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► **To cite this version:**

Pauline Verdurme, Charles Le Losq, Oryaëlle Chevrel, Salomé Pannefieu, Etienne Médard, et al.. Viscosity of crystal-free silicate melts from the active submarine volcanic chain of Mayotte. *Chemical Geology*, In press, 620, pp.121326. 10.1016/j.chemgeo.2023.121326 . hal-03965373

**HAL Id: hal-03965373**

**<https://uca.hal.science/hal-03965373v1>**

Submitted on 31 Jan 2023

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# 1 **Viscosity of silicate melts from the active submarine volcanic chain of** 2 **Mayotte**

3  
4 Pauline Verdurme<sup>a,\*</sup>, Charles Le Losq<sup>b</sup>, Oryaëlle Chevrel<sup>a</sup>, Salomé Pannefieu<sup>b</sup>, Etienne  
5 Médard<sup>a</sup>, Carole Berthod<sup>a,b,c</sup>, Jean-Christophe Komorowski<sup>b</sup>, Patrick Bachèlery<sup>a</sup>, Daniel R.  
6 Neuville<sup>b</sup> and Lucia Gurioli<sup>a</sup>

7  
8 <sup>a</sup> Université Clermont Auvergne, CNRS, IRD, OPGC, Laboratoire Magmas et Volcans, F-  
9 63000 Clermont-Ferrand, France

10 <sup>b</sup> Université Paris Cité, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France

11 <sup>c</sup> Observatoire volcanologique et sismologique de la Guadeloupe, Institut de physique du globe  
12 de Paris, 97113, Gourbeyre, France

13 \* Corresponding author:

14 E-mail address: pauline.verdurme@uca.fr

15 ORCID ID #: 0000-0003-1928-0009  
16

## 17 **Abstract**

18 Following an unprecedented seismic activity that started in May 2018, a new volcanic edifice,  
19 now called Mont Fani Maoré, was constructed on the ocean floor 50 km east of the island of  
20 Mayotte (Indian Ocean). This volcano is the latest addition to a volcanic chain characterized  
21 by an alkaline basanite-to-phonolite magmatic differentiation trend. Here, we performed  
22 viscosity measurements on five silicate melts representative of the East-Mayotte Volcanic  
23 Chain compositional trend: two basanites from Mont Fani Maoré, one tephri-phonolite and two  
24 phonolites from different parts of the volcanic chain. A concentric cylinder viscometer was  
25 employed at super-liquidus conditions between 1500 K and 1855 K and a creep apparatus was  
26 used for measuring the viscosity of the undercooled melts close to the glass transition  
27 temperature. At super-liquidus temperatures, basanites have the lowest viscosity (0.11 - 0.99  
28  $\log_{10}$  Pa·s), phonolites the highest (0.91 - 3.89  $\log_{10}$  Pa·s), while the viscosity of the tephri-

29 phonolite falls in between ( $0.89 - 1.97 \log_{10} \text{ Pa}\cdot\text{s}$ ). Near the glass transition, viscosity  
30 measurements have only been performed for one phonolitic melt because obtaining pure glass  
31 samples for the basanitic and tephri-phonolitic compositions was unsuccessful due to the  
32 formation of nanolites as evidenced by Raman spectroscopy. The phonolite has a viscosity of  
33  $10.19$  to  $12.30 \log_{10} \text{ Pa}\cdot\text{s}$  at  $1058$  to  $986 \text{ K}$ . Comparison with existing empirical models revealed  
34 an underestimation of  $1.2$  to  $2.0 \log$  units at super-liquidus and undercooled temperatures,  
35 respectively, for the phonolite. This emphasizes (i) the lack of data falling along the alkaline  
36 basanite-to-phonolite magmatic differentiation trend to calibrate empirical models, and (ii) the  
37 complexity of modeling the variations in viscosity as a function of temperature and chemical  
38 composition for alkaline composition. The presented new measurements indicate that, at  
39 eruptive temperatures between  $1050 \text{ }^\circ\text{C}$  and  $1150 \text{ }^\circ\text{C}$  ( $1323$ - $1423 \text{ K}$ ), the oxidized, anhydrous,  
40 crystal-free and bubble-free basanite melt is very fluid with a viscosity around  $1 \log_{10} \text{ Pa}\cdot\text{s}$ . In  
41 contrast, the anhydrous phonolitic crystal- and bubble-free melt at eruptive temperatures  
42 ranging from  $800$  to  $1000 \text{ }^\circ\text{C}$  ( $1073$ - $1273 \text{ K}$ ) would have a viscosity around  $6 - 10 \log_{10} \text{ Pa}\cdot\text{s}$ .  
43 Considering that both basanitic and phonolitic melts from the Mayotte submarine volcanic  
44 chain contain less than  $6 \%$  crystals and a significant amount of water, such viscosity values  
45 are probably upper limits. These new viscosity measurements are essential to define eruptive  
46 models and to better understand the storage and transport dynamics of Comoros Archipelago  
47 magmas, and of alkaline magmas in general, from the source to the surface.

48

49 **keywords**

50 viscosity, alkali magmas, submarine volcano, volcanic eruption, Raman spectroscopy

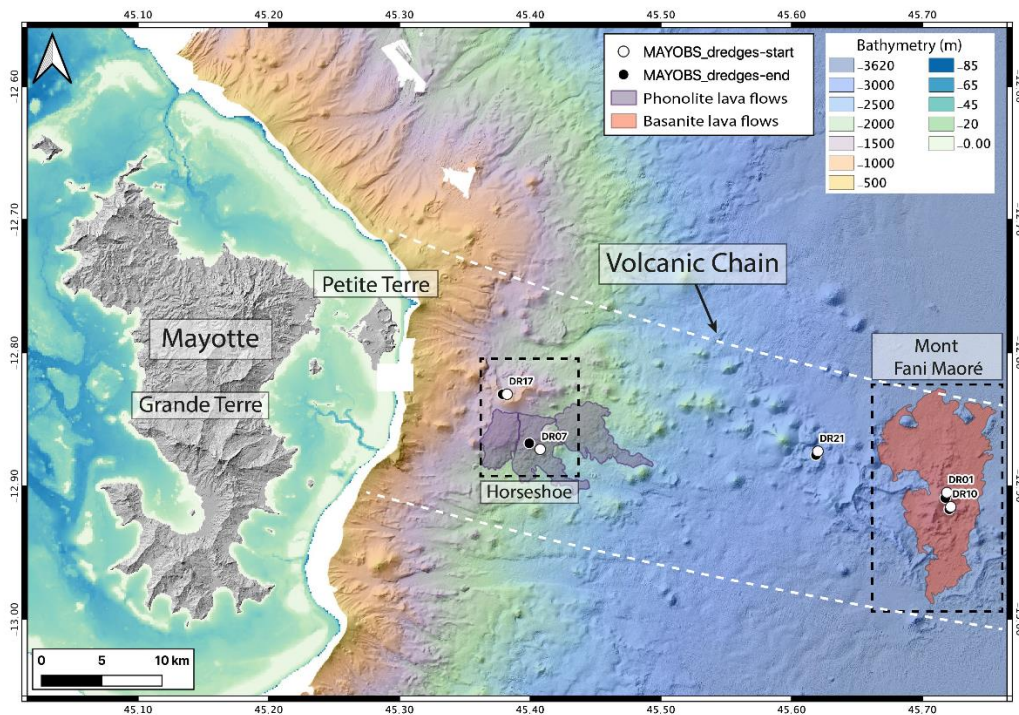
## 51 **1. Introduction**

52 Following an unprecedented intense seismic crisis that started in May 2018 ([Lemoine et al.](#)  
53 [2020](#); [Feuillet et al. 2021](#)), surveys revealed a new submarine volcanic edifice, recently named  
54 Mont Fani Maoré, 936 m tall with a base at around 3,500 m depth, 50 km east of Mayotte  
55 (Indian Ocean). This large eruption has extruded around 6.55 km<sup>3</sup> of basanite magma, with  
56 first estimates of magma transfer rates from source to surface of a minimum of 30 days ([Cesca](#)  
57 [et al. 2020](#); [Berthod et al. 2021a](#)). The source of this magma is evaluated to be located at 30 -  
58 50 km depth into the underlying mantle, with the potential involvement of an intermediate  
59 magma chamber located at ~17 km depth ([Berthod et al. 2021b](#)). This new volcano is an  
60 addition to the East-Mayotte Volcanic Chain, characterized by the emission of magmas falling  
61 along an alkaline basanite-to-phonolite magmatic differentiation trend. This volcanic chain is  
62 quite complex and characterized by large effusive lava flow fields, and by the presence of more  
63 explosive volcanoes (**Figure 1**) ([Puzenat et al. 2022](#); [Gurioli et al. 2023](#); [Komorowski et al.](#)  
64 [2023](#)). To better understand and constrain the storage processes and transfer rates of the  
65 magmas that feed the East-Mayotte Volcanic Chain and the dynamics of their emplacement at  
66 surface, it is necessary to constrain the viscosity of the basanite-to-phonolite melts, which  
67 determines the mobility of magmas toward the surface as well as their fragmentation behavior  
68 in the conduit and flow at the surface ([Dingwell et al. 1996](#); [Papale 1999](#); [Pistolesi et al. 2011](#)).  
69 To constrain the viscosity of magmas and lavas, one can potentially rely on the various  
70 experimental and modeling efforts that led to important improvements in our knowledge of  
71 how it is controlled by temperature ([Vogel 1921](#); [Tamman and Hesse 1926](#); [Fulcher 1925](#);  
72 [Adam and Gibbs 1965](#); [Nascimento and Aparicio 2007](#)), chemical composition ([Bottinga and](#)  
73 [Weill 1972](#); [Shaw 1972](#); [Giordano et al. 2008](#)), volatile elements (e.g., [Whittington et al. 2000](#),  
74 [2001](#)), bubbles ([Lejeune et al. 1999](#); [Llewelin and Manga et al. 2005](#)) and crystal contents  
75 ([Lejeune and Richet 1995](#); [Kolzenburg et al. 2022](#)). Yet, there is not a general and very accurate

76 magma viscosity model to date, due to the complexity in estimating the influence of the  
77 different parameters listed above. Focusing only on the silicate melt phase of magmas and  
78 lavas, many effects exist at the atomic level, such as aluminum coordination changes or metal  
79 cation mixing, that result in strong and non-linear variations in melt viscosity depending on  
80 composition. In particular, in alkali-rich melts, the way Na and K mix may strongly affect  
81 viscosity (Poole, 1948; Richet, 1984; Le Losq and Neuville, 2013, 2017). In Al-poor melts, the  
82 ideal mixing between Na and K results in an important decrease in viscosity (Richet 1984),  
83 whereas in Al-rich melts, Na and K occupy different environments and do not mix ideally (Le  
84 Losq and Neuville 2013, 2017; Le Losq et al. 2021). At a given temperature, this leads to  
85 increasing the melt viscosity by several orders of magnitude upon increasing the melt  
86 K/(K+Na) ratio. Existing parametric viscosity models (e.g., Hui and Zhang 2007; Giordano et  
87 al. 2008) failed to properly reproduce such an effect (Le Losq and Neuville 2013; Robert et al.  
88 2019). Therefore, their accuracy for the prediction of the viscosity of melts along alkaline  
89 magmatic series is to be improved. Only models integrating thermodynamic and physico-  
90 chemical knowledge allow to properly model the way Na and K non-ideal mixing alter the  
91 viscosity of Al-rich alkali melts. However, this has been applied only on quaternary alkali  
92 aluminosilicate melts far from natural compositions (Robert et al. 2019; Starodub et al. 2019;  
93 Le Losq et al. 2021, Neuville and Le Losq 2022). Therefore, to properly constrain the viscosity  
94 of alkali melts, one needs to use *ad hoc* models for increased precision, as it has been performed  
95 for the phonolite of Erebus volcano (Antarctica) for instance (Le Losq et al. 2015a).

96 In this contribution, we present new viscosity measurements performed on samples  
97 representative of the lava compositions that can be found along the East-Mayotte submarine  
98 volcanic chain. Those samples fall along the moderately silica-undersaturated trend (“Karthala-  
99 trend”) of the Comoros Archipelago (Pelleter et al. 2014; Bachèlery et al. 2016; Bachèlery and  
100 Hémond 2016). The studied compositions include two basanite lavas from the recent eruption

101 (Mont Fani Maoré volcano), and two phonolite and a tephri-phonolite samples from other  
102 emission sites in the submarine volcanic chain (Berthod et al. 2021a; Feuillet et al. 2021;  
103 Puzenat et al. 2022; Figure 1). Viscosity is measured with a concentric cylinder viscometer at  
104 super-liquidus temperatures, in the  $10^0 - 10^6$  Pa·s range, and with a creep apparatus close to the  
105 glass transition temperature between  $10^8$  and  $10^{13}$  Pa·s. We also examine the effect of the water  
106 content on the viscosity of the different magma compositions present at Mayotte. The results  
107 (i) provide the foundation necessary to investigate the eruptive behavior of the Mayotte  
108 volcanic system, (ii) strengthen the current knowledge on alkaline series and (iii) could  
109 contribute to refine the pre-existing empirical viscosity models based on chemical composition.



110  
111 **Figure 1.** Geological map of the submarine volcanic chain of Mayotte showing the location of  
112 the dredged samples (DR labels). Mont Fani Maoré lava flows are modified from Feuillet et  
113 al. (2021). Modified from Berthod et al. (2021a). Background is the bathymetry from the  
114 Homonim project (SHOM 2015), DEM Litto3D IGN-SHOM (SHOM 2016) and MAYOBSI  
115 (doi:10.17600/18001217).

116 **2. Materials and Methods**

117 **2.1. Starting material**

118 In this study, the investigated samples were collected by dredging operations that collected  
 119 between 100 - 1000 kg of rocks, from water depths ranging from 1,370 to 3,455 m (**Figure 1**).  
 120 These dredges were operated by the R/V *Marion Dufresne II* and R/V *Pourquoi Pas?* during  
 121 the following oceanographic cruises (**Rinnert et al., 2019**): MAYOBS 1 (**Feuillet, 2019**),  
 122 MAYOBS 2 (**Jorry 2019**), MAYOBS 4 (**Fouquet and Feuillet, 2019**), MAYOBS 15 (**Rinnert**  
 123 **et al., 2020**), and GEOFLAMME (**Rinnert et al., 2021**) (**Table 1**).

124 **Table 1.** Location of the dredges performed during the oceanographic cruises. Latitudes and  
 125 Longitudes are given in decimal degrees (DD).

Dredges	Oceanographic cruises	DOI of the oceanographic cruises <a href="https://doi.org/10.18142/291">10.18142/291</a>	Start dredging			End dredging		
			Latitude	Longitude	Depth	Latitude	Longitude	Depth
<b>DR01</b>	MAYOBS 1	<a href="https://doi.org/10.17600/18001217">10.17600/18001217</a>	-12.905	45.719	3050 m	-12.909	45.718	2820 m
<b>DR07</b>	MAYOBS 2	<a href="https://doi.org/10.17600/18001222">10.17600/18001222</a>	-12.872	45.407	1590 m	-12.868	45.399	1585 m
<b>DR10</b>	MAYOBS 4	<a href="https://doi.org/10.17600/18001238">10.17600/18001238</a>	-12.916	45.722	3120 m	-12.918	45.721	2950 m
<b>DR17</b>	MAYOBS 15	<a href="https://doi.org/10.17600/18001745">10.17600/18001745</a>	-12.831	45.382	1370 m	-12.831	45.379	1340 m
<b>DR21</b>	GEOFLAMME	<a href="https://doi.org/10.17600/18001297">10.17600/18001297</a>	-12.874	45.620	2719 m	-12.876	45.618	2629 m

126  
 127 We selected five samples that represent the diverse composition of the volcanic deposits along  
 128 the Mayotte submarine volcanic chain (**Fig. 1** and **Table 2**). Two samples (MAY01-DR010101  
 129 and MAY04-DR100504, hereafter named DR01 and DR10, respectively) are from Mont Fani  
 130 Maoré, the new volcano that has been in activity from 2018 to 2021 (**Fig. 1**, **Lemoine et al.**

131 [2020; Berthod et al. 2021b; Feuillet et al. 2021](#)). These two samples are fragments of basanite  
132 lava flows collected on the central edifice and were emitted during the first year of the eruption  
133 (before May 2019, [Berthod et al., 2021b](#)). Two other samples (MAY02-DR070201 and  
134 MAY15-DR170404, hereafter named DR07 and DR17, respectively) were recovered at 10 -  
135 15 km east of Petite Terre Island of Mayotte on the “Horseshoe” volcanic structure ([Fig. 1,](#)  
136 [Berthod et al. 2021a; Feuillet et al. 2021; Puzenat et al. 2022; Gurioli et al. 2023](#)). DR07 is a  
137 fragment of a Holocene phonolitic lava flow located on the southeastern part of the Horseshoe  
138 site. DR17 is a phonolitic pyroclast bomb collected at the bottom of the internal western side  
139 slope of the Horseshoe structure. Finally, the fifth sample (GFL-DR2110, hereafter named  
140 DR21) was collected on a seamount located in the middle of the submarine volcanic chain, 10  
141 km west of Mont Fani Maoré ([Rinnert et al., 2021](#)). This sample is a fragment of lava and has  
142 a tephri-phonolitic composition.

143 About 200 g of each sample were used to produce the starting glasses necessary for the  
144 experiments. The rock samples were first crushed and powdered. A portion of the powder was  
145 used for bulk rock chemical analyses and the rest for viscosity measurements. Chemical  
146 analyses have also been performed on the glass chips produced after high- and low-  
147 temperatures viscometry.

148

## 149 **2.2. Major element analyses**

150 Bulk rock major element composition of the starting materials was analyzed at Laboratoire  
151 Magmas et Volcans (LMV, Clermont-Ferrand, France). Samples were crushed into millimeter-  
152 size chips using home-made thermally hardened steel jaws, and powdered in a motorized agate  
153 mortar. Major elements were analyzed by Inductively Coupled Plasma Atomic Emission  
154 Spectrometry (ICP-AES). Powdered samples were melted with LiBO<sub>2</sub> in a magnetic induction  
155 oven at 1100 °C for 5 min using graphite crucibles. The glass beads were then dissolved in a



156 solution of deionized water and nitric acid (1 M) and diluted by a factor of 2000 to produce the  
157 solution analyzed by an Agilent 5800 VDV ICP AES in radial mode. Analytical uncertainties  
158 ( $\pm 2\sigma$ ) vary between 1 and 3 % except for  $K_2O$ ,  $MnO$  (4 %) and  $P_2O_5$  (7 %) for the DR-N  
159 standard (diorite). For the analysis, the plasma flow was 12 L/min, the nebulizer flow was 0.7  
160 L/min and the radio frequency power was 1.2 kW.

161 Glass chips resulting from spindle quench after the super-liquidus viscometry were mounted  
162 as polished sections for chemical analysis carried out with the CAMECA SX Five Tactis  
163 electron microprobe at LMV. We used an accelerating potential of 15 kV at current of 8 nA,  
164 with a defocused beam of 20  $\mu m$  diameter to avoid Na migration under the electron beam.  
165 Natural and synthetic mineral standards, including orthoclase (K, Al), albite (Na), wollastonite  
166 (Si, Ca), fayalite (Fe), forsterite (Mg),  $TiMnO_3$  (Ti, Mn), NiO (Ni),  $Cr_2O_3$  (Cr), and fluorapatite  
167 (P) were used for routine calibration.

168 We also calculated the NBO/T parameter (number of non-bridging oxygens per tetrahedral),  
169 representing the degree of polymerization of the melt following [Mysen et al. \(1982\)](#):

$$170 \quad \frac{NBO}{T} = \frac{(2O)-(4T)}{T}, \quad (1)$$

171 with O the atomic proportion of oxygen atoms and T that of cations entering as network formers  
172 in tetrahedral coordination, namely  $SiO_2$ ,  $TiO_2$ ,  $Al_2O_3$ , and  $Fe_2O_3$  (see for a review of such  
173 concepts [Le Losq et al. 2019](#); [Neuville and Le Losq 2022](#)). Here, the total iron is expressed as  
174  $Fe^{3+}$  as the viscosity experiments were performed in an oxidizing environment. This  
175 assumption is a valid simplification: estimations with the [Moretti \(2005\)](#) model yields 90 % or  
176 more of Fe residing in the melt as  $Fe_2O_3$  at the conditions of our experiments. The NBO/T of  
177 the natural melt may thus be higher, considering the occurrence of reduced iron and its  
178 influence on the melt structure (e.g., [Le Losq et al. 2021](#)).

179        **2.3. High temperature viscometry**

180 High temperature viscosity measurements are performed at super-liquidus conditions, in air,  
181 using a concentric cylinder viscometer (Dingwell 1986; Spera et al. 1988) at the Institut de  
182 physique du globe de Paris (IPGP, Paris, France). Powdered glass samples are first melted at  
183 1800 K in a Pt<sub>95</sub>Au<sub>5</sub> cylindrical crucible (50 mm height, 27 mm inner diameter, 1 mm wall  
184 thickness) in a muffle furnace. Once the crucible is full, it is inserted in the hot zone of the  
185 viscometer vertical tube furnace, heated by Super Kanthal 33 elements in air. The length of the  
186 hot zone was determined by temperature measurement inside the crucible: the vertical and  
187 radial temperature gradients inside the crucible are respectively less than 2 K for 5 cm (vertical  
188 gradient) and 1 K for 2.7 cm (radial gradient) at 1500 K and a little less at 1800 K. The inner  
189 rotating spindle size is 14 mm in diameter, 21 mm in height and has 23° conical extremities to  
190 reduce termination effects and a 5 mm diameter stem. Viscosity is measured using a Rheomat  
191 115 rheometer head, which allows rotating the cylinder at angular velocities ranging between  
192 0.05 and 780 rpm (Neuville 2006). The torque exerted on the cylinder by the sample is recorded  
193 digitally. The correspondence between the exerted torque and the viscosity is calibrated against  
194 the reference sample NBS SRM 710a, for which the viscosity–temperature relationship is  
195 accurately known (Neuville 2006). The accuracy is of the order of 0.02 log Pa·s (e.g., Neuville  
196 2006).

197

198        **2.4. Low temperature viscometry**

199 Low temperature viscosity measurements are performed near the glass transition temperature  
200 ( $T_g$ ) using a creep apparatus at IPGP (Neuville and Richet 1991; Neuville 2006). Glass samples  
201 used for measurements are (i) a parallelepiped of 7.9 mm length, and (ii) a small cylinder of  
202 2.2 mm diameter x 8.1 mm length. A silver cylinder is placed around the sample, creating a  
203 small chamber in which temperature is homogeneous. Lateral and vertical temperature

204 gradients are controlled using two Pt-PtRh<sub>10</sub> thermocouples (ITS90 type S thermocouples);  
205 before and during each measurement, lateral and vertical temperature gradients were always  
206 lower than 0.2 K. To measure sample viscosity at a given temperature, we performed 20 to 30  
207 measurements at different stresses (between 6.4 and 8.2 log Nm<sup>-2</sup>) to check for the occurrence  
208 of a non-Newtonian behavior, which could be a sign of crystallization of the sample. Each  
209 reported viscosity value at a given temperature is the statistical mean of these measurements.  
210 Measurements carried out on the NBS 717 glass show that errors on viscosity measurements  
211 are lower than 0.03 log Pa·s with this apparatus (e.g., [Neuville 2006](#)).

212

### 213 **2.5. Raman spectroscopy**

214 The Raman spectra of the glasses (pre- and post-experiments) were recorded with the Labram  
215 HR Evolution spectrometer available at IPGP, equipped with a Peltier-cooled CCD and a 1800  
216 lines mm<sup>-1</sup> grating. The samples were excited with a Coherent MX 488 nm solid-state laser  
217 focused through a ×50 Olympus objective on the sample surface. The confocal aperture of the  
218 spectrometer was set to 50. With this setup, spectral resolution is ~3 cm<sup>-1</sup> and spatial resolution  
219 is ≤1 cm<sup>-1</sup>. All spectra were recorded with the laser focused at 3-5 μm below the sample's  
220 surface to avoid any surface effects ([Behrens et al. 2006](#), [Schiavi et al. 2018](#)). The laser power  
221 on the sample has been adjusted to lower than 10 mW to avoid any potential damage on the  
222 sample (iron oxidation effects or melting). Potential damage was checked by recording several  
223 spectra on the same spot and varying laser power, without the identification of any effect.  
224 Raman data treatment was performed using the Python programming software, with the *rampy*  
225 open-source software library ([Le Losq 2018](#)).

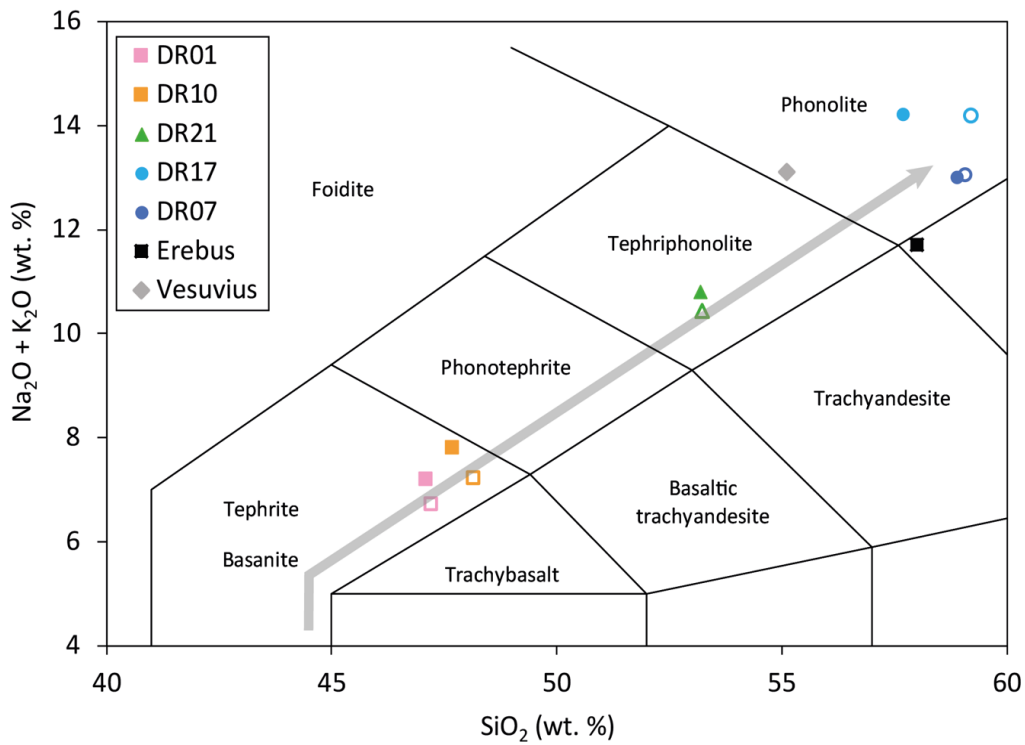
## 226 **3. Results**

### 227 ***3.1. Major element composition***

228 The bulk rock chemical composition of the pre-experiment samples and the post-experiments  
229 glass compositions, determined by ICP-AES and by electron microprobe analyses respectively,  
230 are reported in **Table 2** and presented in **Figure 2**.

231 DR01 and DR10 basanites have a very similar composition between 47.1 and 47.7 wt% SiO<sub>2</sub>  
232 and 7.2 to 7.8 alkali (Na<sub>2</sub>O + K<sub>2</sub>O) wt% (**Fig. 2**). These samples have low MgO contents,  
233 between ~4.8 and ~5.4 wt%, and high FeO<sub>tot.</sub> contents, between ~12.5 and ~12.7 wt% (**Table**  
234 **2**), making them the most iron-rich among all the studied samples from the East-Mayotte  
235 submarine volcanic chain. DR07 and DR17 phonolite samples have similar silica content of  
236 57.7 - 58.9 wt% SiO<sub>2</sub> but DR07 has an alkali content of 13.0 wt% while DR17 is more enriched  
237 in alkali with 14.2 wt% (**Fig. 2** and **Table 2**). Compared to basanite samples, phonolite samples  
238 have lower FeO<sub>tot.</sub> contents, ranging from 6.1 to 7.2 wt%. Finally, sample DR21 has a tepri-  
239 phonolitic composition, falling in between the basanite and phonolite fields of the TAS diagram  
240 (**Fig. 2**) with SiO<sub>2</sub> and alkali contents of 53.2 wt% and 10.8 wt%, respectively (**Table 2**) and  
241 an intermediate FeO<sub>tot.</sub> content of 10 wt%.

242 For most samples, the major element contents show no significant differences before and after  
243 viscosity experiments (**Table 2**). We however note that the composition of DR17 after the high-  
244 temperature experiment has a slightly higher SiO<sub>2</sub> content of 1.5 wt%. This is accompanied by  
245 iron loss of 0.5 wt% (**Table 2**).



246

247 **Figure 2.** TAS diagram showing normalized compositions of the studied samples before (filled  
 248 symbols, ICP-AES) and after (empty symbols, Electron Microprobe) viscosity measurements.  
 249 Compositions of phonolite lavas from Erebus and Vesuvius (GP79, Grey Pumice of the 79  
 250 eruption) are also shown for comparison (data from [Le Losq et al. 2015a](#)). The grey arrow  
 251 shows the moderately silica-undersaturated trend (“Karthala-trend”) identified by [Pelleter et](#)  
 252 [al. \(2014\)](#) and [Bachèlery and Hémond \(2016\)](#). Error bars are smaller than symbol size.

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261 **Table 2.** Normalized major element composition in wt% pre- (first line) and post- (second line)  
 262 experiments of the studied samples. The post-experiment composition are an average of at least  
 263 ten measurements made on glass. The composition of Erebus and Vesuvius GP79 from *Le Losq*  
 264 *et al. (2015a)* are also reported for comparison.

Sample	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO(t)	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	P <sub>2</sub> O <sub>5</sub>	NBO/T
<i>Before and after experiments</i>											
DR01	47.1	3.3	15.3	12.7	0.2	5.4	7.3	4.7	2.5	1.6	0.3
	47.2	3.4	14.8	13.2	0.3	5.4	7.3	4.4	2.4	1.6	-
DR10	47.7	3.1	15.2	12.5	0.2	4.8	6.8	5.2	2.6	1.9	0.3
	48.1	3.2	15.3	12.6	0.3	4.9	6.8	4.6	2.6	1.7	-
DR21	53.2	1.6	16.5	9.8	0.3	1.9	4.2	6.8	4.0	1.0	0.2
	53.2	1.7	16.9	10.0	0.3	2.0	4.5	6.5	4.0	1.0	-
DR07	58.9	0.4	18.2	7.2	0.3	0.4	1.6	7.6	5.4	0.3	0.1
	59.1	0.4	18.6	6.4	0.3	0.3	1.6	7.6	5.7	0.3	-
DR17	57.7	0.1	18.7	6.1	0.3	0.1	1.3	8.6	5.6	0.1	0.1
	59.2	0.1	19.1	5.6	0.2	0.1	1.3	8.4	5.8	0.1	-
Erebus melt	58.0	1.0	20.0	5.4	0.2	0.9	2.6	7.1	4.6	0.0	0.1
Vesuvius GP79	55.1	0.5	19.5	4.1	0.1	1.8	5.3	4.4	8.7	0.0	0.3

265  
 266 The calculated NBO/T values of the oxidized melts range from 0.1 to 0.3 (**Table 2**). DR01 and  
 267 DR10 basanite samples have the highest NBO/T (0.32 and 0.33). On the other hand, DR07 and  
 268 DR17 phonolite samples have the lowest NBO/T (0.10 and 0.11) related to their higher SiO<sub>2</sub>  
 269 content. The tephri-phonolite has an intermediate NBO/T of 0.20.

270

### 271 **3.2. Viscosity data**

272 The viscosity in air for anhydrous and crystal- and bubble-free basanitic, tephri-phonolitic and  
 273 phonolitic melts of the Mayotte submarine volcanic chain are presented as a function of  
 274 temperature in **Figures 3** and **4** and given in **Table 4**. High-temperatures viscosity was obtained  
 275 for all samples while low-temperature viscosity was only obtained for DR07 (**Fig. 3** and **Table**

276 4). As discussed below, this is because it was not possible to prepare the crystal-free glass  
277 samples needed for the low-temperature measurements for the other samples.

278 High-temperature viscosity measurements were performed between 1565 K and 1860 K. The  
279 lower bond of this range is motivated by the fact that, according to the MELTS model (Ghiorso  
280 and Sack 1995), magnetite crystallization is expected to occur below 1450 K for the studied  
281 samples. In this restricted high-temperature range, the relationship between viscosity and  
282 temperature is mostly linear (Fig. 3), in agreement with Bottinga et al. (1982). The data are  
283 therefore interpolated using the Arrhenius relation:

$$284 \log_{10} \eta = A_{Arr} + \frac{E_a}{RT}, \quad (2)$$

285 where  $\eta$  is the melt viscosity (Pa·s),  $A_{Arr}$  is a constant,  $T$  is temperature (K),  $R$  is the gas  
286 constant ( $\text{J mol}^{-1} \text{K}^{-1}$ ), and  $E_a$  is the viscous flow activation energy ( $\text{kJ mol}^{-1}$ ).

287 For DR07, as we obtained both high- and low-temperatures data (Fig. 3 and Table 4), we use  
288 the Vogel-Fulcher-Tamman (VFT; Vogel 1921; Tamman and Hesse 1926; Fulcher 1925)  
289 equation to interpolate the viscosity data:

$$290 \log_{10} \eta = A + \frac{B}{T-C}, \quad (3)$$

291 with  $A$ ,  $B$  and  $C$  the pre-exponential factor, the pseudo-activation energy and the VFT  
292 temperature, respectively. The viscosity-temperature variation of the phonolite melts from  
293 Vesuvius and Erebus are also shown as comparison (Fig. 3).

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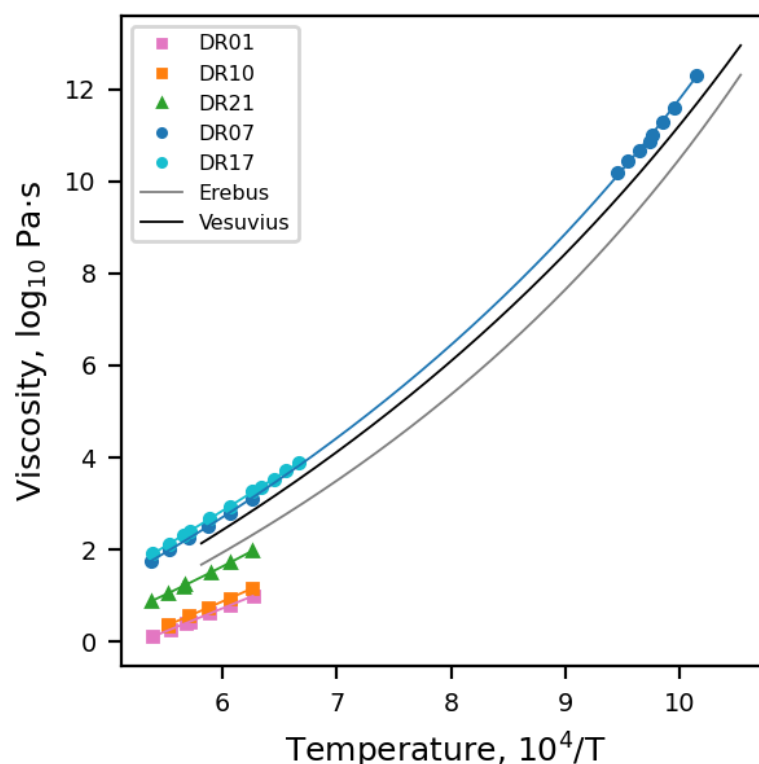
300 **Table 3.** Viscosity measurements of the basanitic-to-phonolitic melts along the Mayotte  
 301 submarine volcanic chain. The uncertainty is equal to or lower than 0.03 log (Pa·s).

Basanitic melt		Tephri-phonolitic melt		Phonolitic melt					
DR01		DR10		DR21		DR07		DR17	
<i>T</i> (K)	<i>η</i> (log <sub>10</sub> Pa·s)	<i>T</i> (K)	<i>η</i> (log <sub>10</sub> Pa·s)	<i>T</i> (K)	<i>η</i> (log <sub>10</sub> Pa·s)	<i>T</i> (K)	<i>η</i> (log <sub>10</sub> Pa·s)	<i>T</i> (K)	<i>η</i> (log <sub>10</sub> Pa·s)
						986	12.30		
						1006	11.61		
						1016	11.30		
						1025	11.00		
						1027	10.87		
						1037	10.67		
						1048	10.44		
						1058	10.19		
								1499	3.89
								1525	3.71
								1551	3.54
								1576	3.37
1595	0.99	1596	1.16	1596	1.97	1597	3.10	1596	3.27
1646	0.79	1648	0.94	1647	1.73	1648	2.80	1646	2.95
1698	0.61	1699	0.74	1696	1.50	1700	2.52	1698	2.67
1749	0.43	1753	0.56			1751	2.25	1750	2.40
1761	0.40			1762	1.23				
				1766	1.20			1766	2.32
1801	0.27	1809	0.38	1809	1.04	1804	2.01	1807	2.12
		1810	0.34						
1855	0.11			1858	0.89	1860	1.75	1855	1.91

302  
 303 At super-liquidus temperatures, the basanitic melts have the lowest viscosity (0.99 to 0.11 log<sub>10</sub>  
 304 Pa·s at 1595 to 1855 K) followed by the tephri-phonolitic (1.97 to 0.89 log<sub>10</sub> Pa·s at 1596 to  
 305 1858 K) and the phonolitic melts (3.89 to 0.91 log<sub>10</sub> Pa·s at 1499 to 1855 K). **Figure 3** shows  
 306 that, at a given temperature (between 1500 to 1855 K), the viscosity of the basanitic melts differ  
 307 from those of the phonolitic ones by about two orders of magnitude. In particular, at 1750 K,  
 308 the viscosity is of 0.43, 1.23 and 2.40 log<sub>10</sub> Pa·s, for the basanitic (DR01), tephri-phonolitic  
 309 (DR21) and phonolitic (DR17) melts, respectively (**Table 3**). At undercooled temperatures, the  
 310 viscosity of DR07 ranges from 12.3 log<sub>10</sub> Pa·s at 986 K to 10.2 log<sub>10</sub> Pa·s at 1058 K (**Fig 3**).



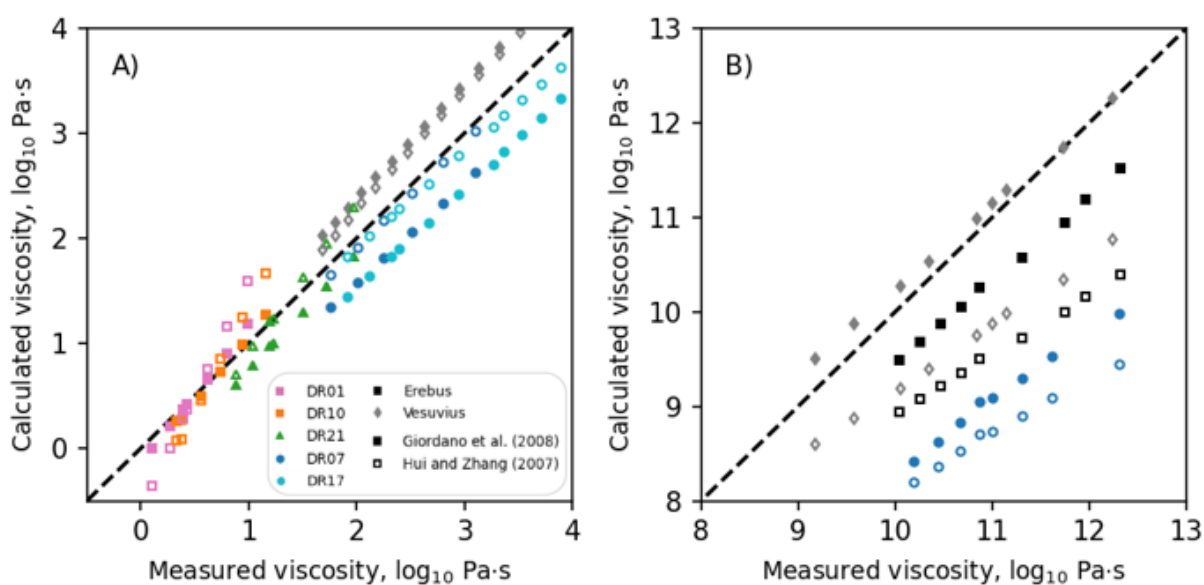
311 We compare our measurements to the [Giordano et al. \(2008\)](#) and [Hui and Zhang \(2007\)](#)  
312 empirical models (hereafter abbreviated as the GRD and HZ models, respectively). Based on  
313 [Figure 4](#), the GRD model tends to fit well the viscosity of basanite while a slight  
314 underestimation is observable for the tephri-phonolitic melts at high-temperatures. This  
315 becomes critical for the phonolite melts for which the GRD model significantly underestimates  
316 the viscosity at both high- and low-temperatures ([Fig. 4](#)). Conversely the HZ model tends to  
317 slightly overestimate the viscosity of the basanites and tephri-phonolites at high-temperatures.  
318 It is in good agreement with the measured viscosity of phonolites at high-temperatures but  
319 significantly underestimates the values at low-temperatures ([Fig. 4](#)). As also shown by [Le Losq](#)  
320 [et al. \(2015a\)](#) the GRD model underestimates by  $\sim 0.5$  log units the viscosity of Erebus ([Fig.](#)  
321 [4](#)). In contrast, for the Vesuvius melt, at super-liquidus temperatures there is a difference of  
322 less than almost  $\sim 0.2$  log units between the predicted and the measured viscosity ([Fig. 4A](#)),  
323 while at low-temperatures the GRD model fits well the viscosity data ([Fig. 4B](#)).



324

325 **Figure 3.** Viscosity ( $\log_{10} \text{Pa}\cdot\text{s}$ ) as a function of inverse temperature ( $\text{K}^{-1}$ ) for the Mayotte  
 326 basanite-to-phonolite magmatic differentiation trend. Squares, diamonds and circles represent  
 327 measurements made on basanites (DR01 and DR10), tephri-phonolite (DR21) and phonolites  
 328 (DR07 and DR17), respectively. Erebus (grey curve) and Vesuvius (black curve) are also  
 329 represented (data from [Giordano et al. 2009](#); [Le Losq et al. 2015a](#)). Solid lines are  
 330 interpolations of the data with the VFT equation ([Eq. 3, Table 4](#)). Error bars are smaller than  
 331 the symbol size.

332



333  
 334 **Figure 4.** Comparison between the measured viscosity ( $\log_{10} \text{Pa}\cdot\text{s}$ ) and the calculated ones  
 335 with the models of [Giordano et al. \(2008\)](#) (filled symbols) and of [Hui and Zhang \(2007\)](#) (open  
 336 symbols) in the super-liquidus (A) and undercooled (B) temperature ranges. Error bars are  
 337 smaller than the symbol size.

338

### 339 **3.3. Raman spectroscopy**

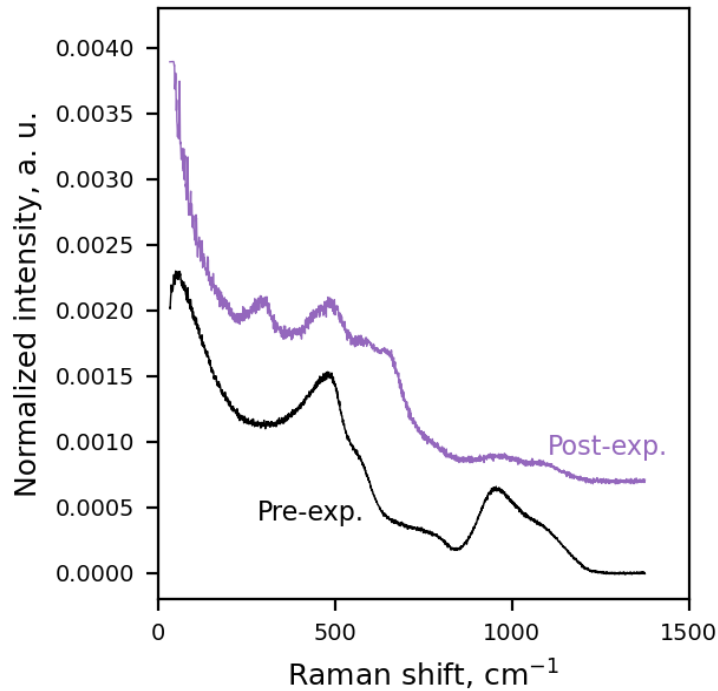
340 To check for crystallization during the low-temperature viscosity experiments, we compare the  
 341 Raman spectra of DR07 before and after the low-temperature viscosity experiments ([Fig. 5](#)).  
 342 Before the experiment, the DR07 glass Raman spectrum exhibits three main broad bands near

343 80, 490 and 1000  $\text{cm}^{-1}$ . The first is the Boson peak assigned to transverse acoustic vibrational  
344 modes promoted in silicate glasses by cooperative inter-tetrahedral vibrations (Buchenau et al.  
345 1986; Malinovsky and Sokolov 1986; Hehlen et al. 2002). This is a universal signature of the  
346 glassy state (Malinovsky and Sokolov 1986), that tends to quickly disappear in the presence of  
347 crystals (e.g., Takahashi et al. 2009). The second is the inter-tetrahedral T-O-T (T = Si, Al)  
348 vibrations in the aluminosilicate network. The third is assigned to intra-tetrahedral T-O  
349 stretching vibration (Mysen et al. 1982; McMillan 1984; Le Losq et al. 2014).

350 After the low temperature viscometry, the recovered DR07 sample does not show the Boson  
351 peak anymore (Fig. 5). This indicates that some degree of crystallization must have occurred  
352 during the viscosity measurements near the glass transition.

353 We also acquired Raman spectra on all the other samples quenched during the initial glass  
354 preparation (see supplementary Fig. A1). Although microscopic inspection of the samples  
355 appeared crystal-free, and although the Raman spectra above 200  $\text{cm}^{-1}$  may not present sharp  
356 peaks typical of crystals, no Boson peak was visible. This indicates that initial glass samples  
357 other than DR07 contained crystals at a sub-nanometric to nanometric scale (see Discussion).

358 Interestingly, we did not succeed to obtain a pure glass for the DR17 phonolite despite a very  
359 similar chemical composition with DR07. This may be explained by the slightly higher  $\text{Na}_2\text{O}$   
360 and  $\text{Al}_2\text{O}_3$  contents that could be responsible for a different behavior. This testifies that  
361 basaltic, tephri-phonolitic and phonolitic crystal-free glass samples of the required size for  
362 low temperature viscosimetry are extremely difficult to obtain.



363

364 **Figure 5.** *Uncorrected Raman spectra of DR07 phonolite products pre (black curve)- and post*  
 365 *(purple curve)-experiments at low-temperature.*

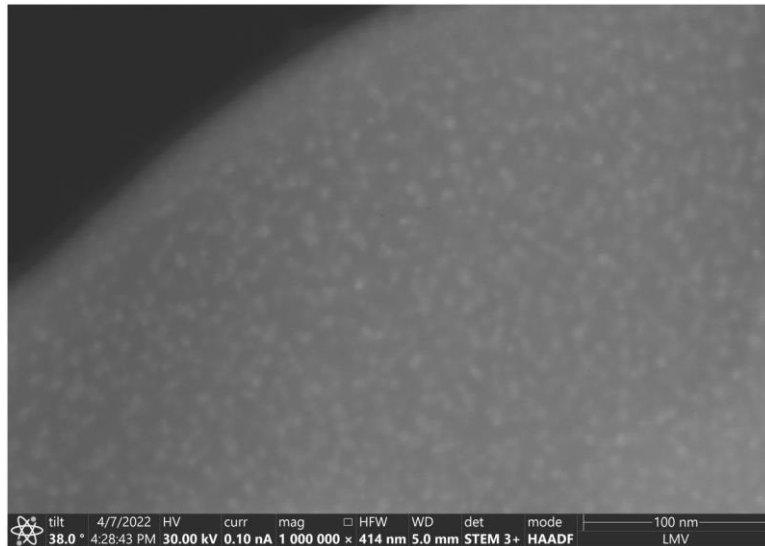
366

## 367 **4. Discussion**

### 368 **4.1. Nanolite crystallization during low-temperature measurements**

369 DR07 is the only sample for which we obtained a glassy piece large enough to perform near-  
 370  $T_g$  viscometry. However, during the experiment, the sample seems to have undergone some  
 371 degree of crystallization as evidenced by the disappearance of the Boson peak (**Fig. 5**). To  
 372 confirm that crystallization occurred we acquired a scanning transmission electron microscope  
 373 (STEM) image with the Helios 5 (ThermoFisher Scientific) scanning electron microscope  
 374 coupled with a focused ion beam (Xe plasma FIB-SEM) at LMV. Images show the presence  
 375 of homogeneously distributed brighter particles at the nanoscale (**Fig. 6**). Similar particles with  
 376 a size ranging from 5 to 30 nm have been observed and characterized as nanolites ([Di Genova](#)  
 377 [et al. 2017, 2018, 2020](#)).

378 The effect of crystallization on viscosity may reach several orders of magnitudes (e.g., [Lejeune](#)  
379 [et Richet, 1995](#); [Costa et al. 2009](#); [Mader et al. 2013](#)), particularly in presence of microlites  
380 ([Del Gaudio et al. 2013](#)) or nanolites ([Di Genova et al. 2020](#); [Le Losq et al. 2021](#)). In the present  
381 case, we observe neither a deviation of the viscosity measurements as a function of time nor  
382 the apparition of a non-Newtonian behavior. Therefore, it seems that the presence of nanolites  
383 only has limited effects on our measurements. However, this finding suggests that nanolite  
384 crystallization happens very quickly in melts of iron-bearing alkaline compositions and even  
385 in silica-rich melts such as phonolitic. To check for this, we placed a DR07 glassy sample in  
386 an annealing furnace and setup the temperature at the glass transition, with a particular care in  
387 avoiding overshooting while checking the sample temperature with a Pt-PtRh10%  
388 thermocouple placed in contact with the sample. The Raman spectra of the retrieved sample  
389 also showed the presence of crystals in the glass after annealing and is identical to the post-  
390 experiment spectra ([Fig. 5](#)). These observations indicate that viscosity measurements for such  
391 compositions in the undercooled temperature domain that were presented in previous studies  
392 ([Giordano et al. 2000, 2005](#); [Giordano and Dingwell 2003](#); [Le Losq et al. 2015a](#); [Whittington](#)  
393 [et al., 2000, 2001](#)) may probably have been also affected by nanolite formation. We add that  
394 because diffusion increases with decreasing Si content ([Zhang and Ni 2010](#)), this effect is  
395 expected to be even stronger for tephri-phonolitic and basanitic composition, preventing pure  
396 glass samples from being obtained. Great care should therefore be taken during low  
397 temperature viscometry even for phonolitic melt, and Raman spectra down to the Boson peak  
398 wavelength range should be systematically acquired.



399

400 *Figure 6. Scanning transmission electron microscope image of the DR07 sample post low-*  
401 *temperatures viscosity measurements.*

402

#### 403 **4.2. Melt viscosity and comparison with models**

404 A discrepancy between experimental measurements and parametric viscosity models, as  
405 observed here ([Fig. 4](#)), has also been previously reported for alkaline composition by [Le Losq](#)  
406 [et al. \(2015a\)](#). There are two likely causes for the significant difference between the GRD  
407 model and the measured viscosity at low temperature. First, unlike at high temperatures, in  
408 undercooled melts, important and nonlinear variations in melt configurational entropy result  
409 from changes in melt composition and drive important changes in viscosity ([Richet 1984](#);  
410 [Neuville and Richet 1991](#); [Le Losq et al. 2021](#)). It is challenging for parametric models to  
411 reproduce such strongly nonlinear behavior of viscosity variations and so, this could explain  
412 the discrepancy between the GRD model and our low-temperature viscosity measurements.  
413 Secondly, the GRD model has been calibrated with a restricted alkaline basanite-to-phonolite  
414 compositional data subset ([Giordano et al. 2008](#)). This could reduce the accuracy of the  
415 viscosity prediction of this model for such compositions. This also explains why the Vesuvius

416 melt is better reproduced as this composition was part of the database used to parametrize the  
417 GRD model (Giordano et al. 2008).

418 Additional comparison has been performed between our viscosity measurements and the  
419 predictions from the Hui and Zhang (2007) model. The HZ model is in relatively good  
420 agreement with our viscosity measurements at high-temperatures (Fig. 4A) whereas it  
421 significantly underestimates by  $\sim 1.3 \log_{10}$  units the viscosity of phonolitic melts at low-  
422 temperatures (Fig. 4B). Similar differences are observed between the HZ model and the  
423 measurements made on the Erebus (Le Losq et al. 2015a) and Vesuvius GP79 (Giordano et al.  
424 2009; Le Losq et al. 2015a) samples. According to Hui and Zhang (2007), the model could be  
425 improved in part by adding new viscosity measurements at low-temperatures and by  
426 considering the effect of ferric and ferrous iron. Also, as for the GRD model, the HZ model is  
427 a general model leading to an increase in uncertainty for specific compositions.

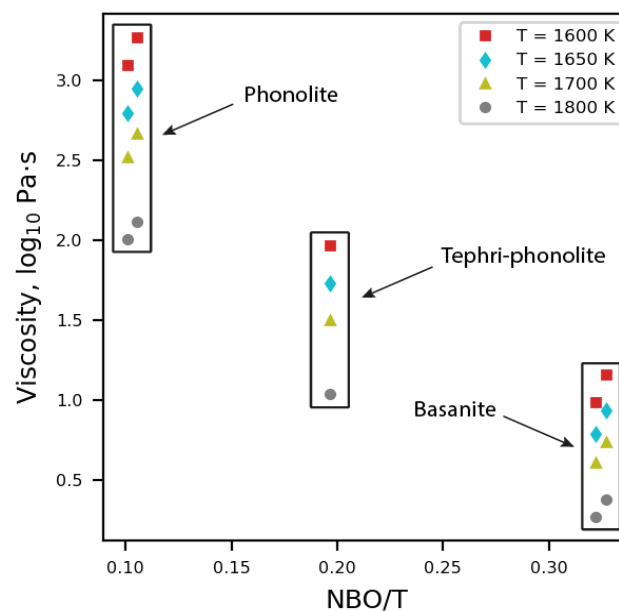
428 In order to refine such pre-existing models and thus improve their accuracy in the viscosity  
429 prediction, it is important to keep measuring natural samples. The present viscometry data  
430 could then be used to improve such models.

431

### 432 **4.3. Polymerization and viscosity**

433 The relationships between the super-liquidus viscosity and melt polymerization, in terms of T-  
434 O-T bridging as quantified by the chemically-derived, oxidized NBO/T (Eq. 1), are illustrated  
435 in Figure 7. At given temperature, basanite melts show the lowest viscosity ( $< 1.2 \log \text{ Pa}\cdot\text{s}$ )  
436 and the highest NBO/T (0.32 to 0.33) whereas phonolite melts have the highest viscosity ( $> 2$   
437  $\log \text{ Pa}\cdot\text{s}$ ) and lowest NBO/T (0.10 and 0.11) (Fig. 7). Accordingly, the tephri-phonolite sample,  
438 having an intermediate chemical composition (Fig. 2 and Table 2), falls in between the basanite  
439 and phonolite compositions (Fig. 7). The fact that the highest viscosity data are obtained for  
440 the samples with the lowest NBO/T, and inversely, is an expected pattern (Mysen et al. 1982,

441 1985; Scarfe et al. 1987; Mysen and Richet 2019). A higher degree of polymerization (lower  
 442 NBO/T) will lead to lesser solutions for performing the cooperative rearrangements of the  
 443 molecular sub-units necessary for melt viscous flow. This results, according to the Adam-Gibbs  
 444 theory of viscous flow, in a lower melt configurational entropy and, hence, in a higher melt  
 445 viscosity (Adam and Gibbs, 1965; Richet, 1984). Such variations in melt polymerization with  
 446 composition, albeit expected, may have an important impact on the eruptive style (see  
 447 Implications).



448  
 449 **Figure 7.** Variations of the viscosity ( $\log_{10} \text{Pa}\cdot\text{s}$ ) at a given temperature relative to composition  
 450 as represented by the ratio of non-bridging oxygens to tetrahedrally-coordinated cations  
 451 ( $\text{NBO}/\text{T}$ ). See also **Table 2** and **3** for chemical compositions and viscosity results, respectively.

452  
 453 **4.4. Influence of iron redox state and volatile concentration on melt viscosity**

454 All the presently reported experiments were performed in air. In addition, we measured the  
 455 viscosity of volatile-free melts, as they degassed upon preparation. Therefore, to provide  
 456 accurate viscosity estimates of the basanite, tephri-phonolite and phonolite melts in a natural  
 457 context, we should take into account the influence of both water and iron oxidation state on the



458 viscosity of the silicate melts. The reduction to Fe<sup>2+</sup> can lead to a decrease in melt viscosity of  
459 0.2 - 0.5 log units at super-liquidus conditions, which can reach up to 1.5 log units at  
460 undercooled temperature conditions (Dingwell and Virgo 1987; Dingwell 1991; Liebske et al.  
461 2003; Drucwrr 2007; Chevrel et al. 2013). Water also strongly participates in lowering the  
462 viscosity of alkaline magmas (e.g., Whittington et al., 2000, 2001).

463 To quantify the effect of water, we here follow the methodology of Le Losq et al. (2015a) who  
464 calculated the relative effect of water on the B and C terms of the VFT equation (Eq. 3) for  
465 phonolite melts, based on previously published data from Whittington et al. (2001). In that way  
466 we use the following equation:

467

$$468 \log_{10} \eta = A + (B^{anh} + K_1 * C_{H_2O} + K_2 * C_{H_2O}^2) / (T - (C^{anh} + K_3 * C_{H_2O} + K_4 * C_{H_2O}^2)), (4)$$

469

470 where K<sub>1</sub>, K<sub>2</sub>, K<sub>3</sub> and K<sub>4</sub> are the parameters of the polynomial functions that describe the effect of the  
471 water concentration in wt%, C<sub>H<sub>2</sub>O</sub>, on the value of the parameters B and C. B<sup>anh</sup> and C<sup>anh</sup> are the VFT B  
472 and C parameters of the anhydrous melt. Regarding the basanite composition no viscosity  
473 measurements at undercooled temperatures were obtained for this study. Hence, we calculated the VFT  
474 parameters of the anhydrous melt by combining our high-temperature data with the low- temperatures  
475 data from Whittington et al. (2000) obtained for an iron-free basanite melt (Table 4). To calculate the  
476 K<sub>1</sub>, K<sub>2</sub>, K<sub>3</sub> and K<sub>4</sub> parameters, we then used then used the VFT parameters published by Whittington et  
477 al. (2000) for a hydrous iron-free basanite. Table 5 reports the K<sub>1</sub>, K<sub>2</sub>, K<sub>3</sub> and K<sub>4</sub> parameters from Le  
478 Losq et al. (2015a).

479

480

481

482

483

484 **Table 4.** Parameters obtained for Arrhenius relation at high-temperature (Eq. 2) and VFT  
 485 parameter over the entire temperature range (Eq. 3). The VFT parameters calculated for the  
 486 basanite DR10 result from a combination of the high temperatures data from this study and  
 487 low-temperatures data from Whittington et al. (2000) (see Discussion).

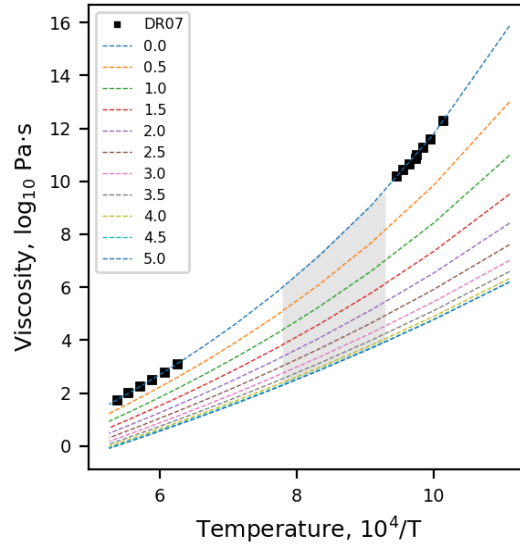
Sample	Temperature range (K)	Viscosity range (log <sub>10</sub> Pa·s)	A <sub>Arr</sub>	E <sub>aArr</sub> (kJ mol <sup>-1</sup> )	RMSE <sub>Arr</sub>	A <sub>VFT</sub>	B <sub>VFT</sub>	C <sub>VFT</sub>	RMSE <sub>VFT</sub>
DR01	1600 - 1855	0.99 - 0.11	-5.34	-44.37	0.003	-	-	-	-
DR10	1600 - 1810	1.16 - 0.34	-5.6	-46.57	0.015	-3.5	4448	645	0.05
DR21	1600 - 1860	1.97 - 0.89	-5.8	-48.23	0.011	-	-	-	-
DR17	1500 - 1855	3.89 - 1.91	-6.46	-53.72	0.008	-	-	-	-
DR07	990 - 1860	12.30 - 1.75	-	-	-	-3.9	7572	517	0.04

488  
489

490 **Table 5.** K parameters for estimating the effect of water on the parameters B and C of the VFT  
 491 equation (Eq. 4) for the phonolite (from Le Losq et al. 2015a) and the basanite melts.

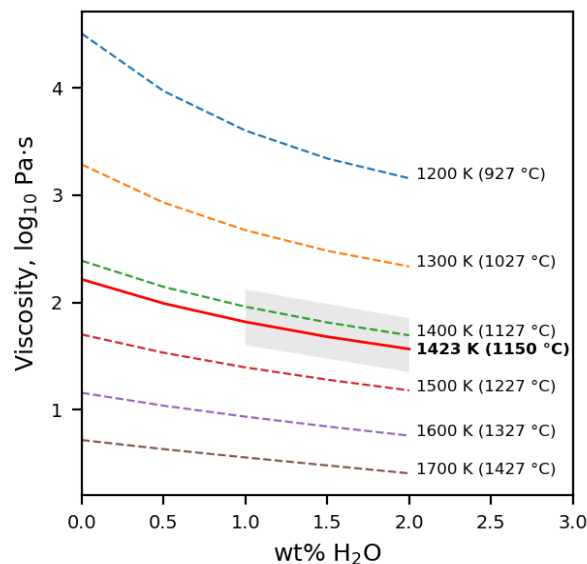
Parameter	Values for phonolite DR07	Values for basanite melts
K <sub>1</sub>	-455.52	403.15
K <sub>2</sub>	32.626	-133.87
K <sub>3</sub>	-110.61	-140.77
K <sub>4</sub>	13.241	32.043

493



494

495 **Figure 8.** Viscosity ( $\log_{10} \text{Pa}\cdot\text{s}$ ) as a function of inverse temperature ( $\text{K}^{-1}$ ) for the DR07  
 496 phonolite. Measured viscosity values (squares) are shown together with values predicted  
 497 (curves) by Eq. 4 for different water concentrations (wt %). Numbers in the label refer to water  
 498 content in wt% and the expected magmatic conditions are indicated by the grey box. The 4.5  
 499 and 5.0 wt% water curves are superposed. Error bars are smaller than the symbol size.



500

501 **Figure 9.** Isothermal viscosity as a function of water concentration (wt%) for the DR10  
 502 basanite. The predicted magmatic conditions proposed by Berthod et al., (2021b) are indicated  
 503 by the red line (1150 °C) and its uncertainty by the grey box ( $\pm 50$  °C).

504 We present the effect of water on the phonolite melt (DR07) in **Figure 8** and on the basanite  
505 melt (DR10) in **Figure 9**. As expected, the water strongly lowers the viscosity for both melts.  
506 The addition of 1 wt% of water reduces the viscosity from 2.50 to 1.68  $\log_{10}$  Pa·s at 1700 K  
507 for the DR07 phonolite. The decreasing effect of water on viscosity is higher at lower  
508 temperatures, where the difference between anhydrous and hydrous melts may reach up to 4  
509 orders of magnitude at 900 K (**Fig. 8**). In contrast, the addition of 1 wt% of water decreases the  
510 viscosity of the basanite melt by almost 0.4  $\log_{10}$  units at 1700 K (**Fig. 9**). These results in  
511 agreement to those reported by [Whittington et al. \(2000, 2001\)](#). Hence, our results confirm that  
512 (i) further addition of water results in a progressively less important decrease in viscosity and  
513 (ii) the effect of water is more effective at reducing the viscosity of polymerized melts  
514 (phonolites) than for depolymerized melts (basanites). As it has been previously demonstrated,  
515 the chemical composition of the melt affects water solution mechanisms, resulting in different  
516 effects on melt structure (e.g., [Whittington et al. 2000, 2001](#); [Xue and Kanzaki 2004, 2006](#);  
517 [Cody et al. 2005](#); [Mysen and Cody 2005](#); [Mysen and Richet 2005](#); [Giordano et al. 2009](#); [Le](#)  
518 [Losq et al. 2015b](#)). For polymerized melts, a common and simple water solution mechanism is  
519 the reaction of water molecules with the silicate network, breaking T-O-T bonds (with T = Si,  
520 Al) (e.g., see [Mysen and Richet 2019](#); [Le Losq et al. 2015b](#)). In contrast, for depolymerized  
521 melts, silicate network and network modifiers (e.g., Ca, Mg) may be implied into the  
522 mechanism leading to slight change of the melt polymerization ([Xue and Kanzaki 2004](#);  
523 [Moretti et al. 2014](#); [Le Losq et al. 2015b](#)). For the present basanite and phonolite melts that  
524 present relatively low NBO/T, the dissolution of water is expected to result in melt  
525 depolymerization, but a stronger effect is expected for phonolites as the mechanism implying  
526 reaction between T-O-T bonds and water should be more active. This can explain observations  
527 made on viscosity (**Figs. 8, 9**).

528 Thus, the reported values for the viscosity of the silicate melts represent upper limits for  
529 aluminosilicate melts due to the effect of water combined with the iron oxidation state.

530

#### 531 **4.5. Comparison of phonolites viscosities from different areas**

532 We can compare the viscosity of the phonolites from Mayotte (DR07 and DR17) to data from  
533 previous studies on similar Al- and alkali-rich compositions. For this, the viscosity  
534 measurements of the phonolitic Erebus (Le Losq et al. 2015a) and Vesuvius GP79 (Giordano  
535 et al. 2009; Le Losq et al. 2015a) magmatic melts have been selected (Fig. 3). The viscosity of  
536 the Vesuvius GP79 sample is lower by an order of magnitude compared to the phonolite from  
537 the East-Mayotte submarine volcanic chain. This may be explained by a lower SiO<sub>2</sub> content  
538 (54 wt% against 59 wt%). Despite a similar composition in SiO<sub>2</sub> and alkali elements (Fig. 2  
539 and Table 2), the viscosity of DR07 differs from the Erebus one by about 0.5 orders of  
540 magnitude. The phonolites of Mayotte (DR07 and DR17) are richer by 1 to 2 wt% of iron  
541 compared to the Erebus sample (Table 2). In contrast to Chevrel et al. (2013), a decrease in  
542 melt viscosity with increasing iron concentration is not observed. The observed viscosity  
543 difference more probably finds its origin in variations in the concentrations of a few elements  
544 such as K, Mg and Ca between the phonolites of Mayotte and Erebus. Indeed, the DR07 and  
545 DR17 samples are slightly more enriched in K<sub>2</sub>O (Table 2). In aluminosilicate compositions,  
546 addition of K<sub>2</sub>O can lead to decreasing the melt configurational entropy and so, to increasing  
547 its viscosity at constant temperature (Le Losq et al. 2013, 2017, 2021; Robert et al. 2019).  
548 Therefore, the difference in K<sub>2</sub>O between the Erebus and Mayotte phonolites could explain the  
549 observed viscosity distinctions. In addition, the Erebus sample contains a bit more alkaline-  
550 earth elements (Mg and Ca), that may act as *network modifiers* and favor the creation of non-  
551 bridging oxygen (NBO) atoms and, hence, a decrease in viscosity (Mysen et al. 1980; Mysen  
552 1995; Richet et al. 1984; Stebbins et al. 1992; Stebbins and Xu 1997). Hence, it seems

553 consistent that the DR07 phonolite from Mayotte has a higher viscosity than the Erebus sample.  
554 As explained previously, no viscosity measurements at undercooled temperatures could be  
555 obtained for the DR17 phonolite. However, according to the viscosity at the super-liquidus  
556 temperatures and the enrichment in alkali (Figs. 2 and 3) we may assume that the viscosity at  
557 undercooled temperatures will be higher than the DR07, Erebus and Vesuvius melts.

558

## 559 **5. Implication on eruption dynamics**

560 Upon magma ascent in a conduit, the style of a volcanic eruption depends on whether the  
561 magma fragmentation threshold is crossed (explosive) or not (effusive). This threshold will  
562 depend on the magma viscosity but also on the ascending rate and bubble overpressure (e.g.,  
563 [Dingwell et al. 1996](#); [Papale 1999](#); [Ma et al. 1999](#); [Pistolesi et al. 2011](#); [Gonnermann et al.](#)  
564 [2011](#); [Gonnermann 2015](#)). Generally, in magmas with high viscosity (silica-rich composition),  
565 the enhanced resistance to bubble expansion favors bubble overpressure and reduces the ability  
566 of volatiles to stream out of the melt. This in turn promotes more efficient magma  
567 fragmentation and a greater explosivity of eruptions. Of course, many exceptions break such  
568 general rules, such as basaltic Plinian eruptions (e.g., [Moitra et al. 2018](#)) and Si-rich lava flows  
569 (e.g., [Fink 1983](#); [Farquharson et al. 2015](#); [Prival 2021](#)). Additionally, eruption style can switch  
570 between effusive and explosive activities during an eruption due to variations in magma  
571 composition, degassing, and crystallization that all drive magma viscosity (e.g., Vesuvius  
572 [Giordano et al. 2009](#); [Myers et al. 2021](#), [Andújar and Scaillet 2012](#); [Popa et al. 2021](#)). Attempts  
573 have been done to define the viscosity threshold to discriminate explosive to effusive eruptions,  
574 but no consensus has yet been achieved ([Papale 1999](#); [Di Genova et al. 2017](#); [Wadsworth et al.](#)  
575 [2018](#)).

576 Our results show that basanite melts have the lowest viscosity of the series followed by the  
577 tephri-phonolite and the phonolite melts and that the presence of water reduces the viscosity.

578 Assuming that the eruptive temperature ( $T_e$ ) of the Mont Fani Maoré basanite lavas is around  
579 1150 °C (Berthod et al., 2021b), and that water content could range from 1.0 to 2.3 wt%  
580 (Berthod et al. 2021b), we thus expect lava viscosity upon eruption to range from 1.5 to 2.5  
581  $\log_{10}$  Pa·s (Fig. 8). Such a viscosity range is close to, for example, the viscosity of subaerial  
582 pahoehoe basaltic lavas from Hawaii (Chevrel et al. 2018) or submarine basaltic lava flows  
583 (McClinton et al. 2014). This low viscosity range is therefore consistent with the effusive  
584 eruption style and morphologies (pillow lavas and pahoehoe lavas) of the lava flow field as  
585 observed at Mont Fani Maoré (Feuillet et al., 2021; Berthod et al. 2021b). However, several  
586 questions remain. Why did the basanite melt come out from the Mont Fani Maoré rather than  
587 ascend upwards to erupt in the Horseshoe site, where its hypothesized magmatic storage is  
588 located (Lemoine et al. 2020; Feuillet et al. 2021)? Does the low viscosity of the basanite melt  
589 favored a lateral pathway? Could the low viscosity favors high flux rates ( $150 - 200 \text{ m}^3 \text{ s}^{-1}$   
590 averaged over the first year, Berthod et al. 2021b) and the long duration eruption if the volume  
591 was there?

592 Around the Horseshoe volcanic site of the Mayotte volcanic chain, a large diversity of edifice  
593 morphologies and deposits textures corresponding to both phonolitic effusive and explosive  
594 eruptions have been observed (REVOSIMA, 2022; Puzenat et al. 2022; Gurioli et al. 2023;  
595 Komorowski et al. 2023). Such diversity can be explained by the fact that, depending on magma  
596 ascent rates, eruptive temperatures and initial water contents, phonolite magmas can be  
597 involved in a broad range of magmatic effusive and explosive activity (e.g., see Andújar and  
598 Scaillet 2012). Indeed, the Mayotte phonolite magmas evolved from basanite magma by ~80%  
599 of fractional crystallization (Berthod et al. 2021a). They are expected to present a high viscosity  
600 at expected eruption temperatures (typically in the 800-1000 °C range, e.g. see Andújar and  
601 Scaillet 2012). However, phonolite magmas are also expected to contain a large amount of  
602 dissolved water. If assuming a phonolite melt stored in a deep magma chamber at  $T_e = 800 -$

603 1000 °C and containing ~5 wt% water (Andújar and Scaillet 2012), this magma could have a  
604 viscosity as low as 2.0 - 3.2 log<sub>10</sub> Pa·s (Fig. 8), not that far from that previously estimated for  
605 the basanite melt. However, upon migration toward the surface, vigorous water exsolution is  
606 expected to occur. According to the composition of the residual glass of the submarine  
607 phonolites from Mayotte (encountered at 1,300 to 1,600 m b.s.l.), the residual water content in  
608 erupted phonolite glass ranges between 0.8 and 1.2 wt% (Thivet, personal communication).  
609 Considering such values, the viscosity of the phonolite melts may increase of up to 4 to 7 orders  
610 of magnitudes upon ascent (Fig. 8). In parallel, the significant water exsolution will lead to the  
611 formation of a magmatic foam of low density in the conduit, reinforcing the buoyant force that  
612 pushes the magma out toward surface. This effect is particularly promoted in silica-rich melts  
613 like phonolites, because their high viscosity favors entrapment of the bubbles (Thomas et al.  
614 1994; Gardner et al. 1996). Therefore, exsolution of water from such magmatic liquids often  
615 results in foaming, largely enhancing the probability of brittle fragmentation of the foam (Jones  
616 et al. 2019; Scheu and Dingwell 2022) in the conduit (e.g., Dingwell, 1996; Papale 1999;  
617 Gonnerman, 2015). Therefore, the amount of water initially stored in the chambers will  
618 determine the eruptive styles. The occurrence of both effusive and explosive phonolite  
619 eruptions at Mayotte, testified by the observed deposits and products, indicate that a broad  
620 range of storage conditions and degassing history exist along this volcanic chain, leading to  
621 open questions regarding the future events that could imply phonolite melts.

622

## 623 **6. Conclusion**

624 This study provides precise quantification of the temperature-viscosity relationship for basanite  
625 to phonolite melts through a large temperature range and discusses the effect of dissolved water  
626 on melt viscosity. We show that at eruption temperatures basanite lavas are more propitious to  
627 erupt in an effusive manner while phonolite may erupt effusively or explosively. However, we



628 do not account for the effect of bubbles and crystals on magma viscosity. Future studies should  
629 accounted for their effect to provide a holistic view of magma rheology which directly impacts  
630 the dynamics and eruption style as well as lava flow emplacement. For this, detailed sample  
631 texture analyses should be undertaken on collected samples to quantify the bubble and crystal  
632 content, shape, size and distribution that are necessary to constrain magma suspension rheology  
633 (see [Harris and Allen 2008](#); [Mader et al. 2013](#); [Kolzenburg et al. 2022](#)). Measuring the  
634 viscosity of magmatic liquids along the Mayotte alkali magmatic differentiation trend  
635 combined with known volatile content, and crystal and bubbles characteristics, is essential to  
636 quantify the evolution of magma viscosity during ascent rate, outgassing and crystallization,  
637 which directly influence the prevailing conditions determining the eruptive style (whether it is  
638 effusive or explosive).

639

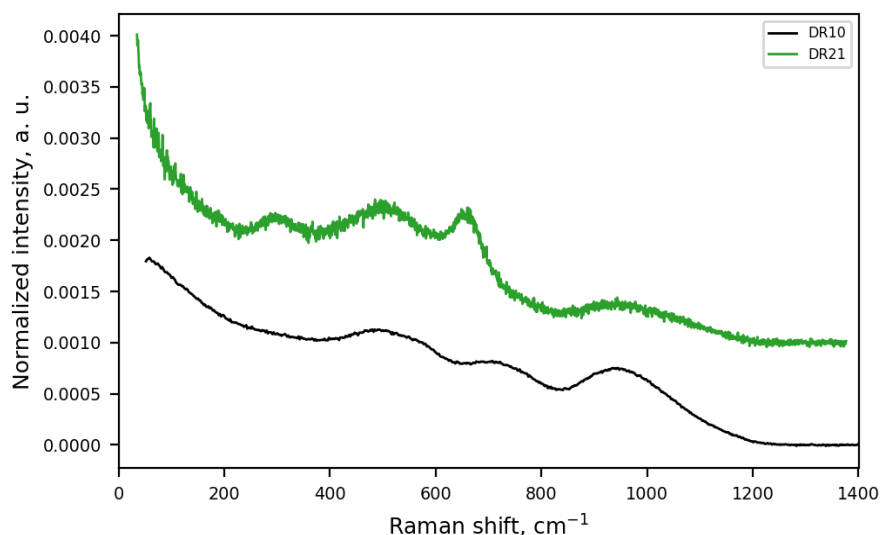
## 640 **Acknowledgments**

641 MAYOBS 1 campaign was funded by the CNRS-INSU TELLUS MAYOTTE program  
642 (SISMAYOTTE project). MAYOBS 1, 2, 4 and 15 campaigns were conducted by several  
643 French research institutions and laboratories (IPGP/CNRS/BRGM/IFREMER/IPGS). All  
644 marine operations are performed as part of the MAYOBS set of campaigns  
645 (<https://doi.org/10.18142/291>) and we thank the captains and crews of the R/V Marion  
646 Dufresne (TAAF/IFREMER/LDA), R/V *Pourquoi Pas?* (GENAVIR/IFREMER, SHOM). We  
647 thank the mission chiefs of the MAYOBS campaigns (E. Rinnert, N. Feuillet, Y. Fouquet, S.  
648 Jorry, I. Thinon, E. Lebas, F. Paquet) and of the GeoFlamme oceanographic campaign (CNFH;  
649 PI C. Cathalot, E. Rinnert, N. Feuillet) for conducting marine operations that benefited this  
650 study and provided dredge samples. We also thank additional scientists on board the MAYOBS  
651 cruises that conducted the dredge operations and processed the samples (P. Besson, M. Bickert,  
652 P. Burckel, B. Caron, C. Deplus, S. Hidalgo, A. Le Friant, S. Nowak). We thank A. Peltier (OVPF-

653 IPGP) and C. Mucig (BRGM Mayotte) respectively the Operational Leader and Co-leader of the  
654 REVOSIMA. We thank the scientists of REVOSIMA consortium for access to data and for  
655 discussions during the Scientific and Technical Committee meetings. Analyses were funded by  
656 the Service National d'Observation en Volcanologie (SNOV, INSU) and the Réseau de  
657 Surveillance Volcanologique et Sismologique de Mayotte (REVOSIMA), a partnership  
658 between the Institut de Physique du Globe de Paris (IPGP), the Bureau de Recherches  
659 Géologiques et Minières (BRGM), and the Observatoire Volcanologique du Piton de la  
660 Fournaise (OVPF-IPGP), the Centre National de la Recherche Scientifique (CNRS), and the  
661 Institut Français de Recherche pour l'Exploitation de la Mer (IFREMER). Since June 2019, all  
662 activities on Mayotte are funded by le ministère de l'Enseignement Supérieur, de la Recherche  
663 et de l'Innovation (MESRI), le Ministère de la Transition Ecologique (MTE), le Ministère des  
664 Outremer (MOM), le Ministère de l'Intérieur (MI), and le Ministère des Armées with the  
665 support of the DIRMOM (Direction Interministérielle aux Risques Majeurs en Outremer) and  
666 the MAPPPROM (Mission d'appui aux politiques publiques pour la prévention des risques  
667 majeurs en Outremer). We thank the IPGP for general funding to the Observatoires  
668 Volcanologiques et Sismologiques (OVS). The data contributes to the Service National  
669 d'Observation en Volcanologie (SNOV). The authors would like to thank IFREMER for their  
670 welcome during the sampling and E. Humler for his support and national funding coordination  
671 (CNRS, REVOSIMA). The authors also thank Thivet S. for the discussion about the water  
672 content within the submarine volcanic samples from Mayotte. Raman and viscosity  
673 measurements were supported *in kind* by funds from the IPGP Geomaterial group. This work  
674 contributes to IdEx Université de Paris ANR-18-IDEX-0001. This project has received funding  
675 from the Institut de Recherche pour le Développement (IRD) and the French Government  
676 Laboratory of Excellence initiative no. ANR-10-LABX-0006. This is **Laboratory of Excellence**  
677 **ClerVolc contribution n° X.**

## 678 Appendix

679 **Figure A1.** Uncorrected Raman spectra of (a.) DR01 basanite (black curve) and (b.) DR21  
680 tephri-phonolite (green curve) products. No Raman Boson peaks are observed on these  
681 samples between 50 and 80  $\text{cm}^{-1}$ .



682

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