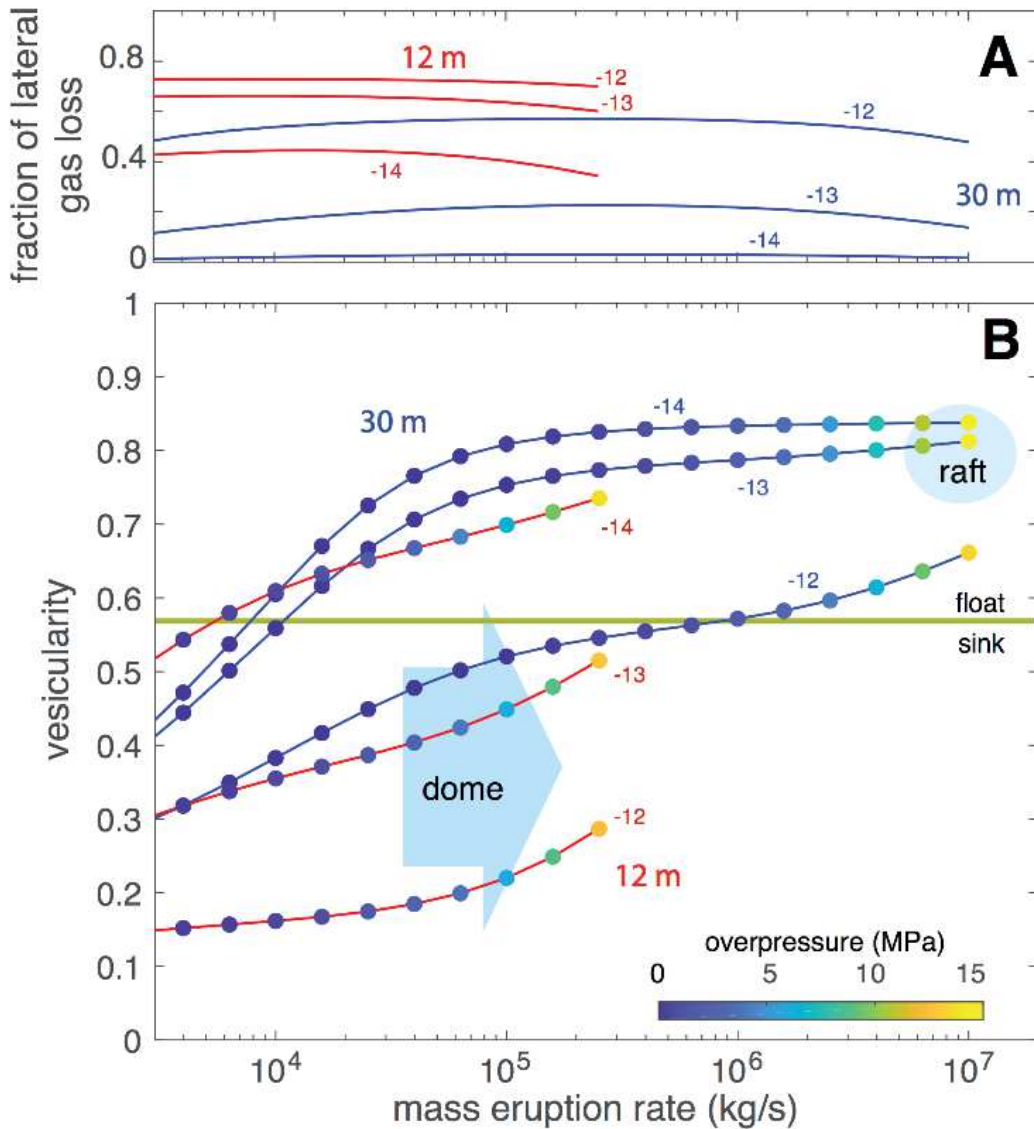
330 Figure 2. Backscatter electron images of representative clast texture from (A) the raft and

331 (B) the dome. In B cristobalite is colored blue. Microlites of plagioclase (white) and

332 pyroxene (dark gray) dominate the groundmass in B. Vesicularities are 83 vol% in A and

333 34.5 vol% in B. Bar in the lower right is 100 μm long.

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336

337 Figure 3. A: Fraction of initial total dissolved water lost to country rocks during ascent.

338 B: Relationship between mass eruption rate and vesicularity at vent. Blue and red curves

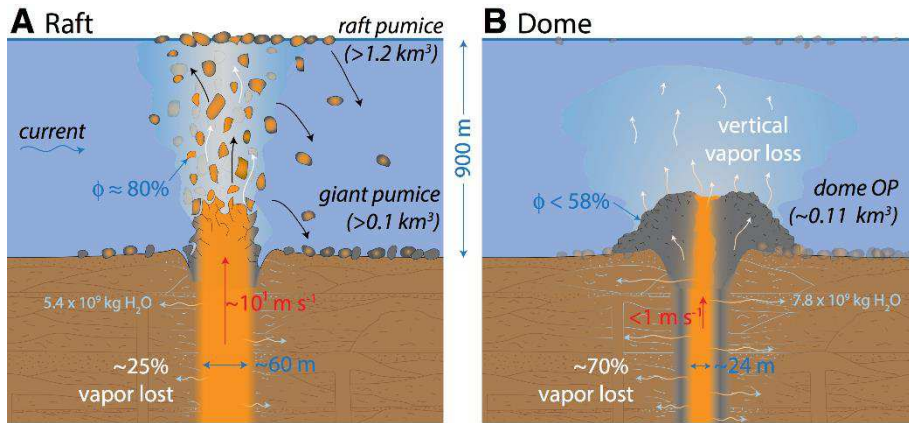
339 indicate conduit radii of 30 and 12 m, respectively. Colors of symbols show overpressure

340 at base of conduit (5 km below seafloor). Horizontal line shows vesicularity needed for

341 clasts to be buoyant prior to ingesting liquid water. In A and B, numbers next to each

342 curve are \log_{10} of permeability (in m^2).

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345 Figure 4. Cartoon illustrating the dynamics that accompanied (A) pumice raft-forming
 346 eruptions and (B) dome-forming eruptions at Havre volcano (southwest Pacific Ocean) in
 347 2012. Subsurface structure is schematic; ϕ is gas volume fraction.

348

349 Supplementary material contains Tables DR1-DR4 and Figure DR5. See online

350 publication.