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1           **The Effects of Digital Elevation Model Resolution on the PyFLOWGO**  
2           **Thermorheological Lava Flow Model**

3  
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13  
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20  **Highlights:**

- 21       • We present a topographic sensitivity analysis for the PyFLOWGO lava flow model.  
22       • Examined DEM resolutions were based on common Earth and planetary datasets.  
23       • A model step size equal to or smaller than the DEM resolution should be used.  
24       • DEM resolution has a quantifiable impact on the modeled flow length (up to 35%).  
25

26 **Abstract (150/150 word limit)**

27 Topography is a fundamental factor influencing the emplacement of lava flows. We assess the  
28 impact of topographic resolution on the thermorheological PyFLOWGO model, specifically with  
29 digital elevation model (DEM) resolutions commonly available for Earth and other planetary  
30 bodies where flow modeling is relevant. We examined PyFLOWGO output parameters (e.g.,  
31 channel length, core temperature, and flow velocity) to assess model sensitivity to topography in  
32 order to document model uncertainties and optimize future application. This study uses rheologic  
33 and topographic data from the Tolbachik, Russia 2012-2013 eruption as model constraints. Using  
34 moderate to lower resolution topographic data overestimates the channelized flow length by up to  
35 35% due to differences in the distribution of slopes topographic variability at different resolutions  
36 and resulting effects on the modeled lava velocity down channel. Determining the impact of  
37 topography on thermorheological lava flow models such as PyFLOWGO is important to correctly  
38 interpret the results for channelized flows.

39

40

41 **Keywords:** FLOWGO, Sensitivity analysis, Planetary science, Volcanology, Volcanic eruption,  
42 Tolbachik, Topography, DEM

43

44 **1 Introduction**

45 The emplacement of a lava flow is controlled by a number of factors, such as the rheology  
46 (i.e., lava viscosity, which depends on eruption temperature, composition, and degree of  
47 crystallization), the effusion rate, external environmental conditions (i.e., gravity, atmospheric  
48 density, and temperature), and importantly, the topography over which it flows. Modeling lava  
49 flow emplacement attempts to simplify these conditions to address specific volcanological  
50 questions. For lava flows on Earth, modeling is commonly performed either once the flow is  
51 emplaced in order to determine the rheological parameters or eruptive conditions (e.g., Harris &  
52 Rowland, 2001; Ramsey et al., 2019; Rowland et al., 2005) or in a prognostic manner for hazard  
53 response and assessment (e.g., Chevrel et al., 2022; Favalli et al., 2005, 2012; Harris & Rowland,  
54 2015). For investigations of inactive/older flows on Earth and other planetary bodies, the goal is  
55 typically to determine the eruption conditions and lava rheology necessary to produce the observed

56 flow length or area (e.g., Flynn et al., 2022; Ganci et al., 2020; Peters et al., 2021). No matter how  
 57 a lava flow model is applied, or the level of complexity included, topography is an essential dataset.  
 58 Therefore, understanding how the spatial resolution of a topographic dataset affects model results  
 59 is critical.

60 For modeling recent terrestrial flows, it is common to have access to digital elevation  
 61 models (DEM) of the pre-flow surface ranging from moderate (~10-50 m/pixel) to high ( $\leq 10$   
 62 m/pixel) resolution that are derived from photogrammetry, LIDAR, or InSAR datasets (Table 1)  
 63 (e.g., Chevrel et al., 2021; Kubanek et al., 2015; Richardson & Karlstrom, 2019). However, pre-  
 64 flow topography is not available for investigations of older lava flows. Some planetary bodies may  
 65 have limited coverage of high to moderate resolution DEMs of a study region, whereas others are  
 66 limited to much lower resolution DEMs (Table 1). The resolution of planetary DEMs and lack of  
 67 pre-flow datasets have resulted in many studies using a constant slope to perform lava flow  
 68 modeling (e.g., Baloga & Glaze, 2008; Hiesinger et al., 2007; Peters et al., 2021). However, several  
 69 previous studies concluded that variations in local topography have a measurable impact on model  
 70 results of a lava flow's emplacement (Glaze & Baloga, 2007; Russo et al., 2022).

71 **Table 1:** DEM datasets commonly available for Earth, Mars, and Venus.

<b>Planetary Body</b>	<b>Instrument</b>	<b>Spatial Resolution</b>	<b>Global Surface Coverage (%)</b>	<b>References</b>
<i>Earth</i>	ASTER <sup>a</sup>	30 m/pixel	100%	(Hirano et al., 2003)
	SRTM <sup>b</sup>	30 m/pixel	100%	(Farr et al., 2007)
	InSAR	12 m/pixel	~97%	(Rizzoli et al., 2017)
	ArcticDEM	2 m/pixel	~23%	(Takaku et al., 2020)
<i>Mars</i>	MOLA <sup>c</sup>	463 m/pixel	100%	(Zuber et al., 1992)
	MOLA/HRSC <sup>d</sup>	200 m/pixel	100%	(Ferguson et al., 2018)
	HRSC	50 m/pixel	~44%	(Jaumann et al., 2007)
	CTX <sup>e</sup>	~20 m/pixel	~17%	(Breton et al., 2019)
	HiRISE <sup>f</sup>	~1-2 m/pixel	~0.4%	(Sutton et al., 2022)
<i>Venus</i>	Altimeter	10-20 km/pixel	98%	(Ford & Pettengill, 1992)
	SAR Stereo	1-2 km/pixel	20%	(Herrick et al., 2012)

72 <sup>a</sup> Advanced Spaceborne Thermal Emission and Reflection radiometer

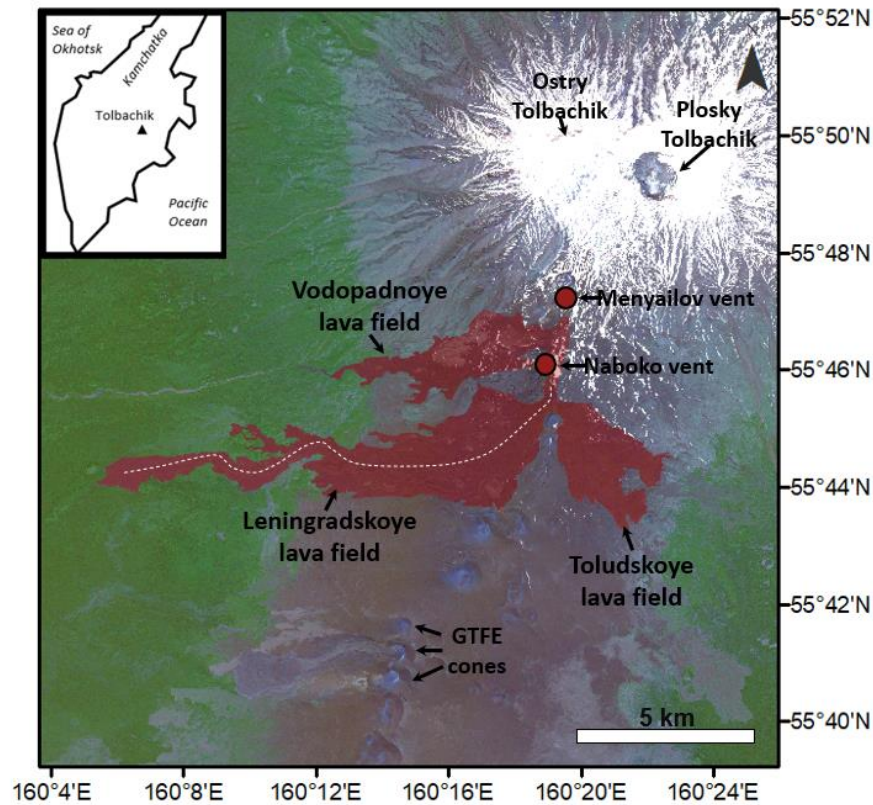
- 73 <sup>b</sup> Shuttle Radar Topography Mission  
74 <sup>c</sup> Mars Orbiting Laser Altimeter  
75 <sup>e</sup> Context Camera  
76 <sup>f</sup> High Resolution Imaging Experiment

77

78         There are numerous lava flow emplacement models, and many have associated sensitivity  
79 analyses to topography, including DOWNFLOW (Favalli et al., 2011), MAGFLOW (Bilotta et al.,  
80 2019), MULTIFLOW (Richardson & Karlstrom, 2019), and Q-LAVHA (Mossoux et al., 2016).  
81 However, this has not been done for FLOWGO/PyFLOWGO (Chevrel et al., 2018; Harris &  
82 Rowland, 2001). The focus of this study, therefore, is to investigate the impact of topographic data  
83 resolution on the FLOWGO thermorheological model's output with specific focus on commonly  
84 available DEM resolutions for Earth and other planetary bodies. We use a terrestrial study case  
85 that was presented in Ramsey et al. (2019): the 2012-2013 Tolbachik eruption (Figure 1). It was  
86 chosen because Ramsey et al. (2019) already performed the FLOWGO modeling and provided the  
87 input rheological parameters that best fit the lava flow.

### 88 **1.1 The 2012-2013 Tolbachik Eruption**

89         The Tolbachik volcanic complex is one of three active volcanoes (Klyuchevskoy,  
90 Bezymianny, and Tolbachik) that make up the Klyuchevskoy Volcano Group (KVG) in  
91 Kamchatka (Russia). Tolbachik is comprised of two stratovolcanoes, Ostry and Plosky Tolbachik,  
92 with multiple cones and lava flows to the south (Figure 1).



93

94 **Figure 1.** The Tolbachik volcanic complex comprised of Ostry and Plosky Tolbachik. To the immediate south of the  
95 volcanic complex, are the three flow fields (shown in red) and two vents (red circles with black outline) formed during  
96 the 2013 Tolbachik Fissure Eruption (TFE-50). Further south are three cones formed during the Great Tolbachik  
97 Fissure Eruption (GTFE). The white dashed line indicates the path of the central channel used for the PyFLOWGO  
98 model. Base image from Planets RapidEye Sensor (~5 m/pixel, visible), acquired on 18 June 2014.

99 Historically, the complex has produced eruptions throughout the Holocene and more recent  
100 times (Churikova et al., 2015). The two most notable eruptions from the complex are the 1975-  
101 1976 Great Tolbachik Fissure Eruption (GTFE) and the 2012-2013 Tolbachik Fissure Eruption  
102 (TFE-50), named after the 50<sup>th</sup> anniversary of the Institute of Volcanology and Seismology. During  
103 the GTFE new monogenetic cones and several associated lava flows were emplaced along the  
104 southern slope of Plosky Tolbachik (Figure 1) (Fedotov et al., 1984). However, most of the  
105 deposited material is pyroclastic, infilling the area around the cones. The smaller TFE-50 eruption  
106 produced mainly lava flows originating from vents that formed monogenetic cones closer to the  
107 main volcanic edifice and along the same path that produced the GTFE (Churikova et al., 2015).

108 During the TFE-50 eruption, which began on 27 November 2012, both Hawaiian and  
109 Strombolian eruption styles occurred, producing basaltic trachyandesite lava flows (Belousov &

110 Belousova, 2017). The average effusion rate, measured with aerial photogrammetric surveys, early  
111 in the eruption was  $\sim 440 \text{ m}^3/\text{s}$ , which gradually waned until the eruption ended on 5 September  
112 2013 (Belousov & Belousova, 2017; Ramsey et al., 2019). At the end of the eruption three  
113 prominent flow fields had formed, Vodopadnoye, Leningradskoye, and Toludskoye (Figure 1).  
114 The Vodopadnoye flow field (area  $6.17 \text{ km}^2$ , volume  $0.04 \text{ km}^3$ ) originated from the Menyailov  
115 vent and is comprised mostly of ‘a‘ā flows (Belousov et al., 2015; Belousov & Belousova, 2017).  
116 Both the Leningradskoye (area  $22.44 \text{ km}^2$ , volume  $0.35\text{-}0.39 \text{ km}^3$ ) and Toludskoye (area  $8.4 \text{ km}^2$ ,  
117 volume  $0.1 \text{ km}^3$ ) flow fields originated from the Naboko vent (Belousov et al., 2015). The  
118 Leningradskoye flow field contains mostly ‘a‘ā and the Toludskoye contains ‘a‘ā, pāhoehoe and  
119 various transitional lava types (Belousov et al., 2015; Belousov & Belousova, 2017). The total  
120 volume of lava produced from the TFE-50 eruption makes it one of the largest in history and one  
121 of the most thermally radiant eruptions in the modern satellite era (Ramsey et al., 2019). This study  
122 focuses specifically on the large, channelized lava flow that formed early in the eruption in the  
123 Leningradskoye flow field.

## 124 **2. FLOWGO model**

125 Lava flow modeling is a powerful tool for hazard mitigation and interpreting emplacement  
126 processes as well as understanding the evolution of thermorheological properties of flowing lava.  
127 One of the most extensive and widely utilized lava flow models is FLOWGO/PyFLOWGO (Table  
128 2). The original FLOWGO model was recoded and updated in the Python scripting language to  
129 PyFLOWGO, which allows for improved iteration and customization (Chevrel et al., 2018). For  
130 all modeling discussed throughout the rest of the manuscript, PyFLOWGO is used. As of 2023,  
131 twenty-two peer-reviewed publications have used FLOWGO or PyFLOWGO to investigate both  
132 active and inactive lava flows on Earth at a variety of locations. It has also been utilized to  
133 investigate volcanism on four other planetary bodies (Table 2).

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140 **Table 2.** Examples of studies employing FLOWGO/PyFLOWGO

Study Locations	References
<i>Earth</i>	
Cameroon	(Wantim et al., 2013)
D.R. Congo	(Mossoux et al., 2016)
Galapagos	(Rowland et al., 2003)
Hawai'i	(Chevrel et al., 2018; Harris et al., 2022; Harris & Rowland, 2015, 2001; Mossoux et al., 2016; Riker et al., 2009; Robert et al., 2014; Rowland et al., 2005; Thompson & Ramsey, 2021)
Italy	(Harris et al., 2007, 2011; Wright et al., 2008)
Kamchatka	(Ramsey et al., 2019)
La Reunion	(Chevrel et al., 2018, 2022; Harris et al., 2016, 2019; Peltier et al., 2022; Rhéty et al., 2017)
<i>Planetary</i>	
Mars	(Flynn et al., 2022; Rowland et al., 2004)
Mercury	(Vetere et al., 2017)
Moon	(Lev et al., 2021)
Venus	(Flynn et al., 2023)

141  
142 FLOWGO, originally described in Harris and Rowland (2001) is a one-dimensional,  
143 thermorheological model applicable to open channelized flows. It calculates the thermal budget  
144 and resulting lava properties (i.e., core and surface temperature, and crystallinity) of a lava volume  
145 down channel at each pre-defined interval step. At each step, the resulting depth-averaged viscosity  
146 and velocity are obtained. The model is only applicable for cooling-limited flows and to the well-  
147 established channel region, and not the dispersed flow region that extends downslope beyond the  
148 channel. For flows on Earth, the dispersed zone may be a region up to ~3 km beyond the end of  
149 the channel (Dietterich & Cashman, 2014; Harris et al., 2022; Lipman & Banks, 1987). For  
150 channelized flows on Mars, the dispersed zone may be up to ~49 km long (Flynn et al., 2022).

151 Thermal budgeting includes the radiative ( $Q_{\text{rad}}$ ) heat loss from the flow surface, convective  
152 ( $Q_{\text{conv}}$ ) heat loss to the atmosphere, conductive ( $Q_{\text{cond}}$ ) heat loss to the channel walls and floor, and  
153 heat loss due to vaporization of precipitation ( $Q_{\text{rain}}$ ) (where applicable). The heat gains are due to  
154 latent heat of crystallization ( $Q_{\text{cryst}}$ ) and viscous dissipation ( $Q_{\text{visc}}$ ) within the flowing lava. Under  
155 typical terrestrial conditions, heat loss from the modeled lava is dominated by radiative heat flux,  
156 which is a function of the temperature and effective emissivity of the lava's radiating surface  
157 (molten and crusted fractions) (Ramsey et al., 2019; Rowland et al., 2004).



158 PyFLOWGO makes five assumptions that define the model and where it is applicable. (1)  
159 The lava flow velocity is defined by the effusion rate, which is held constant downflow, in a  
160 channel with a defined depth and width (either estimated for an active flow or measured directly  
161 on an older flow) on a given slope. (2) The lava flow must be confined to an open channel with no  
162 continuous roofing or tube formation. (3) The vertical thermal structure of the lava is divided into  
163 three layers: a cooler basal crust, a homogeneous high-temperature molten core, and a radiating  
164 upper surface. (4) The model simulates the propagation of a channel-confined lava unhindered by  
165 the flow front or levee formation. (5) The flow is cooling- and not supply-limited.

166 As a cooling-limited flow slows down because of heat loss and increasing viscosity, it  
167 spreads laterally, commonly marking the end of the central channel, and will stop after it has cooled  
168 to an extent that its rheological behavior impedes forward motion (Chevrel et al., 2018). In  
169 practice, PyFLOWGO has one of three stopping conditions: (1) the modeled velocity reaches zero;  
170 (2) the modeled temperature of the flow core reaches the point of solidus; (3) the modeled yield  
171 strength of the flow core increases such that the flow is unable to advance. As the modeled velocity  
172 reaches zero the simulated channel width will increase substantially.

173 The main goal of this work is to investigate the effects of DEM resolution on  
174 PyFLOWGO's results. The rheologic input parameters are the same as provided by Ramsey et al.  
175 (2019) for the TFE-50 Leningradskoye channelized lava flow (Table 3), with the exception of the  
176 lava's thermal emissivity module. As shown by Ramsey et al. (2019) and later by Thompson and  
177 Ramsey (2021), PyFLOWGO is also sensitive to the assumed emissivity value chosen, with lava  
178 flow runout distance varying by ~6%. As part of this work, we also explore the effect of the three  
179 different emissivity modules (where emissivity is considered constant, bimodal, and a function of  
180 surface temperature) within PyFLOWGO and their effectiveness to reproduce the observed length  
181 of a lava channel for the TFE-50 eruption. Between the three emissivity module options there is a  
182 700 m (~6%) difference in modeled channel length. Full details of this investigation can be found  
183 in the Supplemental Material. For the DEM analysis, we choose to use the variable emissivity  
184 option because, similar to Thompson and Ramsey (2021), it appears to most accurately reproduce  
185 a flow's emplacement compared to the other emissivity options in PyFLOWGO.

### 186 **3 Methods: Topographic Study**

187           Following the argument presented in Bilotta et al. (2019) for the MAGFLOW model, the  
188 resolution of a DEM has both a direct and a secondary impact on modeling. The direct impact is  
189 the resolution of the DEM, and the secondary impact is the model step size (i.e., how frequently  
190 the model accounts for changes in topography). Both models (MAGFLOW and PyFLOWGO)  
191 require topography and a step size. The model step size defines the distance (in meters) that the  
192 simulated lava advances per each model iteration. PyFLOWGO allows the user to define a step  
193 size that is independent of the DEM resolution. The influence of DEM resolution versus the model  
194 step size is investigated using a convergence analysis (see section 3.2).

195           PyFLOWGO does not directly intake a DEM profile. It requires a slope transect that  
196 follows the path of the lava flow's central channel. For each DEM resolution, a slope map was  
197 created using ArcGIS v10.7 and the "Slope" tool.

#### 198 **3.1 Topographic Data**

199           To assess the impact that DEM resolution has on any lava flow emplacement model,  
200 numerous approaches are possible. For locations where multiple DEMs of different resolutions are  
201 available, comparing the model outputs of each is possible (Stefanescu et al., 2012). The  
202 generation of artificial terrain through Monte Carlo algorithms that control the variation and  
203 amount of noise/error is another approach (Richardson & Karlstrom, 2019). For this analysis, a  
204 deterministic approach similar to the method of Bilotta et al. (2019) was used. We used a high  
205 resolution (5 m) DEM as the baseline and derived multiple lower-resolution DEMs (10, 15, 20,  
206 30, 50, 100 and 200 m) for comparison. It is important to note, however, that a DEM degraded to  
207 a lower resolution from a higher resolution DEM is not identical to a DEM produced at that same  
208 spatial resolution. However, for this study, the smaller-scale differences between these two  
209 approaches are not considered critical to the final model output.

210           The baseline topographic data pre-date the TFE-50 eruption and therefore represent the  
211 pre-flow topography. Colleagues (Viktor Dvigalo and Alexander Belousov) at the Institute of  
212 Volcanology and Seismology, Russia, generated the data necessary to produce the DEM using  
213 aerial images. The point cloud data provided were used to create the 5 m DEM with ArcGIS v10.7  
214 and the "3D Analyst" toolbox, and the "Terrain to Raster" tool. This resolution was chosen because

215 it is similar to the DEMs produced from CubeSat data and photogrammetry (i.e., Planet Labs)  
216 (Ghuffar, 2018). The moderate and low resolution DEMs represent common scales from other  
217 remote sensing data (Table 1). Lastly, for comparison we performed simulations on a constant  
218 slope that is the average value obtained on the 5 m DEM ( $11.3^\circ \pm 0.11$ ) flow path transect. Using  
219 a constant slope is relevant because this approximation is commonly used for planetary lava flow  
220 modeling (e.g., Baloga & Glaze, 2008; Garry et al., 2007; Peters et al., 2021).

### 221 **3.2 Convergence Analysis**

222 A necessary step for assessing a model is a numerical convergence analysis. For  
223 PyFLOWGO, a convergence analysis consists of varying the step size and assessing the value  
224 returned for the variables of interest. Stabilization of the variable of interest within an acceptable  
225 range determines whether the step size is sufficient. Step sizes were chosen to determine if there  
226 is any relation between the employed step size and the resolution of the DEM. We therefore tested  
227 step sizes of 1 m, 5 m, 10 m, 15 m, 30 m, 50 m, 100 m, and 200 m on the slope profiles extracted  
228 from the 5 m reference DEM and from the lower resolution DEMs with pixel size of 15 m, 30 m,  
229 200 m, as well as the constant slope profile. To assess the convergence analysis, the maximum  
230 distance traveled by the channel and final lava core temperature are used. This convergence  
231 analysis is similar to the one performed by Chevrel et al. (2018).

## 232 **4 Results**

### 233 **4.1 Effects of DEM resolution on slope values**

234 PyFLOWGO does not directly use a DEM to initiate the model but rather a 1D slope profile  
235 that is derived from the DEM and drawn along the flow path. The elevation value of a pixel of the  
236 DEM has a direct impact on the derived slope value, hence DEM resolution greatly matters (Chang  
237 & Tsai, 1991). The focus of the analysis is on the change of slope values along the DEM transect  
238 due to changing DEM resolutions. Statistics of the slope transects are shown in Table 4. As the  
239 resolution of the DEM decreases (from 5 to 200 m pixel size), the average, maximum, standard  
240 deviation, and variance of the slope decrease. A decreasing slope average from lower resolution  
241 DEMs has been observed in previous studies (e.g., Chang & Tsai, 1991; Evans, 1980; Kienzle,  
242 2004) and is well documented. The results here support these previous studies and reconfirm the  
243 relationship between DEM resolution and derived slope values.

244 The slope values implemented in PyFLOWGO depend also on the step size (i.e., which  
 245 effectively resamples the downflow slopes) so they are not always identical to the slope values  
 246 derived from the topographic profile (see section 4.2). The resampled slope values nonetheless  
 247 follow the same trend as DEM resolution decreases.

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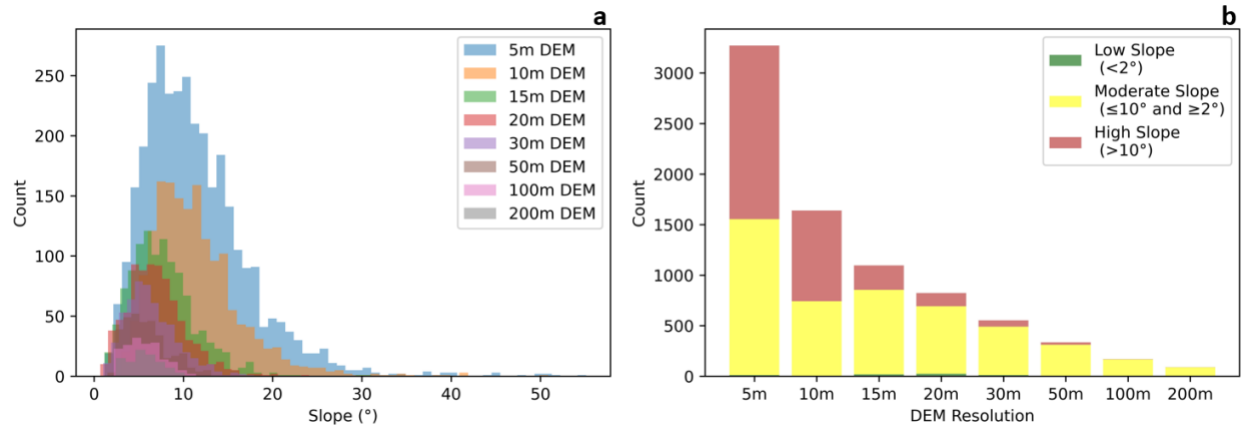
251

252 **Table 4.** Slope statistics from the channel transect for the eight DEM resolutions.

	<b>5 m DEM Slope</b>	<b>10 m DEM Slope</b>	<b>15 m DEM Slope</b>	<b>20 m DEM Slope</b>	<b>30 m DEM Slope</b>	<b>50 m DEM Slope</b>	<b>100 m DEM Slope</b>	<b>200 m DEM Slope</b>
<b>Data Points (count)</b>	3272	1640	1097	824	553	335	169	89
<b>Average Slope (°)</b>	11.4	11.4	7.70	7.01	6.39	5.91	5.60	5.60
<b>Median Slope (°)</b>	10.3	10.5	7.23	6.63	6.00	5.52	5.34	5.49
<b>Minimum Slope (°)</b>	1.19	3.03	1.16	0.69	1.00	1.49	2.20	2.43
<b>Maximum Slope (°)</b>	55.1	50.9	31.0	24.1	17.9	13.1	11.4	10.1
<b>Standard Deviation (°)</b>	6.07	5.09	3.65	3.27	2.87	2.41	2.01	1.87
<b>Variance (°)</b>	36.8	25.9	13.3	10.7	8.21	5.78	4.03	3.44

253

254 The distribution of slope values for each transect is a right skewed distribution (Figure 2a).  
 255 For the 15 m and lower resolution DEMs, as the DEM resolution decreases, the proportion of  
 256 higher slope values ( $>10^\circ$ ) significantly decreases (Figure 2b). These higher slope values are  
 257 averaged out as the DEM resolution decreases, whereas the proportion of moderate slope ( $\leq 10^\circ$   
 258 and  $\geq 2^\circ$ ) values increases. As DEM resolution decreases, moderate resolution slopes become  
 259 dominant. This indicates that the higher slope values cover a proportionally smaller part of the  
 260 downslope transect than the moderate values. This trend is also reflected in the decreasing average  
 261 slope values (Table 4). In the transect used for this study, there is not a consistent relationship  
 262 between the proportion of moderate slopes and DEM resolution when considering the full range  
 263 of DEM resolutions.



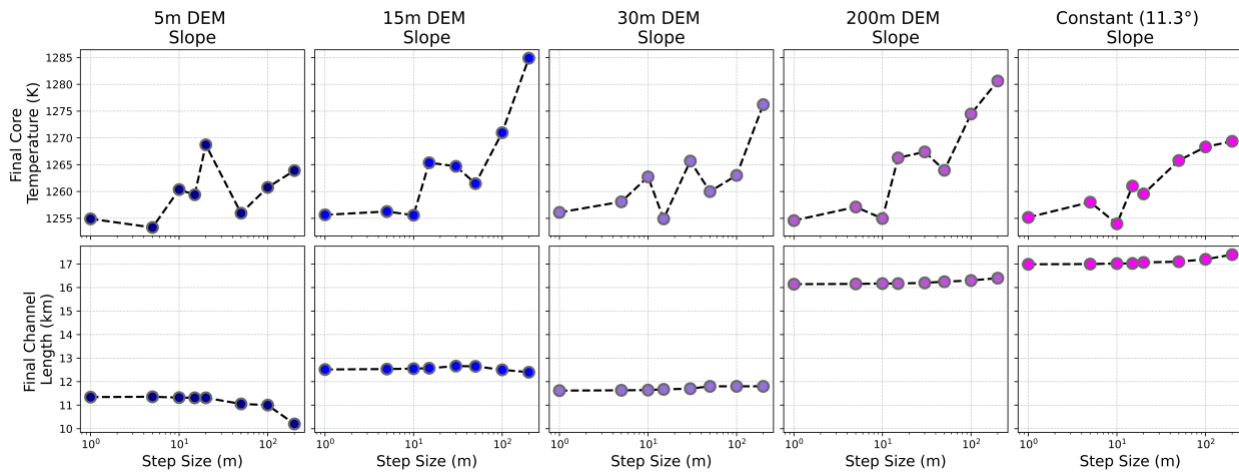
264  
265 **Figure 2.** (a) Distribution of DEM slope values for the eight DEM resolutions. (b) Slope values distributed into  
266 three groups of low (<2°), moderate (≤10° and ≥2°), and high (>10°) values. Low slope (<2°) values are so few that  
267 they do not appear on the graph.

## 268 4.2 Convergence Analysis

269 A convergence analysis was used to assess the impact of step size in relation to the DEM  
270 resolution. Figure 3 shows how the final core temperature (K) and the final channelized flow  
271 distance (km) vary with changing step size in the model.

272 As the step size decreases, the final core temperature generally decreases, stabilizing for a  
273 step size of 5 m to 10 m. The final channelized flow length stabilizes at a step size of 10 m and  
274 below for the DEM resolutions of 5 m, 15 m, and 30 m. However, for the 200 m and constant slope  
275 a step size of 15 m can be used without significantly impacting the modeled channel length. Note  
276 that the final channelized flow length does vary with DEM resolution (being higher for the 200 m  
277 DEM than for the <30 m DEM) and will be discussed in section 4.3. The convergence analyses  
278 for core temperature and channel length is consistent with Chevrel et al. (2018) that in most cases  
279 a step size of 10 m or less is preferred to ensure stabilization of the model. If a future study is  
280 focused on core temperature, then a smaller step size may be needed. This may also be true for  
281 other parameters (i.e., viscosity, crystal fraction, yield strength) and highlights the need for a  
282 convergence analysis for any output parameter of interest. Other lava flow models that incorporate  
283 a step size (i.e., MAGFLOW) have also determined a step size between 5 m and 10 m is optimal  
284 (Bilotta et al., 2019). However, the smaller the step size, the longer the computation time, which  
285 is also a function of the length of the lava flow modeled. If rapid results for hazard response are  
286 required, the choice of step size may be a necessary consideration.

287



288

289 **Figure 3.** Convergence of the PyFLOWGO runs at 1, 5, 10, 15, 30, 50, 100, 200 m step size for the four different  
290 DEM derived slopes and the one constant slope ( $11.3^\circ$ ). The final channel length (bottom row) represents the distance  
291 the flow traveled once the velocity reaches zero.

292

293 If the step size is smaller than the resolution of the DEM, the model interpolates a profile  
294 between the two points and calculates slope values for each step. If the model step size is larger  
295 than the DEM resolution, the model averages over the slope values inside the step. Therefore, the  
296 amount of topographic variability can also impact the choice of step size. If there is little  
297 topographic variability (i.e., a smooth plain), then a larger step size could be used with potentially  
298 minor impact on the model. However, if the topography is highly variable (i.e., valleys, obstacles,  
299 etc.) then a smaller step size is required. To accurately and fully account for the topography present  
300 in a transect, a step size equal to or smaller than the DEM resolution should be used. To allow for  
301 a uniform comparison for the remainder of this manuscript, results using a 5 m step size are  
302 presented for all DEM resolutions examined.

### 303 4.3 Modeled Channel Length

304 We define the baseline model using the pre-flow 5 m DEM and all rheological model inputs  
305 from Ramsey et al. (2019), excluding the variable emissivity module as discussed in the  
306 Supplemental Material. To initialize the model and match the observed flow length of 11.3 km,  
307 Ramsey et al. (2019) used an ASTER-derived 30 m DEM and an effusion rate of  $278 \text{ m}^3/\text{s}$ . This  
308 initial effusion rate was taken from an estimation of the time averaged discharge rate (TADR)

309 obtained by the Middle Infrared Observation of Volcanic Activity (MIROVA) system (Coppola et  
310 al., 2016). Here, using the new pre-flow 5 m DEM and variable emissivity module, we obtained  
311 the desired channel length of 11.3 km for an effusion rate of 395 m<sup>3</sup>/s. This value is slightly above  
312 the estimation from MIROVA but is reasonable as it falls in the range determined for the early  
313 phase of the eruption via field observations (aerial surveys) (440 m<sup>3</sup>/s), the MIROVA system (278  
314 m<sup>3</sup>/s), and the TanDEM-X satellite (417 m<sup>3</sup>/s) (Dvigalo et al., 2013; Kubanek et al., 2017; Ramsey  
315 et al., 2019). The effusion rate of 395 m<sup>3</sup>/s is closer to the values of ~400 m<sup>3</sup>/s or more determined  
316 by TanDEM-X and aero-photogrammetric observations that occurred during the first few days of  
317 the eruption.

318 Varying the resolution of the DEM, all simulated flows traveled farther than the simulation  
319 on the 5 m DEM for the same effusion rate (Figure 4). The simulation using the 10 m DEM resulted  
320 in a 12.8 km flow (12% increase); the 15 m DEM resulted in a 12.5 km flow (10% increase); the  
321 20 m DEM resulted in a 12.4 km flow (9% increase); the 30 m DEM resulted in a 11.6 km flow  
322 (3% increase); the 50 m DEM resulted in a 11.5 km flow (1% increase); the 100 m DEM resulted  
323 in a 13.1 km flow (14% increase); and the 200 m DEM resulted in a 16.2 km flow (35% increase).  
324 The simulation on the constant slope the flow reached 17.0 km (40% increase) (Figure 4). All  
325 model runs stopped because the flow velocity reached zero. Good model fits (<5% increase in flow  
326 length) were obtained with the 5 m (baseline model), 30 m and 50 m DEMs. The 10 m, 15 m, 20  
327 m, 100 m, 200 m, and constant slope had worse fits compared to the baseline model. Based on the  
328 slope statistics in Table 4, specifically the decreasing trend in average slope values with decreasing  
329 DEM resolution, the non-regular relationship between DEM resolution and modeled channel  
330 length was not expected. Analyses of slope values at Tolbachik (Figure 2b) show that the  
331 proportions of low, moderate, and high slopes do not systematically change with decreasing DEM  
332 resolution and may have an effect on modeled channel length.

#### 333 **4.4 Effusion Rate Variations**

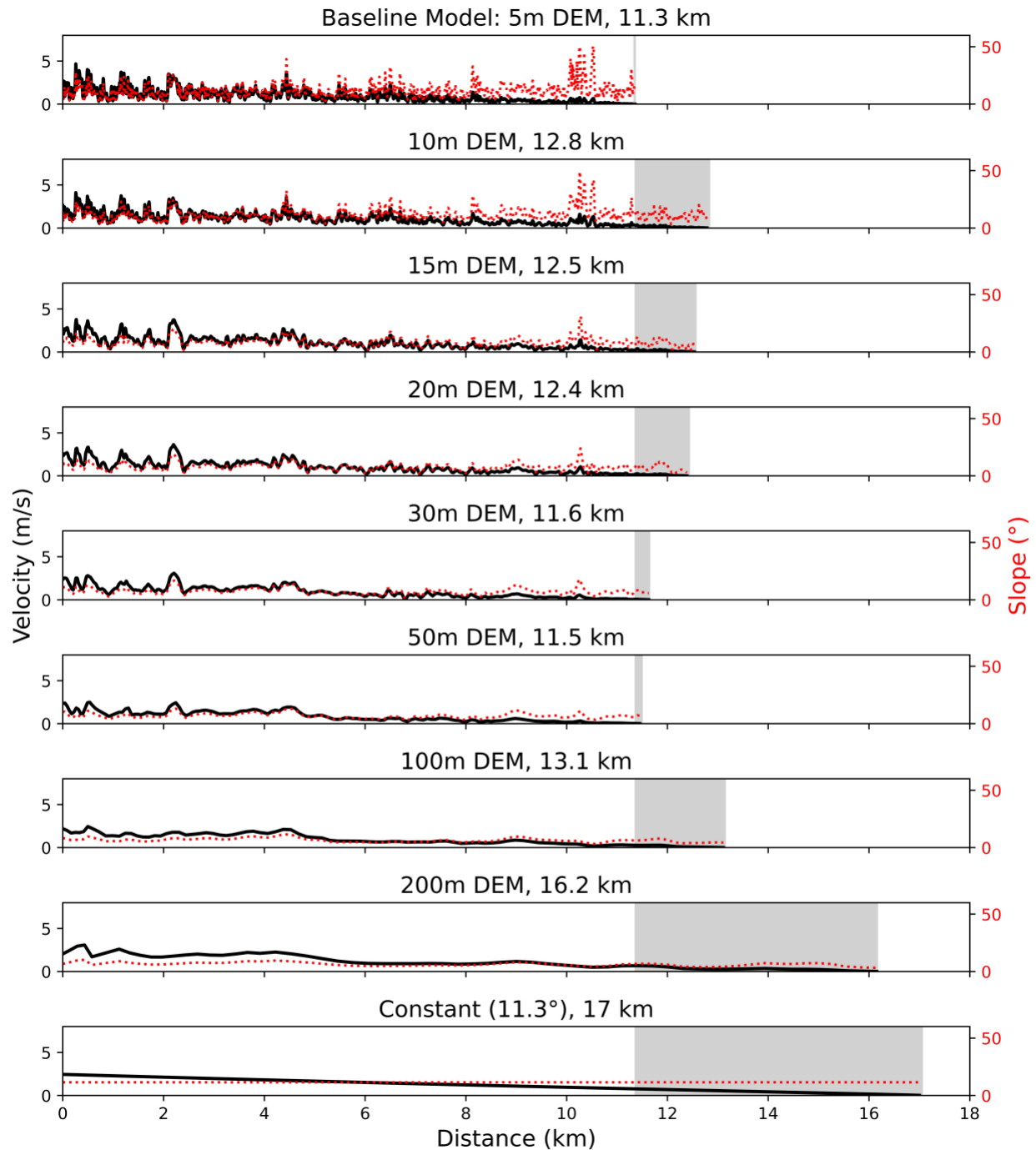
334 Effusion rate is one input for PyFLOWGO (or any lava flow model) that greatly influences  
335 lava flow length but is subject to large uncertainty (Harris et al., 2007; Rowland et al., 2005). An  
336 effusion rate is typically calculated indirectly from satellite thermal data, or directly from  
337 observations made in the field, and used in the model to match a defined channel/flow length or  
338 area. To assess the relationship between effusion rate and the resolution of the DEM, PyFLOWGO  
339 was run iteratively with different effusion rates for all seven DEM resolutions plus the constant



340 slope to match the 11.3 km channel length (Figure 5). All other input parameters remained  
341 constant, including the 5 m step size, for all DEM resolutions. The 5 m DEM is the baseline model  
342 as defined in section 4.3.

343 Figure 5 shows results for effusion rates for the eight DEM resolutions; these follow a non-  
344 regular trend, similar to the channel length results shown in Figure 4. Effusion rates similar to the  
345 baseline model at Tolbachik were found for the 30 and 50 m DEMs, whereas significantly lower  
346 values for the 10 m, 15 m, 30 m, 100 m and 200 m DEMs are needed to match the final channel  
347 length. This pattern is attributed to the decrease in topographic variability of the Tolbachik slope  
348 transect (Table 4 and Figure 2). For a low-resolution DEM (100 m and 200 m) or constant slope,  
349 there is substantially less topographic variability expressed in the transect. A markedly smaller  
350 effusion rate is necessary to reach the 11.3 km distance. With less topographic variability  
351 influencing the modeled flow, a lower effusion rate can therefore produce a lava flow of the same  
352 length. This could potentially lead to an underestimation of the effusion rate when using the model  
353 with lower resolution DEMs (e.g.,  $\geq 100$  m) to determine a flow's emplacement conditions.

354 For many terrestrial volcanic eruptions, effusion rates spike early then follows an  
355 exponential decay (e.g., Bonny et al., 2018; Harris et al., 2000; Ramsey et al., 2019; Wadge, 1981;  
356 Wright et al., 2001). Most channelized lava flows are formed during this early period in the  
357 eruption (Rowland et al., 2005). The average initial effusion rate determined for 27 - 29 November  
358 2012 from field observations at Tolbachik (i.e., aerial surveys) was  $\sim 400$  m<sup>3</sup>/s (Dvigalo et al.,  
359 2014; Kubanek et al., 2017). The observed flow length of 11.3 km occurred on 1 December 2012,  
360 early in the eruption. Using the higher resolution topographic data required a higher effusion rate  
361 to reach the same distance (Figure 5). In other words, the estimate of the effusion rate that produced  
362 a given flow should improve with higher resolution topography.

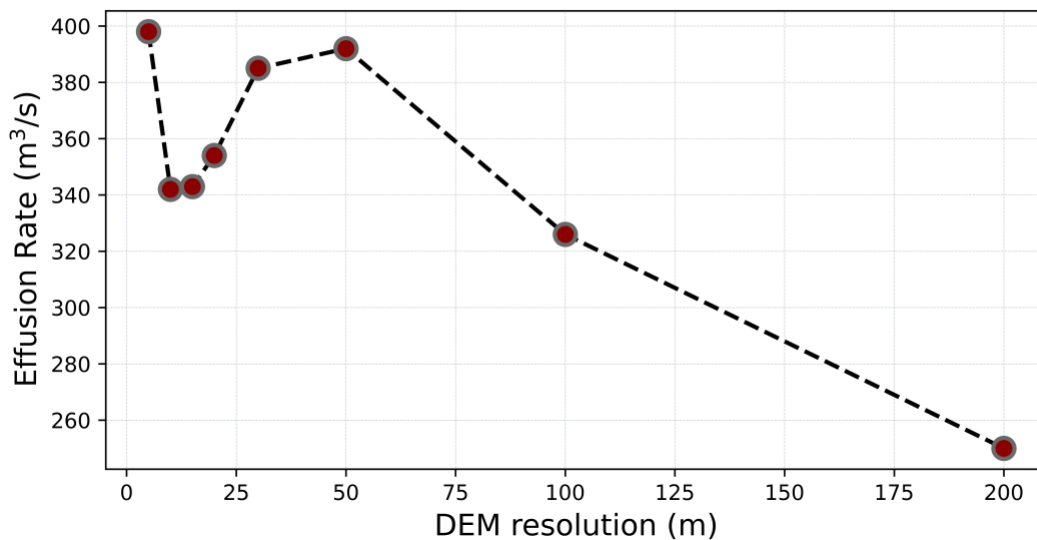


363

364 **Figure 4.** Model runout distances at a constant effusion rate ( $395 \text{ m}^3/\text{s}$ ) for eight DEM resolutions, and a constant  
365 slope. The black line represents lava velocity in the channel as a function of distance from the source. The red dotted  
366 line is the slope profile. The grey vertical line in the baseline model plot is the 11.3 km flow length observed on 1  
367 December 2012, and the grey regions indicate the difference between the flow length of baseline model and the  
368 relevant model case. All model results shown here used a 5 m step size.

369

370



371

372 **Figure 5.** Required effusion rates as a function of DEM resolution to reach the 11.3 km channel length for the 5 m  
373 (395 m<sup>3</sup>/s), 10 m (341 m<sup>3</sup>/s), 15 m (342 m<sup>3</sup>/s), 20 m (353 m<sup>3</sup>/s), 30 m (384 m<sup>3</sup>/s), 50 m (391 m<sup>3</sup>/s), 100 m (325 m<sup>3</sup>/s)  
374 and 200 m (249 m<sup>3</sup>/s) model runs, and for the constant slope (11.3°) (248 m<sup>3</sup>/s). All model results shown here used a  
375 5 m step size.

376

#### 377 4.5 PyFLOWGO Velocity

378 PyFLOWGO accounts for the change in topography through a modified version of Jeffreys  
379 equation to determine the flow velocity (Moore, 1987). To assess the lack of consistent channel  
380 length and effusion rate trend with decreasing DEM resolution observed in Figure 4 and 5, we next  
381 divided each modeled flow into three segments (proximal, medial, and distal) to compare changes  
382 in flow velocity as a result of the DEM resolution in those segments (Table 5). This was intended  
383 to explore how the distribution of slopes downflow might affect model results for different DEM  
384 resolutions as the higher slopes are reduced/smoothed in lower resolution DEMs. Proximal  
385 represents the first 10% of the channel length, medial the middle 65%, and distal the final 25%.  
386 These percentages are similar to those used for terrestrial and Martian lava flows in prior studies  
387 (Garry et al., 2007; Peitersen & Crown, 1999). The constant slope model run is not assessed  
388 because variations in topography are not present.

389

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**Table 5:** Average modeled velocity for the proximal, medial, and distal flow portions. The last two columns are the average and median flow velocity over the entire modeled channel length.

DEM Resolution	<i>Modeled Channel Velocity (m/s)</i>				
	Proximal (first 10%)	Medial (middle 65%)	Distal (final 25%)	Average Over Channel Length	Median Over Channel Length
5 m	1.68	1.07	0.27	0.92	0.76
10 m	1.87	1.12	0.34	0.99	0.89
15 m	1.95	1.12	0.30	0.99	0.85
20 m	1.88	1.09	0.25	0.96	0.85
30 m	1.69	1.01	0.25	0.88	0.72
50 m	1.55	0.96	0.23	0.83	0.61
100 m	1.75	1.08	0.22	0.93	0.74
200 m	2.30	1.22	0.24	1.08	0.91

393

394 There is a decrease in the average and median flow velocities for the 10 m to 50 m DEM  
 395 resolutions (Table 5). The same trend is not fully observed in the proximal, medial, or distal  
 396 regions. There is an increase in velocity in the 10, 15, and 20 m DEM results for the proximal,  
 397 medial, average, and median relative to the 5 m DEM results. For the distal region, the increase is  
 398 seen in the 10 and 15 m results and not for the 20 m DEM. These trends follow the lack of  
 399 consistent relationships with DEM resolution for channel length and effusion rate and are  
 400 attributed to the specific topographic characteristics of the Tolbachik slope transect that influences  
 401 flow velocity.

402 The decrease in flow velocity is attributed to the change in topographic variability at  
 403 different locations along the slope transect (Table 4 and Figure 4). The slope transect used for the  
 404 modeling contained large topographic changes that take up a proportionally smaller area than do  
 405 the small to moderate changes (Figure 2). As large topographic changes (and by extension large  
 406 slopes) are lost in the lower resolution DEMs, the flow is unable to maintain the same velocity to  
 407 allow it to reach the same length as those derived from using the higher resolution DEMs. For high  
 408 to moderate resolution DEMs (10 to 50 m), the fewer large slope changes in the transect leads to  
 409 a lower average flow velocity. However, for the low resolution DEMs (100 m and 200 m) and  
 410 constant slope there is an increase in both flow length and velocity in comparison to the moderate  
 411 resolution DEMs (Figure 4 and Table 5). With these lower resolution DEMs, the topography has

412 smoothed to the point where there is little to no variability left to influence the final length (Table  
413 4 and Figure 2). Although there are not high slope values to increase velocity, there are also not  
414 the corresponding decreases in slope that reduce velocity. The lower resolution DEMs can  
415 maintain a higher velocity and thus a longer flow distance.

416 For the case shown here, topographic effects and the resulting changes in velocity of the  
417 modeled lava flow might be site specific to the Tolbachik volcanic complex. These results  
418 highlight the importance for future studies to examine the magnitude and distribution of slope  
419 elements at the scale of the DEM to assess how they may influence the modeling results.

420

## 421 **5 Discussion**

422 The resolution of the DEM used for PyFLOWGO modeling has a measurable impact on  
423 the resulting channel length. These findings differ from Harris et al. (2016) who used two different  
424 topographic datasets, a 30 m DEM from the SRTM and a 10 m DEM produced from LIDAR, to  
425 compare the effect of different DEM resolutions on the FLOWGO model for a channelized lava  
426 flow from the December 2010 eruption of Piton de la Fournaise (Harris et al., 2016). Their study  
427 found that there is little difference between the two resolutions but did not elaborate on the details,  
428 since the focus of the study was on comparing field measurements to model results and not a  
429 detailed topographic sensitivity analysis. Comparing the results here for only the 30 m and 10 m  
430 DEMs, the length difference is 1.16 km or 9.5%, a potentially significant difference for lava flow  
431 hazard assessments and maps.

432 The changes in topographic variability related to the DEM resolutions and the effect on the  
433 calculated velocity partially explains the trend in flow lengths (Figure 2, Figure 4, and Table 5).  
434 However, we cannot rule out the possibility that the trend in channel length may be impacted by  
435 specific aspects of the topography at Tolbachik. The slope profile used in the study shows large  
436 variations near the proximal and distal end of the flow (Figure 4). These variations in slope are not  
437 fully averaged out except in the 100 m and 200 m DEM (Figure 2). Previous studies have shown  
438 that changes in topography at different points along a flow's emplacement can impact the final  
439 flow length and morphology (Flynn et al., 2021; Gregg & Fink, 2000; Guest et al., 1987).  
440 Assessing how different unique combinations of topographic variability impact lava flows  
441 emplacement is outside the scope of this work but should be further investigated.

442 Results from the topographic sensitivity analysis indicate that using a higher resolution  
443 DEM requires a higher effusion rate to model the same channel length especially for regions with  
444 large pre flow slope variability. Moderate to low resolution DEMs will underestimate the modeled  
445 effusion rate in these same circumstances. The remainder of the discussion centers on the  
446 implications of using different DEM resolutions for modeling both active and inactive lava flows.

## 447 **5.1 Active Flows**

448 PyFLOWGO is increasingly used to make hazard assessments for new volcanic events  
449 within minutes to hours of an eruption occurring (Chevrel et al., 2022; Harris et al., 2019; Peltier  
450 et al., 2021, 2022). In these cases, lava flow composition (i.e., rheological parameters) is usually  
451 determined based on past eruptions at that volcano. The two remaining inputs needed for the model  
452 are the effusion rate and the topography. Effusion rate for active eruptions is commonly calculated  
453 from repeat visible, thermal, or radar satellite data (Harris et al., 2007 and methods described  
454 within). Common tools for determining effusion rates from satellite data are the MIROVA  
455 (Coppola et al., 2016), HOTVOLC (Gouhier et al., 2016), and HOTSAT (Ganci et al., 2011)  
456 monitoring systems and data from the TanDEM-X satellite (Kubanek et al., 2021). Our results  
457 indicate that using the highest resolution DEM available, with an effusion rate from early in the  
458 eruption, produces the best fit to the observed channel length (provided the model is calibrated to  
459 that DEM resolution and chosen model step size as we show here). In an ideal situation, a high-  
460 resolution topographic dataset (i.e., 10 m DEM or better) is available coupled with effusion rate  
461 measurements made as soon as possible at the start of the eruption. However, if a study uses a  
462 moderate resolution DEM in combination with PyFLOWGO, an overestimate of the final channel  
463 length is possible depending on the pre-flow topography.

464 For some volcanoes (e.g., Etna, Kilauea, Piton de la Fournaise), a 5 m (or better) DEM may  
465 be available. However, for many other volcanoes, especially those in remote locations, high  
466 resolution DEMs are not readily available or feasible to acquire. In those cases, a moderate  
467 resolution DEM (i.e., 30 m), such as the SRTM DEM or ASTER global DEM (GDEM) are freely  
468 available and provide global coverage. However, the SRTM DEM provides a snapshot of  
469 topography from February 2000, potentially missing important topographic changes since that  
470 time (van Zyl, 2001). In contrast, version 3 of the ASTER GDEM is more current, but averages  
471 together ASTER individual scene DEMs from 2000-2011. Averaging scenes together for the  
472 GDEM does produce a more accurate DEM, devoid of clouds and other errors, but it also averages

473 any topographic change during that period. Alternatively, an ASTER individual scene DEM (30  
474 m/pixel) could be used. However, as highlighted in Flynn & Ramsey (2020), these scenes can  
475 contain substantial amounts of noise or interference from clouds and shadowing, despite being  
476 more representative of recent topographic change.

477 The impact of noise (and/or rapidly changing topography) in a DEM could be reduced  
478 using the “slope smoothing” function in PyFLOWGO. This function interpolates a line over the  
479 DEM based on a set number of points, defined by an individual before the model starts. It is  
480 important to note that, although using this function may reduce noise in a transect, it also removes  
481 naturally occurring topographic features smaller than the smoothing function. As highlighted in  
482 Harris et al. (2019) and Chevrel et al. (2022), calibration of PyFLOWGO for a given DEM  
483 resolution and step size is mandatory before using it for any hazard assessment. This ensures that  
484 the modeled lava flow distance for a given effusion rate is as accurate as possible.

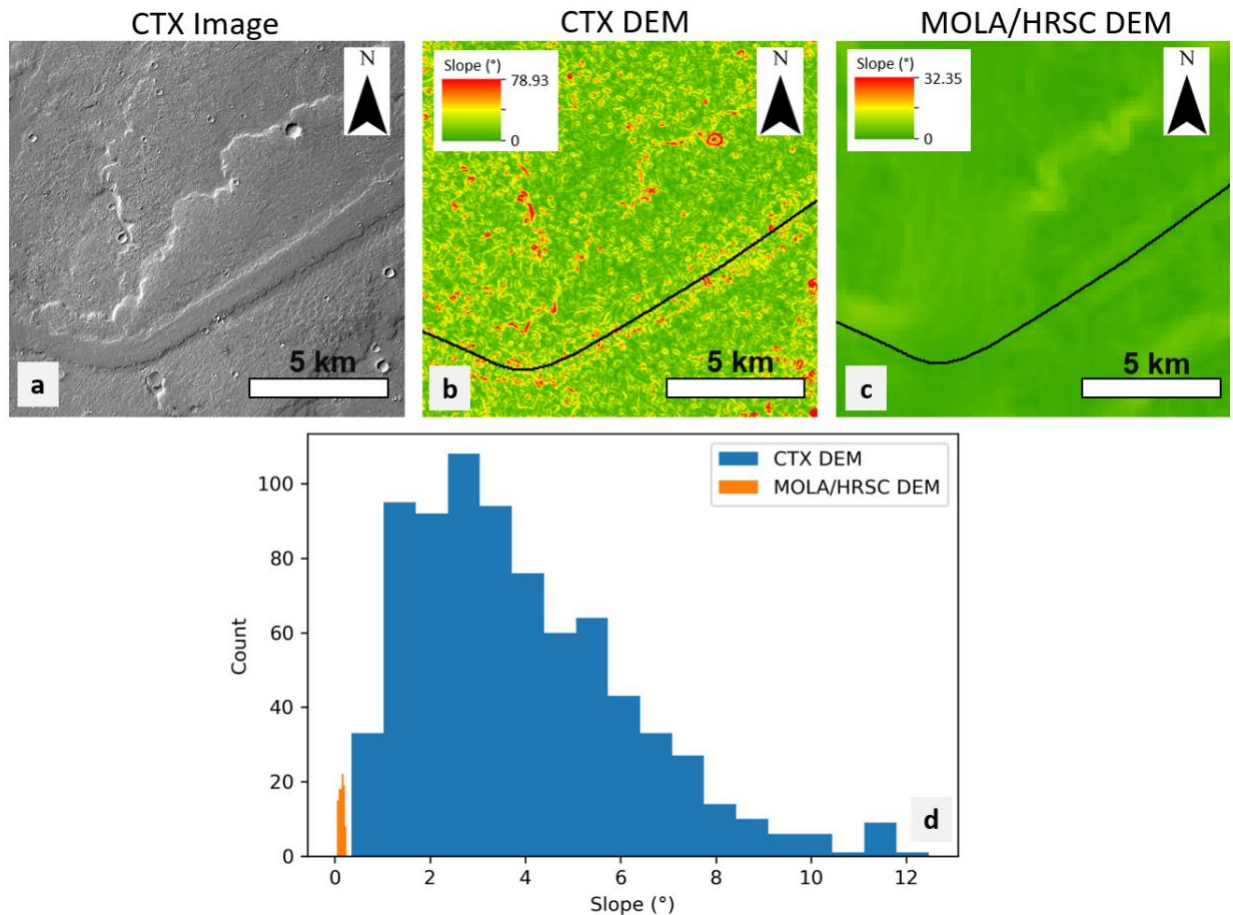
## 485 **5.2 Inactive Flows**

486 PyFLOWGO can also be used to investigate the evolution of the thermorheological  
487 properties of already emplaced/inactive flows on Earth and other planetary bodies (cf. Table 2)  
488 (e.g., Lev et al., 2021; Ramsey et al., 2019; Rowland et al., 2004). Morphological parameters of  
489 terrestrial lava flows and rheological properties of the lava can be established through field  
490 campaigns and sample analysis, respectively. However, for flows on other planetary bodies,  
491 rheological conditions typically are assumed using an Earth analog or modeling (Chevrel et al.,  
492 2014). As discussed in Section 1, DEM resolution is limited for other planetary bodies, and pre-  
493 flow topography has not been available. Therefore, post-emplacment topography is assumed to  
494 be a suitable proxy.

495 Comparing the results for the 200 m DEM (the same resolution as the MOLA/HRSC global  
496 DEM for Mars) to the results for the constant slope, the modeled channel length difference is 850  
497 m (5%). This is a potentially significant difference where attempting to reproduce very long (e.g.,  
498 100s of km) channelized lava flows observed on Mars or Venus. However, the difference in  
499 effusion rate required to model a given flow using the 200 m DEM and constant slope scenarios is  
500 negligible 0.4% (Figure 5). In order to apply PyFLOWGO to planetary lava flows the use of lower  
501 resolution DEMs or a constant slope is commonly required and will introduce uncertainty into the  
502 results. Results presented here quantify that uncertainty, which can aid future applications to  
503 planetary lava flows.



504 A potentially important factor typically overlooked in studies that use a constant slope for  
505 lava flow modeling is the value chosen. As shown in Table 4, as the resolution of the DEM  
506 decreases the average slope value also decreases and becomes less representative of the local  
507 topography. Figure 6 shows that the change in measured slope value relates to the resolution of  
508 the topographic dataset using data commonly available on Mars (i.e., a CTX and MOLA/HRSC  
509 DEM).  
510



511  
512 **Figure 6:** (a) A CTX image (P02\_001775\_1593) of a channelized lava flow located south of Arsia Mons, Mars. The  
513 central channel has been filled with a secondary lava flow, which distinctly outlines its path. (b) CTX slope map of  
514 the same location in (a) and the black line indicating the location of the central channel. (c) MOLA/HRSC slope map,  
515 derived from the MOLA/HRSC DEM. (d) Histogram distribution of slope values along the black line in (b) and (c).  
516 The average value for the CTX and MOLA/HRSC slope transects were 3.92° and 0.14°, respectively.  
517

518 Previous modeling of Martian lava flows has used a constant slope (i.e., the standard  
519 rheologic approach described in Hiesinger et al., 2007). This value is usually derived from HRSC  
520 or MOLA data (e.g., Hiesinger et al., 2007; Pasckert et al., 2012; Peters et al., 2021). As shown in  
521 Table 4, Figure 2, and Figure 6, the higher resolution DEMs can provide a large range of slope  
522 values dependent on the resolution of the topographic dataset. This range in slope values can have  
523 a quantifiable effect on modeling results for effusion rate, viscosity, and yield strength (Russo et  
524 al., 2022). However, using a constant slope derived from the regional topography may be the best  
525 option if no pre-flow topography is available.

526 To assess the effects of the constant slope values we perform an additional model run using  
527 the average slope value derived from the 200 m DEM ( $5.6^\circ$ ) (Table 4). We compare the  $5.6^\circ$   
528 constant slope model run to those from the  $11.3^\circ$  constant slope results. Our results show that the  
529 differences between a simulation on a constant slope of  $5.6^\circ$  or  $11.3^\circ$  is only 960 m (5%). Given  
530 the large (nearly two times) difference in average slope value, this is a relatively minor difference  
531 in distance. Although using a constant slope for lava flow modeling is not optimal, given the  
532 limitations of planetary datasets, errors related to this approximation are acceptable for a first order  
533 estimation of the lava flow emplacement dynamics.

## 534 **6 Conclusions**

535 The main objective of this work was to investigate the effect of DEM resolution on  
536 PyFLOWGO modeling. We used the 11.3 km-long lava flow emplaced during the Tolbachik 2012-  
537 2013 eruption as a reference and the same PyFLOWGO rheologic input parameters as provided  
538 by Ramsey et al. (2019) with the exception of the emissivity model for that same flow. Our  
539 approach consisted of comparing PyFLOWGO results on DEMs of decreasing resolution (from 5  
540 m to 200 m).

541 Using a convergence analysis, we emphasize the importance of choosing an appropriate  
542 model step size to account for topography. In agreement with Chevrel et al. (2018) we show that  
543 a step size of 10 m or smaller should be used for a DEM with resolutions between 5 and 200 m.  
544 Choosing an appropriate model step size in relation to the DEM resolution is a necessary step to  
545 ensure accurate model results.

546 The topographic sensitivity analysis results indicate that DEM resolution is an important  
547 parameter to consider with PyFLOWGO. Using a moderate to low resolution ( $\sim 10$  m to  $\sim 200$  m)  
548 DEM overestimates the channelized flow length, as does using a constant slope equal to the

549 average slope value. In addition, assessing the level of topographic variability in a modeled lava  
550 flows path is an important consideration. This information supports the fact that regardless of the  
551 level of complexity in a lava flow model, topography and its accuracy are critical parameters.  
552 Studies using PyFLOWGO for active lava flow emplacement should strive to use the most recent  
553 pre-flow DEM, especially for cases where the topography is frequently changing. If moderate to  
554 low resolution DEMs are the only available options, either the effusion rate will be underestimated,  
555 or the final channel length overestimated. However, where applying PyFLOWGO to inactive lava  
556 flows (on Earth or other planetary bodies), a moderate to low resolution DEM may be the only  
557 option. Interpreting these model results should be done with the expectation of higher error rates  
558 in either the effusion rate or modeled channel length.

559         With the wide range of applications for PyFLOWGO, across multiple planetary bodies,  
560 understanding the impact of topographic datasets resolution is important for accurate modeling  
561 results. The topographic sensitivity analysis presented here provides useful information that can  
562 be used to improve the application of PyFLOWGO for hazard assessments and investigation into  
563 the emplacement of channelized flows on Earth and other planetary bodies.

564

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572 available for download at <https://github.com/pyflowgo>. PyFLOWGO results can be found at Flynn  
573 (2023). This is a Laboratory of Excellence ClerVolc contribution no°594.

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855 **Table 3.** PyFLOWGO input models, modules, and parameters that were used to simulate the 11.3 km channelized  
 856 Leningradskoye lava flow from 1 December 2012 (e.g., Ramsey et al., 2019). The \* next to step size indicates that  
 857 this parameter was varied for a part of this study. For a full breakdown of abbreviations in the table see Chevrel et al.  
 858 (2018).

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<b>Models</b>	<b>Selection</b>	<b>References</b>
Crystallization Rate	basic	Harris and Rowland (2001)
Melt Viscosity	vft	Giordano et al. 2008)
Relative Viscosity	er	Einstein-Roscoe model from Chevrel et al. (2018)
Yield Strength	ryerson	Ryerson et al. (1988)
Crust Temperature	constant	Harris and Rowland (2001)
Effective Crust Cover	basic	Harris and Rowland (2001)
Vesicle Fraction	constant	Harris and Rowland (2001)
<b>Heat Budget Modules</b>	<b>Selection</b>	<b>References</b>
Radiation	yes	Harris and Rowland (2001) and Ramsey et al. (2019)
Conduction	yes	Harris and Rowland (2001)
Convection	yes	Harris and Rowland (2001)
Rain	no	Harris and Rowland (2001)
Viscous Heating	no	Harris and Rowland (2001)
<b>Initial Input Parameters</b>	<b>Value</b>	<b>References</b>
Step Size*	10 m	Harris and Rowland (2001)
Effusion Rate*	278 - 400 m <sup>3</sup> s <sup>-1</sup>	MIROVA (this study)
Width	30 m	Ramsey et al. (2019)
Depth	6.1 m	Ramsey et al. (2019)
Gravity	9.81 ms <sup>-2</sup>	

Eruption Temperature	1355.15 K	Plechov et al. (2015)
Crystal Fraction	25%	Plechov et al. (2015)
Dense Rock Equivalent Density	2630 kg m <sup>3</sup>	Plechov et al. (2015)
Vesicle Fraction	0.06	Plechov et al. (2015)
Stefan-Boltzmann constant	5.60E-08	
Emissivity Crust	0.95	Ramsey et al. (2019)
Emissivity Uncrusted	0.6	Lee & Ramsey, (2016)
Basal Temperature	773.15 K	Harris and Rowland (2001)
Distance from Core to Base	19%	Harris and Rowland (2001)
Wind Speed	5 ms <sup>-1</sup>	Harris and Rowland (2001)
Air Friction Factor	0.0036	Harris and Rowland (2001)
Air Temperature	273.15 K	Harris and Rowland (2001)
Air Density	0.4412 kg m <sup>-3</sup>	Harris and Rowland (2001)
Air Specific Heat Capacity	1099 J kg <sup>-1</sup> K <sup>-1</sup>	Harris and Rowland (2001)
Buffer	140 °C	Harris and Rowland (2001)
Crust Cover Fraction	0.9	Ramsey et al. (2019)
Velocity Dependency of Crust	-0.16	Ramsey et al. (2019)
Crust Temperature	773.15 K	Belousov et al. (2015)
a vft	-4.55 Pa s	Volynets et al. (2015)
b vft	6887.303 J mol <sup>-1</sup>	Volynets et al. (2015)
c vft	527.44 K	Volynets et al. (2015)
Crystal Growth	0.37	Ramsey et al. (2019)
Solid Temperature	1253.15 K	Ramsey et al. (2019)
Latent heat of Crystallization	350000 K kg <sup>-1</sup>	Harris and Rowland (2001)

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