1	Experimental constraints on the rheology, eruption and
2	emplacement dynamics of analog lavas comparable to Mercury's
3	northern volcanic plains
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### 20

## 21 Key Points:

22	•	New viscosity data for Mercury northern volcanic plains lavas are presented.
23	•	Mercury lavas show shear thinning behaviour with a decrease of viscosity of ca. 1 log
24		unit as shear rate ( $\dot{\gamma}$ ) varies from 0.1 to 5.0 s <sup>-1</sup> .
25	•	Heat loss during lava flow and emplacement implies that high effusion rates, >10000
26		m <sup>3</sup> /s, are required to cover large distances as observed by MESSENGER (NASA).

#### 27 Abstract

We present new viscosity measurements of a synthetic silicate system considered an analogue 28 for the lava erupted on the surface of Mercury. In particular, we focus on the northern volcanic 29 plains (NVP), which correspond to the largest lava flows on Mercury and possibly in the Solar 30 System. High-temperature viscosity measurements were performed at subliquidus conditions 31 (1569–1502 K) to constrain the viscosity variations as a function of crystallinity (from 0 to 28%) 32 33 and shear rate (from 0.1 to 5 s<sup>-1</sup>). Melt viscosity shows moderate variations (4 - 16 Pas) in the temperature range 1736–1600 K. Experiments performed below the liquidus temperature show 34 an increase in viscosity as shear rate increases from 0.1 to 5 s<sup>-1</sup>, resulting in a shear thinning 35 behaviour, with a decrease in viscosity of ca. 1 log unit. The low viscosity of the studied 36 composition may explain the ability of NVP lavas to cover long distances, on the order of 37 hundreds of kilometres in a turbulent flow regime. Using our experimental data we estimate that 38 lava flows with thickness of 1, 5 and 10 m are likely to have velocities of 4.8, 6.5 and 7.2 m/s 39 respectively, on a 5° ground slope. Numerical modelling incorporating both the heat loss of the 40 lavas and its possible crystallization during emplacement allows us to infer that high effusion 41 rates (> 10000  $\text{m}^3/\text{s}$ ) are necessary to cover the large distances indicated by satellite data from the 42 MESSENGER spacecraft. 43

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## 48 **1 Introduction**

49	The eccentricity of the orbit of Mercury, in combination with the planet's vicinity to the
50	Sun, is responsible for its very long days (~ 59 terrestrial daytimes) and, locally, extremely high
51	surface temperatures. The daylight temperature at perihelion, estimated on the surface at the
52	equator, is ~700 K, whereas it decreases to ~350 K at 85°N. During the night, the lack of a
53	shielding atmosphere produces a high loss of thermal energy due to radiation and temperature
54	decreases to ~100 K [Paige et al., 1992; Vasavada et al., 1999].
55	The surface of Mercury is dominated by a secondary volcanic crust, the majority of
56	which formed between 4.2 and 3.5 Ga [Head et al., 2011; Weider et al., 2012; Denevi et al.,
57	2013; Byrne et al., 2016], with minor explosive volcanic activity until ~ 1.0 Ga [Thomas et al.,
58	2014]. Geochemical mapping using the X-Ray Spectrometer (XRS) and Gamma-Ray
59	Spectrometer (GRS) of the MErcury Surface, Space ENvironment, GEochemistry, and Ranging
60	(MESSENGER) spacecraft [Solomon et al., 2001] revealed that the volcanic crust is Mg-rich and
61	Al- and Ca-poor in comparison with terrestrial and lunar crustal material [Nittler et al., 2011;
62	Weider et al., 2012, 2015; Peplowski et al., 2015]. Mercury's crust is also strongly depleted in Fe
63	[Izenberg et al., 2014; Weider et al., 2015]. This is most likely due to extreme partitioning of
64	iron into the core [Hauck et al., 2013] during early differentiation of the planets under highly
65	reducing conditions (IW-3 to IW-7 with IW being the iron- wüstite oxygen fugacity buffer)
66	[Malavergne et al., 2010; McCubbin et al., 2012; Zolotov et al., 2013; Namur et al., 2016a]. The
67	extremely high sulfur contents measured by MESSENGER (1-3 wt.%; [Weider et al., 2015]) can
68	also be explained by differentiation under reducing conditions [Namur et al., 2016a], as sulfur
69	solubility in silicate melts increases with progressively reduced oxygen fugacity conditions

[*McCoy et al.*, 1999; *Berthet et al.*, 2009; *Zolotov et al.*, 2013; *Cartier et al.*, 2014; *Namur et al.*,
2016a].

The largest effusive events on Mercury occurred at the highest latitudes of the northern 72 hemisphere and are represented by lavas with the highest SiO<sub>2</sub>- and Al<sub>2</sub>O<sub>3</sub>-contents and the 73 lowest MgO-contents detected on the planet [Weider et al., 2015; Namur et al., 2016b]. These 74 lavas belong to a single smooth plain deposit referred to as the northern volcanic plains (NVP) 75 [Weider et al., 2012; Denevi et al., 2013], which was dominantly formed between 3.7 and 3.5 Ga 76 2011; [Head et al., 2011; Weider et al., 2012; Ostrach et al., 2015; Byrne et al., 2016] and covers 77  $\sim 6\%$  of the surface of the planet [Denevi et al., 2013]. A notable characteristic of this geological 78 sector of Mercury is that it contains well-preserved lava flows with a low crater density. Some of 79 these flows can be followed for distances exceeding 100 km [Byrne et al., 2013; Hurwitz et al., 80 2013], making them extremely useful to gain a better understanding of the dynamics of magma 81 emplacement on planetary surfaces in general, and on Mercury in particular. Another distinct 82 feature of these lava flows is their very high Na<sub>2</sub>O (up to 8 wt.%) [Peplowski et al., 2014, 2015], 83 which have been explained by 15–30 % melting of a plagioclase-bearing lherzolitic mantle 84 source [Namur et al., 2016b; Vander Kaaden and McCubbin, 2016] and low Al<sub>2</sub>O<sub>3</sub> contents, 85 86 responsible for the low viscosity of the lavas [Charlier et al., 2013; Sehlke and Whittington, 2015]. 87

Several authors have investigated the nature of lavas constituting the NVP through
morphological and compositional analyses [*Head et al.*, 2011; *Ostrach et al.*, 2015; *Weider et al.*,
2015], mineralogical analysis [*Namur and Charlier*, 2017; *Vander Kaaden et al.*, 2017], flow
modelling [*Byrne et al.*, 2013] and rheological measurements of analog lavas [*Sehlke and Whittington*, 2015]. The origin of such large lava flows and their lateral extent are controversial

and may be due to the low-viscosity of the lava [*Stockstill-Cahill et al.*, 2012] and/or very high
effusion rates [*Head et al.*, 2011; *Ostrach et al.*, 2015; *Sehlke and Whittington*, 2015; *Namur and Charlier*, 2017]. However, to the best of our knowledge, there is presently no work that takes
into consideration the possible variation of heat loss of the lava and its relationship with effusion
rates.

In order to better understand the mechanisms of lava emplacement in the NVP, we 98 provide new viscosity measurements using synthetic material with a composition that has been 99 estimated using the most recent XRS and GRS data from MESSENGER [Weider et al. 2015; 100 101 Peplowski et al. 2015; Table 1]. Viscosity measurements were performed at both superliquidus and subliquidus temperatures, at varied shear rates, and are combined with numerical models in 102 order to propose a hypothesis for the rheological behaviour, eruption and emplacement dynamics 103 of NVP lavas. We propose that the formation of very large lava flows may adequately be 104 explained by a combination of low viscosity (10-20 Pa s) and high effusion rate (> 10000  $\text{m}^3/\text{s}$ ). 105 In particular, we show that the effusion rates necessary to produce NVP lava flows are 106 comparable to those observed in large igneous provinces (LIPs) on Earth, which could be 107 consistent with formation of NVP lavas by adiabatic decompression of the mantle source [Namur 108 109 et al., 2016b]. Our work also shows the critical effect of low Al<sub>2</sub>O<sub>3</sub> and high Na<sub>2</sub>O contents on lava rheology and we believe that accurately measuring these elements and their variability 110 111 across the planet should be a priority target of the BepiColombo mission [Benkhoff et al., 2010]. 112

#### 113 **2 Starting materials and experimental techniques**

The elemental compositions of NVP lavas were obtained from XRS (normalized to Si) and GRS measurements and presented by *Nittler et al.* [2011], *Weider et al.* [2012; 2015] and

116	Peplowski et al. [2014, 2015]. They were recalculated by Namur et al. [2016b] on an oxide basis.
117	These authors combined individual maps of Mg/Si, Ca/Si, Al/Si and S/Si and only calculated
118	chemical compositions for pixels for which those four ratios were available assuming that the
119	sum of major oxides is 100 wt.%. The main advantage of this method is that Si contents do not
120	need to be arbitrarily fixed [Vander Kaaden and McCubbin, 2016]. According to Peplowski et al.
121	[2014, 2015], Na2O is high in NVP lavas. Consequently, Namur et al. [2016b] used a Na/Si ratio
122	of 0.20, similar to the average Na/Si ratio of NVP lavas presented by Peplowski et al. [2014].
123	This procedure resulted in the acquisition of a large compositional range for NVP lavas spanning
124	from 55 to 66 wt.% SiO <sub>2</sub> , 8 to 20 wt.% MgO, 3 to 9 wt.% CaO and 9 to 16 wt.% Al <sub>2</sub> O <sub>3</sub>
125	[Peplowski et al. 2015; Weider et al. 2015; Namur et al., 2016b].
126	As reported by Nittler et al. [2011] and Weider et al. [2015], Mercury's surface contains
127	high abundances of sulfur (1-3 wt.%). Sulfur solubility in silicate melts increases with
128	decreasing oxygen fugacity [Berthet et al., 2009; Cartier et al., 2014; McCoy et al, 1999; Namur
129	et al., 2016a]. Oxygen fugacity during mantle melting and volcanic eruptions on Mercury is
130	traditionally considered as being between IW-2 and IW-7 [McCubbin et al., 2012; Zolotov et al.,
131	2013] although new models show that most lavas were formed at IW-5.4±0.4 [Namur et al.,
132	2016a]. Therefore, the potential effect of sulfur on the rheology of Mercurian lavas might need to
133	be considered. However, it was demonstrated that S has a very minor effect on the
134	polymerization of silicate melts and, hence, on their rheology [Scaillet, 2015]. In addition, NVP
135	lavas contain the lowest S contents among Mercurian magmas (0.5 to 2.7 wt.% S with a median
136	value of 1.64 wt.% Namur et al., [2016a]). Therefore, in this study we consider that sulfur plays
137	a minor role in modulating the physical properties of NVP lavas.

According to the above considerations, for this experimental study, we prepared a S-free 138 representative composition of NVP lavas (Table 1). We concentrated on lava compositions from 139 the northernmost regions of NVP (>  $70^{\circ}$  North), which have the lowest Al<sub>2</sub>O<sub>3</sub>-contents [*Weider*] 140 et al., 2015] but also the highest Na<sub>2</sub>O contents [Peplowski et al., 2014] and which have not yet 141 been experimentally investigated for viscosity characterization. The synthetic composition was 142 143 prepared at the Petro-Volcanology Research Group laboratories of the University of Perugia (hereafter PVRG labs). Our composition has a ratio of non-bridging oxygen to tetrahedrally 144 coordinated cations equal to 0.89 reflecting its high degree of depolymerisation. This 145 corresponds well with compositions reported by Stockstill-Cahill et al. [2012] and Vander 146 Kaaden and McCubbin [2016]. When plotted in a total alkali versus silica (TAS) diagram, our 147 experimental composition lies between trachy-andesite and trachy-dacite fields, similar to the 148 compositions investigated by Sehlke and Whittington [2015] and Vander Kaaden and McCubbin, 149 [2016]. 150

Five hundred grams of glass were prepared by melting a mixture of oxides and 151 carbonates at 1873 K for 4 hours in a Pt80Rh20 crucible in air. Melting was performed in a 152 Nabertherm HT 04/17 MoSi<sub>2</sub>-heated box furnace (Nabertherm GmbH, Lilienthal, Germany). The 153 154 melt was poured on to a brass plate to quench. To ensure homogeneity, the quenched melt (glass) was crushed, re-melted and quenched again using the same technique. This technique ensures 155 156 compositional homogeneity of the glass [Vetere et al., 2015]. Qualitatively, high fluidity was 157 observed during quenching suggesting a low viscosity of the melt. Viscosity measurements have been performed in a Gero HTRV 70-250/18 high-158

temperature tube furnace with MoSi<sub>2</sub> heating elements (Gero GmbH, Neuhausen, Germany)

operating up to 2073 K at room pressure. Thermal ramps can be precisely controlled via

computer using the Eurotherm iTools v. 9.57.11 software (Eurotherm, Worthing, West Sussex, 161 UK). Viscosity measurements were performed with a rotational Anton Paar RheolabQC 162 viscometer head at the PVRG labs. This instrument consists of a sample-filled crucible and a 163 rotating measuring spindle that is immersed into the sample. The crucible hosting the silicate 164 melt is made of Pt80-Rh20 with an inner diameter of 37 mm (outer diameter 40 mm) and height of 165 70 mm. The spindle is made of Al<sub>2</sub>O<sub>3</sub> with a circular section of 12.2 mm in diameter and is fixed 166 with a standard collet chuck to the head of the viscometer. The lower end of the Al<sub>2</sub>O<sub>3</sub> rod is 167 sheathed by a tight-fitting Ptso-Rh20 foil (0.2 mm thick) in order to prevent any contamination of 168 the silicate melts during the experimental runs. The rotational viscometer allows measurements 169 under controlled shear rate ( $\dot{\gamma}$ ). This allows us to investigate possible shear thinning (an increase 170 of viscosity with decreasing  $\dot{\gamma}$ ) or shear thickening (a decrease of viscosity with decreasing  $\dot{\gamma}$ ) 171 effects. Methods and procedures described by Dingwell [1986] and Ishibashi [2009] were 172 applied in order to determine melt and melt + crystals viscosities. With this equipment, viscosity 173 can be measured in the range from 0.1 to 10<sup>5</sup> Pa s [Hess et al., 1996]. The viscometer was 174 calibrated against NIST 717a standard glass, for which the temperature-viscosity relationship is 175 accurately known (https://www.nist.gov). Reproducibility of measurements on the standard glass 176 is on the order of  $\pm 0.03$  log units. Since NVP melt viscosity was expected to be low, before 177 running experiments we calibrated the viscometer using a Wacker silicone standard having 178 viscosity of 10 Pas [Spina et al., 2016a, 2016b]. One hundred measurements were performed and 179 the results showed good reproducibility, with average values of  $9.7 \pm 0.3$  (standard deviation) Pa 180 181 s.

The furnace hosting the experimental charge is equipped with aluminium cooling heads.These are positioned on the top and the bottom openings of the furnace tube in order to prevent

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overheating of the viscometer head. Cooling is achieved by a continuous flux of cold water (20° 184 C). The furnace can move vertically using two pneumatic cylinders (details in Morgavi et al. 185 [2015]). This has two major advantages as it allows us to: (1) carefully prepare the experimental 186 geometry by precisely positioning the outer and inner cylinders outside the furnace; (2) bring the 187 furnace directly to the experimental temperature while the sample is still outside the furnace, 188 189 preventing the sample from undergoing the entire thermal ramp. Two cross-mounted Thorlabs single-axis translation stages with a standard micrometer 190 allow the correct positioning of the spindle and the crucible. An alumina (Al<sub>2</sub>O<sub>3</sub>) rod (length=600 191 mm; FRIATEC Aktiengesellschaft, Mannheim, Germany) is fixed to the lower part of the 192 structure holding the outer cylinder. Temperature was monitored using an in-house built S-type 193 thermocouple (Pt<sub>10</sub>Rh<sub>90</sub> vs. Pt) within an Al<sub>2</sub>O<sub>3</sub> sheath, positioned at the bottom of the crucible. 194 As the rotation of the viscometer prevents the use of thermocouples directly wired to a controller, 195 OMEGA wireless thermocouple transmitters UWTC-Series were employed (OMEGA 196

Engineering, INC., Stamford, Connecticut, USA). Uncertainty on temperature measurements ison the order of 0.5 K.

Prior to viscosity measurements, ca. 70 g of melt was stirred at 1773 K for 2 hours at strain rates ( $\dot{\gamma}$ ) of 5–10 s<sup>-1</sup>. This allowed for the complete removal of possible gas bubbles and the attainment of a compositionally homogeneous melt [*Dowty*, 1980; *Lofgren*, 1983; *Davis and Ihinger*, 1998; *Armienti*, 2008; *Iezzi et al.*, 2008, 2011; *Pupier et al.*, 2008; *Vetere et al.*, 2013a, 2015]. Samples used in the subliquidus temperature experiments were first melted at superliquidus conditions. The temperature was then decreased continuously to the required subliquidus value at a rate of 5 K/min. At the end of the experiments, samples were quenched by moving them into the cooled head of the furnace. The quench rate was of the order of 100 K/min
which was sufficient to avoid the formation of quench crystals.

#### 208 **3 Analytical methods**

209 Experimental samples were cored out from the outer cylinder after quenching, mounted in epoxy, ground flat and progressively polished using diamond paste for textural and chemical 210 analysis. The composition of phases, i.e. the starting glass (for superliquidus experiments), the 211 glass matrix and crystalline phases of run-products (for subliquidus experiments) were analyzed 212 213 using a CAMECA SX100 electron microprobe analyser (EMPA) at the University of Hannover (Germany). Analyses were performed with an accelerating voltage of 15 kV. For glasses, we 214 used a beam current of 8 nA and a defocused beam of 10 µ**□**. Mineral analyses were performed 215 with a beam current of 15 nA and a focused beam (1 µ**□**). Counting time on peak was 15–20 s 216 217 (7.5-10 s for background) for each element. For glasses and minerals, we used the following standards for Ka X-ray line calibration: albite for Na, orthoclase for K, wollastonite for Si and 218 219 Ca, Al<sub>2</sub>O<sub>3</sub> for Al, TiO<sub>2</sub> for Ti, MgO for Mg. Raw data were corrected with the CATZAF software and results are reported in Table 1 and Table S1-S2 (supplementary information). A high-220 resolution scanning electron microscope with field-emission gun (FE-SEM LEO 1525 - ZEISS), 221 installed at the Department of Physics and Geology (University of Perugia), was used to collect 222 back-scattered electron (BSE) images of the experimental charges. Additional details about the 223 analytical conditions of FE-SEM and EPMA, as well as data reduction procedures, are reported 224 in Vetere et al. [2015] and Namur et al. [2016a]. 225

Abundance and distribution of mineral phases in each experimental charge were determined on BSE images using the Image-ProPlus 6.0 software. This software was also used to calculate length-width aspect ratios (AS) of the crystals by applying an automatic ellipse fitting

229	procedure. When crystals were in contact, aspect ratios were measured manually. Details on the
230	analytical protocol for image analysis are given in previous works [Iezzi et al., 2008, 2011;
231	Lanzafame et al., 2013]. Representative BSE images are reported in Figure 1. In particular, the
232	estimation of mineral phase proportions (in area %) was made by linking grey-level values of
233	BSE images with compositions. No stereological correction was applied [Iezzi et al., 2008;
234	Vetere et al., 2010 and 2013a-b]. Magnification used in image acquisition ranged from 150× to
235	1600× depending on the size, shape and amount of crystalline phases. For each sample, five to
236	ten BSE images, cut perpendicularly to the rotational axis, were collected and analyzed on
237	different parts of the polished section to ensure that results were statistically significant.

#### 238 **4. Results**

### 239 *4.1 Mercury's superliquidus melt viscosity and modelling*

The liquidus temperature (T<sub>L</sub>) of our NVP composition was estimated using the alpha-MELTS software package [*Asimow, et al.*, 2004; *Smith and Asimow*, 2005] providing a value for T<sub>L</sub> = 1581 K. Experimental data are in good agreement with this estimate; indeed, the experiment performed at 1569 K contained only 2.1 area % of crystals indicating that the liquidus temperature is slightly higher. This is confirmed by the experiment at T=1600 K in which no crystals were detected.

Twenty-seven superliquidus experiments in the temperature range between 1600 and 1736 K were performed in order to determine the dependence of melt viscosity on temperature (**Table 2** and **Figure 2**). In these experiments, melt viscosity ranges from 4.0 Pa s and 16.3 Pa s. Each of the viscosity values presented in Table 2 is an average of 100 to 500 measurements, collected on timescales from 120 to 240 minutes, with torque values measured continuously. The majority of the experiments were performed at a shear rate of 5 s<sup>-1</sup>. We selected this value

252	because of the very low viscosity of the investigated composition. At lower shear rate the
253	instrument's torque limit is reached and measurements are not feasible. Experiments at 1736 K
254	(M3a-b), 1711 K (M7a-b) and 1691 K (M10a-b) were repeated twice and show a high
255	reproducibility (Table 2). The dependence of viscosity upon shear rate ( $\dot{\gamma}$ ) was investigated with
256	a series of 7 experiments performed at 1672 K with $\dot{\gamma}$ from 5 to 10 s <sup>-1</sup> (samples M16a-g in Table
257	2). Larger strain rates were not applied in order to prevent the melt to spill out from the crucible
258	due to its high fluidity. No measurable effect of shear rate on viscosity was detected at
259	superliquidus temperature (Table 2).
260	The obtained viscosity dataset was used to develop an empirical model of viscosity as a
261	function of temperature. The model is based on the Vogel-Fulcher-Tammann (VFT) equation
262	[ <i>Vogel</i> , 1921]:

263 
$$\log \eta (Pa s) = A + B/(T - T_0)$$
 (1)

where T is the temperature in Kelvin and A, B, and T<sub>0</sub> are fitting parameters, representing the 264 265 pre-exponential term, the pseudo-activation energy (related to the barrier of potential energy obstructing the structural rearrangement of the liquid), and the VFT temperature, respectively. 266 The VFT approach accounts for the non-Arrhenian temperature dependence of melt viscosity. 267 Data were fitted using a non-linear least-square regression providing the following parameters: A 268 = - 4.30 (Pa s), B = 6244.9 (K) and  $T_0 = 471.2$  (K). This relationship reproduces our 269 experimental data with a r<sup>2</sup> value of 0.99 (Figure 2). Note that the fitting parameters reported 270 above are only valid for high-temperature viscosity data. In fact, when comparing our VTF 271 parameters with those presented in Sehlke and Whittington [2015] for melts considered similar to 272 those erupted on Mercury, we observe a general agreement with terms A and B, but a 273

disagreement with the T<sub>0</sub> parameter. This reflects the inability of the above model to reproduce
data below the liquidus temperature.

The comparison of viscosity data presented here to those recently published by Sehlke 276 and Whittington [2015] shows a maximum difference of 5.0 Pa s log (not Cable 1 and Figure 277 3) mainly due to the different chemical composition of the silicate melt used in the experiments. 278 279 In this respect, an important feature of our composition is the high Na<sub>2</sub>O (8.85 wt.%) and low Al<sub>2</sub>O<sub>3</sub> (8.95 wt.%) contents, in agreement with MESSENGER data for the most evolved lavas of 280 the NVP [Peplowski et al., 2014, 2015]. These compositional characteristics are responsible for 281 the low viscosity of our silicate melt. Previous experimental data (e.g. Le Losq and Neuville, 282 2013; Vetere et al., 2014; Stabile et al., 2016) are in agreement with the lower melt viscosity 283 described here. 284

285

#### 286 *4.2 Mercury's subliquidus viscosity*

The viscous behaviour of the melt below  $T_L$  was investigated using 15 experiments in the temperature range of 1569–1502 K at three different shear rates: 0.1, 1.0 and 5.0 s<sup>-1</sup>. Prior to cooling the melt was kept at 1673 K for 2 h in order to erase possible crystal nuclei. The final dwell temperature was reached using a cooling ramp rate of 20 K/min. Experiments were run for up to 24,000 s. In this temperature range, crystals nucleated and grew, increasing from 2 to 28 area %, as temperature decreased (**Table 3** and **Figure 4a-b**).

The relative abundance of mineral phases, crystal sizes, and shapes do not vary significantly at different shear rates. Olivine (pure forsterite) is the liquidus phase in all experiments. Clinopyroxene appears at a temperature of 1520 K and remains stable in the system down to 1502 K. Clinopyroxene does not show significant compositional evolution, varying

297	from W039En61 at 1520 K to W037En62 at 1502 K, although a slight increase of sodium content is
298	observed (from 0.25 to 0.95 wt.%) with decreasing temperature (Table S2, supplementary
299	materials). Image analysis estimates of crystal fractions and those obtained by mass balance
300	calculations agree well and are reported in Table 3 and in Figure 5.
301	The compositional evolution of the melt during crystallization is shown in Figure 6 (see
302	also electronic supplementary material). SiO2 increases from 62.35 to 65.95 wt.% as temperature
303	decreases from liquidus (TL) to 1502 K. Similar trends are observed for Al2O3, TiO2, and K2O,
304	whereas CaO and MgO decrease. Na2O shows a more scattered behaviour, presumably due to the
305	combined effect of devolatilization of this element at high temperature and its incorporation into
306	clinopyroxene crystals.
307	Viscosity vs. time at constant temperature shows a typical S-shape curve, as reported in
308	Figure 7a-c-e. On the right panels of the figure, representative BSE images of experimental
309	samples are also shown for three selected samples (M34, M35 and M36; table 3) at 1510, 1520
310	and 1533 K at shear rate ( $\dot{\gamma}$ ) of 5.0 s <sup>-1</sup> ( <b>Figure 7b-d-f</b> ).
311	As shown in Table 3 and Figures 4 and 7, viscosity at $\dot{\gamma} = 0.1 \text{ s}^{-1}$ varies between 2.72–
312	4.01 Pa s [log ( $\eta \mathbf{a}$ ] for temperatures ranging from 1533 to 1502 K. As $\dot{\gamma}$ values increase to 1.0
313	and 5.0 s <sup>-1</sup> , viscosity [log ( $\eta \mathbf{I}$ ] varies between 1.93–3.04 Pa s (at 1545 K) and 2.25–3.36 Pa s (at
314	1502 K), i.e. viscosity decreases as shear rate increases. This points to a shear thinning behaviour
315	of the partly crystallized melt. From the curves displayed in Figure 7, the time for crystal
316	nucleation and growth can be evaluated at different temperatures [Vona et al., 2011]. Crystal
317	growth appears to be inversely correlated to the shear rate indicating that the higher the applied
318	shear rate, the lower the time needed for crystals to nucleate and grow. For example, at 1510 K
319	the time to reach a constant viscosity value (plateaux in Figure 7a) at $\dot{\gamma} = 0.1 \text{ s}^{-1}$ is slightly longer

than 4.0 hours. At the same temperature, viscosity reaches a constant value after about 2.0 hours at  $\dot{\gamma} = 5.0 \text{ s}^{-1}$ . As temperature increases to 1533 K, the time taken to reach a constant viscosity in the partly crystallized system is about 2.0 hours and 1.0 hour, for  $\dot{\gamma}$  of 0.1s<sup>-1</sup> and 5.0 s<sup>-1</sup>, respectively (Figure 7e).

324

#### 325 *4.3 Further rheological considerations*

During crystallization, the shear thinning behaviour becomes evident for all experiments, with viscosity decreasing as the shear rate increases (Figure 4). Noteworthy is the fact that the shear-thinning behaviour also arises at low crystal fractions ( $\Phi_c=0.05$ ) and increases at higher crystal contents (Figure 4a). Following the work from *Sehlke and Whittington* [2015], *Sehlke et al.*, [2014] as well as from *Vona et al.*, [2011 and 2013], we calculate the flow index using the linear regression coefficients derived from our experimental data set (**Table 3** and **Figure 8 a** and **b**; see also supporting information files for details).

Flow index values are relatively low compared to those estimated by *Sehlke et al.* [2014] 333 for Hawaiian basalts, as well as from Sehlke and Wittington [2015] for Mercury's analog 334 compositions (Fig. 8b). This is presumably due to the large aspect ratios of crystals in our study 335 (up to 14) that can strongly influence flow index values [Mader et al. 2013]. This highlights the 336 highly non-Newtonian behaviour of our partly crystallized analog composition. As an example, 337 at a temperature of 1533 K and crystal content of ca. 9.0 area % (sample M36 in Table 3) the 338 flow index (n) is 0.53. Under these conditions, an increase in shear rate from 0.1 to 5 s<sup>-1</sup> results in 339 a decrease in viscosity by a factor of 6. In a similar way, experiment M33 performed at 1502 K 340 (crystal content of 28 area % with n = 0.42) shows a decrease in viscosity by a factor of 10 (see 341 Table 3 and Figures 4 and Figure 8). A comparison between the rheological behaviour of our 342

343 composition with data from *Sehlke and Wittington* [2015] is provided in Figure S1

344 (supplementary materials), where the change in apparent viscosity with crystal fraction is shown.

Results indicate a good general agreement between literature data and our experimental results.

346 A slight deviation is observed at the lowest shear rate possibly due to differences in crystal

347 shapes and distributions.

Flow curves for Mercury lavas superimposed on the pahoehoe to `a`a transition threshold 348 diagram, derived from Sehlke et al. [2014], are presented in Figure 9. Mercury's lavas show 349 similar characteristics, but at slightly lower crystal fractions (0.05) compared to those studied by 350 Sehlke et al. [2014]. As the melt crosses the liquidus temperature (1581 K), pseudoplasticity 351 arises (flow index n < 0.7), as shown by the fact that the magma lies in the transition threshold 352 zone (TTZ) in Figure 9. It is not surprising that pseudoplastic behaviour is found at very low 353 crystal content. In fact, Ishibashi and Sato [2007] detected pseudoplastic behaviour in alkali 354 olivine basalt at  $\Phi$ as low as 0.05 (olivine, plagioclase, and spinel). In addition, Ishibashi [2009] 355 detected pseudoplastic behaviour in basalt from Mount Fuji between  $\Phi$  of 0.06 to 0.13 due to 356 suspended plagioclase crystals. The transition from pahoehoe to `a`a for the analog Mercurian 357 lava studied here begins at a temperature of  $\sim 1533 \pm 10$  K. 358

In modelling viscosity vs. crystal content, the relative viscosity (defined as  $\eta_r = \eta_{eff}/\eta_m$ , where  $\eta_{eff}$  is the effective viscosity of the suspension with a volume fraction of crystals, and  $\eta_m$  is the viscosity of the melt; see supplementary materials for details) is one of the most used parameter. **Figure 10** shows the variation of  $\eta_r$  as crystallinity ( $\Phi$ ) increases. Our experimental data can be described by the Einstein-Roscoe equation (see supplementary information) for relatively high  $\dot{\gamma}$ . Lowering  $\dot{\gamma}$  to the value 0.1 s<sup>-1</sup> results in higher  $\eta_r$  (reaching values up to ca. 15) matching results reported by *Vona et al.*, [2011] for crystallinity higher than 20 vol.%.

#### 366 **5. Discussion**

According to results presented by Byrne et al. [2013] on the lava flows forming the NVP, 367 some important constraints emerge: i) NVP lavas can flow over very long distances, on the order 368 of hundreds of kilometres; ii) the channels filled by the lavas appear quite variable in terms of 369 their width, spanning from a few hundreds meters to tens of kilometres; iii) the morphological 370 characteristics of lava flows, observed using high-resolution images of the planet surface, 371 372 indicate that lavas were emplaced as turbulent flows [Byrne et al., 2013]. These constraints must be taken into account when attempting to shed new light upon the mechanisms that might have 373 contributed to the emission and emplacement of lava flows forming the NVP. In the following 374 discussion, the new experimental data presented in this work are integrated with the above 375 constraints, in order to refine our understanding of the dynamics of NVP lava flows. Notably, in 376 conducting both superliquidus and subliquidus experiments, we attempt to constrain the 377 behaviour of lavas in terms of velocity and distance covered by the flows, as well as the possible 378 effusion rates. 379 The initial issue to consider is the slope of the terrains on which the lava was emplaced. 380

Although it would be preferable to use pre-eruption topographic data to estimate lava flow velocity, this is not possible at present and we must rely upon present day, post-NVP emplacement, topography [e.g. *Byrne et al.*, 2013]. Hereafter, we consider average slope values in between 0.1°–5°; these are of the same order of magnitude as those used in other studies of lava flows on the surface of Mercury [e.g., *Byrne et al.*, 2013; *Hurwitz et al.*, 2013].

As the emplacement of lavas on Mercury likely occurred in a turbulent regime [*Byrne et al.*, 2013] we can use the approach proposed by *Williams et al.* [2001], which is valid for high Reynolds numbers (*Re>>*2000), to infer the possible velocity of lava flows. This method allows for the estimation of lava flow velocity u (*m/s*), considering the ground slope  $\theta$  (°) and the lava friction coefficient  $\lambda$ :

391 
$$u = \sqrt{\frac{4gh\sin(\theta)}{\lambda}}$$
 (2)

392 where  $g(m/s^2)$  and h(m) are the acceleration due to gravity and the lava flow thickness,

respectively. The acceleration due to gravity on Mercury is  $g=3.61 \text{ m/s}^2$  (i.e. almost 1/3 of that of the Earth; *Mazarico et al.* 2014).  $\lambda \mathbf{n}$  can be calculated as follows:

395 
$$\lambda = \frac{1}{[0.79\ln(Re) - 1.64]^2}$$
 (3)

396 where *Re* is the Reynolds numbers defined as:

$$Re = \frac{2\rho uh}{\eta} \tag{4}$$

 $\rho$  is the bulk lava density (2450 kg/m<sup>3</sup> for the melt considered here, calculated following *Ochs and Lange* [1999]) and *η* is the lava viscosity (Pa s).

With regard to the thickness of the flows to be used in the above equations, the total 400 thickness of NVP lavas has been estimated to be ~ 0.7-1.8 km [Ostrach et al., 2015; Head et al., 401 2011; Klimczak et al., 2012; Byrne et al., 2013]. However, this thickness is likely to be the result 402 of the superimposition of different lava flows, whose individual thicknesses are presently 403 unknown. Some constraints can be derived from terrestrial analogous lavas. Among them, 404 Hawaiian lavas with rheological behaviours similar to the silicate melt considered here typically 405 show individual flow thicknesses ranging from 1 to 5 m [Griffiths, 2000]; furthermore, 406 komatiites, also commonly considered similar to Mercurian lavas [Weider et al., 2012], show 407 flows with a slightly greater thickness of about 10 m [Williams et al., 2001]. Accordingly, values 408 of *h* from 1 to 10 m are used in equations 2 and 4. 409

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The values of viscosity to be used in the above equations depend upon the eruptive 410 temperature. Possible eruption temperatures for Mercurian lavas are estimated to be around 1623 411 K [Charlier et al., 2013; Namur et al., 2016b]. As the estimated liquidus temperature for the melt 412 used in the experiments is similar (1600 K) to the suggested eruptive temperature, we started our 413 modelling considering 1623 K as a representative eruptive temperature of the lava that is initially 414 erupted. Changing the eruptive temperature to 1600 K does not affects the outcomes of the 415 model presented below. According to our experimental data and modelling (Figure 2), at 1623 K 416 the viscosity of the silicate melt is ca. 13 Pa s. Assuming constant effusion rates, calculated 417 velocities for lava thicknesses of 1, 5 and 10 m, considering varying ground slopes (from 0.1 to 418  $5^{\circ}$ ) are shown in **Figure 11**. As expected, results show that lava flows with larger thickness 419 consistently have larger velocities (see also Table S3 in the supporting materials). In addition, in 420 order to maintain the turbulent regime (*Re* larger than 2000) for lava emplacement (*Byrne et al.*, 421 2013), minimum velocities must be larger than ca. 1.2 m/s (on a slope of 0.3°), and ca. 0.7 m/s 422 (on a slope of  $0.1^{\circ}$ ) for lava thicknesses of 5 and 10 m, respectively. Note that for lava 423 thicknesses on the order of 1.0 m, the turbulent regime is only possible for slopes larger than 10°, 424 a value which appears unlikely on the basis of recent studies in this sector of the planet's surface 425 426 [e.g., Byrne et al., 2013; Hurwitz et al., 2013]. As the slope increases, the lava flow velocity can increase up to values of the order of 6.5 and 7.2 m/s, when flowing on a 5° ground slope, for 427 thicknesses of 5 and 10 m, respectively. 428 429 In order to incorporate the possible effects of heat loss during the lava flow in our model,

430 we performed thermal balance calculations using the FLOWGO model [*Harris and Rowland*,

431 2001]. In the model we consider that lavas can flow in channels with variable widths from 100,

432 1000 and 30000 m, as observed for the NVP on Mercury [*Byrne et al.*, 2013]. In addition, we

consider, as minimum and maximum velocities of the lava, the values obtained above for a lava 433 flow of thickness h=10 m flowing on a slope of  $0.1^{\circ}$  (u=0.7 m/s) and for a lava flow of the same 434 thickness flowing on a slope of  $5^{\circ}$  (u=7.2 m/s). Note that, according to the above discussion, 435 these end-member values of *u* include all possible values of velocity for a lava flowing down a 436 slope at an angle between 0.1 and  $5^{\circ}$ , with a thicknesses ranging from 5–10 m. A further 437 constraint to be considered in the model is the emissivity value of the lava. According to Harris 438 [2013] this parameter shows little variability, even among very different magmatic compositions. 439 In particular, it ranges from ca. 0.8 to ca. 0.9 from basalt to trachyte; accordingly, in the model, 440 we used an emissivity value of 0.85. The differences produced when this parameter is changed 441 are negligible. The FLOWGO model accounts for the formation of a crust developing on the 442 outer part of the lava acting as a thermal insulation boundary and limiting the heat loss. In 443 addition, turbulence cannot be directly included in FLOWGO. However, the large velocity 444 values used in the model can be considered as a proxy for turbulence implying that larger 445 velocities produce larger covered areas by the lava and, hence, larger heat losses. Parameters 446 used in the model are reported in Table S4 (supplementary materials); details about the 447 FLOWGO model can be found in Harris and Rowland [2001]. 448

Figure 12 shows the variation of heat loss (in K/km) of the lava as a function of effusion rate (in m<sup>3</sup>/s). The plot displays six curves corresponding to lava flows with the aboveconsidered velocities (0.7 and 7.2 m/s) flowing in channels with widths of 100, 1000 and 30000 m. The graph shows that the curves have a similar behaviour and tend to saturate towards constant values of heat loss at high effusion rates (of the order of 10<sup>4</sup>, 10<sup>5</sup> and 10<sup>7</sup> for channels having width of 100, 1000 and 30000 m, respectively). At lower effusion rates the heat loss dramatically increases (Figure 12). The curves corresponding to the lava flowing with a velocity of 0.7 m/s indicate that, at the same value of effusion rate, the heat loss is always lower in
comparison to the lava flowing with a velocity of 7.2 m/s. From Figure 12, it also emerges that
the channel width plays a major role in modulating heat loss. In fact, to maintain a similar
amount of heat loss, strongly increasing effusion rates are required as the width of the channel
increases.

As stressed above, a fundamental constraint to be satisfied is that a turbulent regime 461 likely characterized the emplacements of the lava for distances on the order of 100 km [Byrne et 462 al., 2013]. According to the above discussion, Reynolds numbers larger than 2000 are possible 463 considering the eruptive temperature T=1623 K (and relative value of viscosity,  $\eta$ =13 Pa s), and 464 lava velocities larger than 0.7 m/s, according to the slope values and lava thicknesses reported 465 above (see supporting information, Table S3). Keeping lava velocity constant in the limits 466 imposed above (i.e. 0.7 and 7.2 m/s), we can use the results shown in Figure 12 to evaluate how 467 heat loss impacts lava rheology and the consequent dynamic regime of lava emplacement. As the 468 lava cools down during its flow, its viscosity is expected to increase leading to a decrease of the 469 Reynolds number and resulting, eventually, in the suppression of the turbulent regime. 470 According to our experimental results and modelling (Figure 2), the viscosity of the analogous 471 melt studied here shows minimal variations (13–20 Pa s) in the temperature range 1623–1581 K 472 (i.e. from eruptive to liquidus temperature; Figure 3). This implies that the Reynolds number 473 remains relatively constant during this temperature drop ( $\Delta T=42$  K, i.e. 1623–1581 K) allowing 474 the turbulent regime to remain almost unchanged. Considering that the lava has to cover a 475 distance on the order of 100 km in turbulent regime [Byrne et al., 2013], this would correspond 476 to a heat loss of ca. 0.4 K/km. According to the model shown in Figure 12, a heat loss of 0.4 477 K/km can be obtained at different effusion rates depending on the width of the channel in which 478

the lava flows. In particular, effusion rates larger than ca.  $10^4$ ,  $10^5$  and  $10^6$  m<sup>3</sup>/s are required for 479 channel widths of 100, 1000 and 30000m, respectively. It is noteworthy that these estimates are 480 in concurrence with independent results given by Keszthelvi and Self [1998] for the emplacement 481 of long (on the order of 100 km) basaltic lava flows. In particular, Keszthelvi and Self [1998] 482 report that in order to cover these large distances, a heat loss equal to or lower than 0.5 K/km is 483 necessary. Furthermore, these authors suggest that effusion rates larger than  $10^3$ – $10^4$  m<sup>3</sup>/s, and 484 velocities on the order of 4-12 m/s are also required. These values are comparable to those 485 arising from our model. From this discussion, it is therefore clear that the Mercurian analog melt 486 studied here is able to flow over the long distances observed for the NVP in turbulent regime, 487 under the effusion rates given above. Significantly, the required effusion rates (and relative 488 single lava flow thickness) are comparable to those estimated for some of Earth's basaltic 489 eruptions forming the so-called LIPs [e.g. Bryan et al., 2010; Head et al. 2011]. Consequently, 490 the mechanism of eruption and emplacement of terrestrial flood basalts can be considered, as a 491 first approximation, as a plausible geologic analogue for the Mercury's NVP lavas in agreement 492 with the results reported by Vander Kaaden and McCubbin [2016]. 493

A further issue that might deserve consideration is the possible effect of cooling of the 494 495 lavas due to the temperature difference between diurnal and nocturnal times on Mercury. During a Mercurian day (corresponding to 59 terrestrial days), surface temperature can reach 725 K. At 496 night, the surface temperature drops to about 90 K [Strom, 1997]. In these conditions, the high 497 498 temperature of the planet may limit heat dissipation during daytime eruptions and, consequently, enhance the emplacement of lava flows. Conversely, during the night, the temperature drop 499 could act as a limiting factor for the distance the lava is able to flow. However, in modelling lava 500 501 flows, the formation of a crust developing on the outer part of the lava must be considered. This

effect is considered in the FLOWGO models presented above. The formation of this crust acts as 502 thermal insulation for the lava and, therefore, the effect of temperature difference between 503 Mercurian day and night can be considered negligible. 504 According to the above discussion, it is apparent that NVP lavas were able to cover large 505 distances without undergoing strong degrees of crystallization. Accordingly, most of their 506 507 solidification history would have occurred when they stopped flowing and emplaced, defining the present morphology of Mercury's NVP. The causes why lavas stopped flowing on the 508 surface of the planet can be variable. One possibility might be that the topography of the planet 509 played a key role, for example because of the presence of low topographies due to pre-existing 510 impact craters that allowed the lavas to stagnate, lose heat and solidify. 511 Further evidence that solidification of the lavas must have occurred mostly after 512 emplacement is provided by results from crystallization experiments. Lavas can strongly reduce 513 their ability to flow when approaching the so-called "maximum packing fraction". The 514 maximum packing fraction strongly depends on the aspect ratio of crystals [Mader et al., 2013]. 515 Our experiments indicate that crystal aspect ratios range, on average, between 2 and 14. 516 According to *Mader et al.* [2013] this corresponds to maximum packing fractions from ca. 52% 517 518 to 28%. As here we are evaluating the least favourable conditions under which the lavas can flow, we considered the lowest maximum packing fraction (i.e. 28%) as the reference crystal 519 fraction that would strongly reduce (or stop) the ability of the lava to flow. Our experiments 520 521 (Figure 4a-b) indicate that temperature needs to drop to 1502 K to reach a crystal content of ca. 28%. The question is whether the lava in these conditions can preserve the turbulent flow 522 regime, as required by morphological constraints provided by [Byrne et al., 2013]. Our 523 524 experimental results presented in Figure 4 and Table 3 show that the rate of increase in viscosity

due to crystallization is different depending on the applied shear rate, due to a shear thinning
behaviour of the lava. As an example, at T=1520 K (corresponding to a crystal content of ca.
10%; Table 3) viscosity changes from 144 Pa s to 1259 Pa s, as the shear rate decreases from 5 to
0.1 s<sup>-1</sup>. A similar behavior is observed for lower temperatures, i.e. larger crystal contents, up to
the threshold limit of 28% crystal content (Figure 4).

530 The Reynolds number (equation 4) can be used to assess whether the lava can emplace in turbulent flow conditions, i.e. Re larger than 2000. In the following calculations, we consider a 531 lava flow with thickness of 10 m flowing at velocities from 0.7 to 7.2 m/s (see above) and having 532 viscosities determined by the amount of crystals formed upon solidification (Figure 4). Note that 533 for lavas with thicknesses lower than ca. 10 m, considering the viscosities for crystal-bearing 534 lavas with a crystallinity equal to or larger than 5% (Figure 4 and Table 3) and for any of the 535 above velocities, *Re* is always lower than 2000; in these conditions the laminar regime prevails. 536 For lava thicknesses on the order of 10 m, i.e. maximum thickness considered here, the laminar 537 fluid dynamic regime governs most of the behavior of the flowing lava. However, according to 538 our experimental results, there are cases in which the turbulent regime might still persist, even 539 for partially crystallized systems. In particular, Re can attain values above 2000 if viscosity is 540 541 lower than 144 Pa s and flow velocity is larger than 6 m/s. From Figure 4, these viscosity values correspond to partially crystallized systems with crystal contents up to ca. 10% for shear rates of 542 5 s<sup>-1</sup>. Lower shear rates shift *Re* towards values lower than 2000, driving the system towards 543 544 laminar conditions. These considerations highlight that, whilst for some narrow combinations of parameters (i.e. viscosity, velocity, lava thickness), the lava could potentially flow in turbulent 545 conditions below the liquidus temperature (TL), in most cases it behaves as a laminar system. 546 547 However, this is contrary to morphological features observed from satellite images and

548 considered to reflect a turbulent emplacement of the lava [Byrne et al., 2013]. These

549 considerations corroborate the idea proposed above, that crystallization of lavas is a process that 550 most likely occurred after emplacement.

551 6. Conclusions

The new experiments and modelling presented in this work allow us to shed new light on the mechanisms that determined the emplacement of what is believed to be one the largest volcanic deposits in the Solar System [*Byrne et al.*, 2016]; the northern volcanic province on planet Mercury. The high Na<sub>2</sub>O content (~8.8 wt.%) of the experimental starting material plays an important role in reducing lava viscosity, as confirmed by concentric cylinder high temperature viscosity measurements.

The viscosity of our Mercury analog silicate melt measured at superliquidus temperature conditions slightly increases from 4 to 16 Pa s in the temperature range 1736–1600 K. In the temperature range 1569–1502 K (subliquidus), viscosity increases due to the combined effect of progressive crystallization (from 2 to 28 area %) and chemical evolution of the melt. Here, a shear-thinning behavior was observed when varying strain rates from 0.1 and 5 s<sup>-1</sup>. Lava viscosity decreases by ca. 1 log unit as shear rate varies from 0.1 and 5 s<sup>-1</sup>.

These new viscosity measurements were used to model the behaviour of Mercurian lavas during emplacement. Merging experimental data and numerical modelling leads to the conclusion that the emplacement of lavas in turbulent conditions, as claimed by previous works [e.g. *Byrne et al.*, 2013], defines a geologic scenario in which lavas might have travelled long distances without undergoing strong degrees of crystallization. Therefore, it is possible to infer that solidification (crystallization) of the lavas mostly occurred after emplacement on the surface

- of the planet. Effusion rates were estimated to be in the order of  $10^4$ – $10^7$  m<sup>3</sup>/s, comparable to
- 571 those estimated for some Earth's basaltic eruptions forming the LIPs.
- 572

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# **Table captions**

832	Table 1. Electron microprobe analyses of the starting material compared to literature data. The
833	starting composition represents an average of 50 measurements on the synthetic starting glass
834	material. SW-2015 refers to literature data presented in Sehlke and Whittington [2015]. Std
835	refers to the standard deviations. The NBO/T values (non-bridging oxygens (NBO) per
836	tetrahedrally coordinated cation (T) [Mysen and Richet, 2005]). NBO/T values for Sehlke and
837	Whittington [2015] chemical compositions refer to calculation results considering (a) ferrous
838	only, (b) ferrous and ferric, and (c) ferric only.
839	
840	Table 2. Experimental conditions and results of viscosity measurements $(\eta)$ using the concentric
841	cylinder apparatus. $\dot{\gamma}$ (in s <sup>-1</sup> ) refers to the applied shear stress; std refers to standard deviation of
842	viscosity measurements.
843	
844	Table 3: Experimental conditions and results of viscosity measurements $(\eta)$ during
845	crystallization experiments at different shear rate ( $\dot{\gamma}$ ). Rheological parameters [flow index (n) and
846	consistency $(k)$ ] and crystal contents (both area% and vol.%) are also reported.
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#### 859 Figure Captions

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Figure 1. Back scattered electron images of representative experimental products. a) Experiment at temperature T=1502 K and shear rate  $\dot{\gamma}$ =5.0 s<sup>-1</sup>; b) experiment at same temperature as (a) with a shear rate  $\dot{\gamma}$ =0.1 s<sup>-1</sup>. The two experiments show comparable crystal contents. Labelled crystal phases are olivine (Ol), clinopyroxene (Cpx) and glass.

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Figure 2. Viscosity data for melts at superliquidus conditions. The continuous line represents the
 predictive model given in Eq. (1). Black full triangles are repeated measurements; green squares
 represent experiments performed at different shear rates.

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Figure 3. Variation of viscosity as a function of temperature showing the comparison between

data presented in this work and literature data using possible Mercury compositions. Icp-HCT

and E\_B refer to basaltic komatiites and Enstatite Basalt, respectively, as reported in *Sehlke and Whittington* [2015]

874

Figure 4. a) Variation of viscosity  $[log(\eta)]$  as a function of crystal content ( $\Phi$  area %) at different shear rates ( $\dot{\gamma}$ ); b) variation of viscosity  $[log(\eta)]$  as a function of temperature for crystallization experiments at different shear rates ( $\dot{\gamma}$ ). Measurement errors in Figure 4b are

comparable or lower than symbol size.

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Figure 5. Comparison between crystal content obtained by image analysis (area %) and mass
balance calculations (vol.%). Errors in area % are on the order of 10 % relative.

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Figure 6. Change of the residual glass composition (in wt.% of element concentrations) upon
crystallization of the silicate melts. Triangle, circle and square indicate experiments performed at
shear rate of 5.0, 1.0 and 0.1 s<sup>-1</sup>, respectively.

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Figure 7. a-c-e): Variation in viscosity as a function of time and temperature at different shear
rates; b-d-f): corresponding BSE images and phases after reaching the saturation threshold (flat

part of profiles in the plots on the left) for experiments performed at  $\dot{\gamma} = 5 \text{ s}^{-1}$ . Dashed-lined 889 arrows indicate flow direction; Ol and Cpx refer to olivine and clinopyroxene, respectively. 890 891 892 Figure 8. a) Linear regressions of  $\ln(\sigma)$  (shear stress) against  $\ln(\dot{\gamma})$  (shear rate) for the studied composition. Linear regressions for each dataset are also shown. Note the drastic change in slope 893 when passing from the super-liquidus experiment (1600 K) to the crystal-bearing experiments 894 due to the onset of crystallization; b) flow index derived from the linear regression of data in (a). 895 Estimated rheological parameters (*n* and *k*) are given in Table 3. 896 897 Figure 9. Flow curves for the analog Mercury lavas from this work superimposed on the 898 pahoehoe to `a`a transition diagram from Sehlke et al. [2014]. Coloured lines from right to left 899 refer to data from Sehlke et al. [2014] for Hawaiian lavas; dots refer to data from this study. Lava 900 on Mercury start the transition from pahoehoe to `a`a at a temperature of  $\sim 1533 \pm 10$  K. The 901 question mark, as reported in Sehlke and Whittington [2015], emphasizes the approximate 902 location of the end of the transition threshold zone TTZ. 903 904 **Figure 10.** Relationship between relative viscosity  $(\eta_r)$  and crystal fraction 905 906 melt in this study (coloured symbols) and literature data. Curves correspond to different models: 907 ER, Einstein-Roscoe model [Einstein, 1906; Roscoe, 1952]; KD, Krieger-Dougherty model 908 [Krieger and Dougherty, 1959]; Sato model [Sato, 2005]; Mader model [Mader et al., 2013]; 909 Vona model [Vona et al., 2011] see also supplementary information for details). 910 911 Figure 11. Variation of lava flow velocity as a function of topographic slope for three lava flows 912 with thickness of 1 m, 5 m and 10 m (Eq. 2-4). 913 914 Figure 12. Variation of heat loss (in K/km) as a function of effusion rates for a lava flowing in 915 100, 1000 and 30000 m width channels, with velocities of 0.7 and 7.2 m/s respectively. Details 916 917 are provided in the main body of text.

## Tables

# Table 1

This work		SW-2015				
	wt%	std	Enstatite bas	NVP	NVP-Na	IcP-HCT
SiO <sub>2</sub>	61.48	0.36	55.06	57.10	55.02	53.30
TiO <sub>2</sub>	0.36	0.01	0.18	0.96	0.89	0.89
$Al_2O_3$	8.95	0.11	13.07	15.27	14.88	12.31
FeO	-	-	0.29	3.61	2.88	3.31
Mn0	-	-	0.14	0.25	0.25	0.23
MgO	14.12	0.21	19.72	16.53	13.59	22.53
Ca0	6.81	0.09	12.37	4.95	4.29	6.29
Na <sub>2</sub> O	8.85	0.21	0.04	0.29	6.25	0.16
K <sub>2</sub> O	0.21	0.02	0.08	0.31	0.22	0.19
Tot	100.78		100.95	99.27	98.27	99.21
NBO/T	0.89		1.00ª	0.64 a	<b>0.68</b> a	1.05 <sup>a</sup>
-			_	0.50 <sup>b</sup>	0.55 <sup>b</sup>	0.86 <sup>b</sup>
			_	0.41 c	0.53 <sup>c</sup>	0.79 °

Table	2 2
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#	Т (К)	η (Pa s)	std	Ý	
M29	1600	16.3	0.27	5	
M28	1610	14.9	0.58	5	
M32	1618	12.6	0.43	5	
M27	1621	14.1	0.41	5	
M26	1630	12.9	0.35	5	
M31	1639	10.3	0.06	5	
M15	1652	9.7	0.05	5	
M14	1662	8.7	0.07	5	
M13	1671	7.9	0.04	5	
M12	1681	7.2	0.05	5	
M17	1687	6.7	0.11	5	
M10a	1691	6.8	0.05	5	
M10b	1691	6.7	0.08	5	
M9	1701	6.0	0.06	5	
M7a	1711	5.5	0.05	5	
M7b	1711	5.7	0.03	5	
M 6	1720	5.0	0.03	5	
M 5	1730	4.6	0.04	5	
M 3a	1736	4.3	0.03	5	
M 3b	1736	4.0	0.04	5	
M 16a	1642	10.7	0.17	5	
M 16b	1642	11.2	0.15	5	
M 16c	1642	11.2	0.26	6	
M 16d	1642	11.3	0.23	7	
M16e	1642	11.2	0.25	8	
M16f	1642	11.4	0.21	9	
M16g	1642	11.2	0.16	10	

Table 3

#	Temperature	Log η Pa s	Log η Pa s	Log η Pa s	Flow index	K	Olivine	Pyroxene	Crystallinity	Crystallinity mass balance
	К	γ̇=5 s⁻¹	ý=1 s⁻¹	γ̇=0.1 s⁻¹	n	Pa s	area %	area %	area %	vol %
M33	1502	3.04	3.36	4.01	0.4201	1172± 175	$17.6 \pm 3.0$	9.9 ± 1.3	$27.5 \pm 3.0$	23.7
M34	1510	2.67	3.18	3.59	0.4551	593± 53	11.5 ± 1.9	9.9 ± 0.8	21.4 ± 2.2	16.3
M35	1520	2.16	2.69	3.10	0.4568	207±30	9.9 ± 0.2	$1.7 \pm 0.2$	11.8 ± 1.2	10.1
M36	1533	1.99	2.47	2.72	0.5272	131±22	8.8 ± 0.9	-	8.8 ± 0.9	8.2
M37	1545	1.93	2.25		0.5333	61± 5	$5.0 \pm 0.7$	-	$5.0 \pm 0.7$	7.3
M38	1569	1.63			-	-	$2.1 \pm 0.3$	-	2.1 ± 0.3	3.9

Figure 1



## Figure 2







Figure 4







## Figure 6

















Figure 11



Figure 12





