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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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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 (≤ 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.

Planetary Body	Instrument	Spatial Resolution	Global Surface Coverage (%)	References
<i>Earth</i>	ASTER ^a	30 m/pixel	100%	(Hirano et al., 2003)
	SRTM ^b	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 ^c	463 m/pixel	100%	(Zuber et al., 1992)
	MOLA/HRSC ^d	200 m/pixel	100%	(Ferguson et al., 2018)
	HRSC	50 m/pixel	~44%	(Jaumann et al., 2007)
	CTX ^e	~20 m/pixel	~17%	(Breton et al., 2019)
	HiRISE ^f	~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)

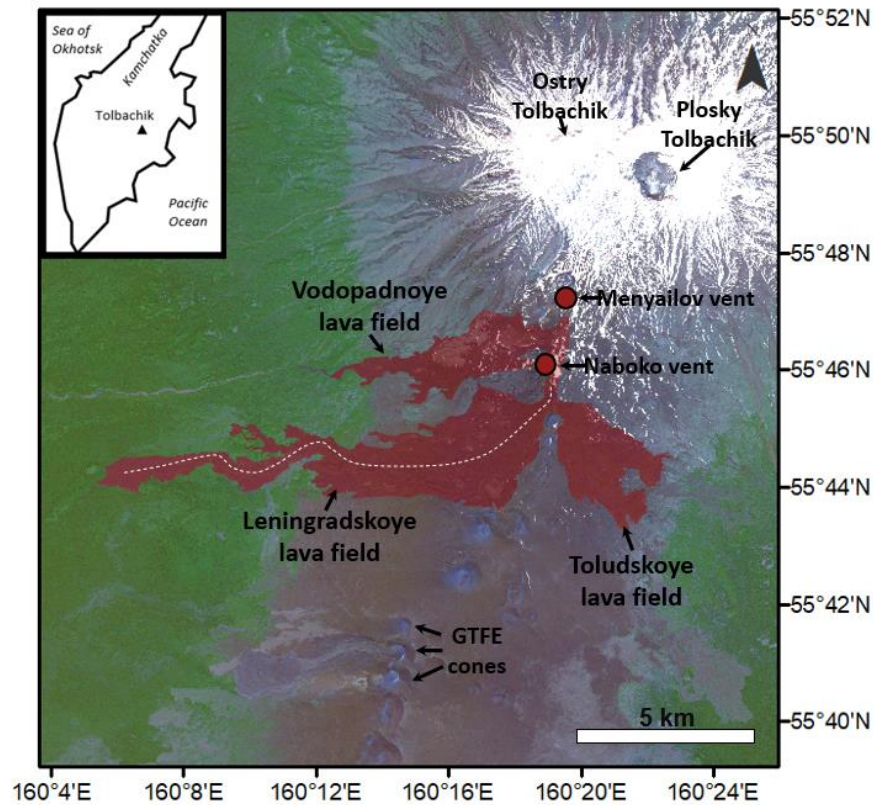
72 ^a Advanced Spaceborne Thermal Emission and Reflection radiometer

- 73 ^b Shuttle Radar Topography Mission
74 ^c Mars Orbiting Laser Altimeter
75 ^e Context Camera
76 ^f 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 50th 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 km^2 , volume 0.04 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 km^2 , volume $0.35\text{-}0.39 \text{ km}^3$) and Toludskoye (area 8.4 km^2 ,
117 volume 0.1 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_{rad}) heat loss from the flow surface, convective
152 (Q_{conv}) heat loss to the atmosphere, conductive (Q_{cond}) heat loss to the channel walls and floor, and
153 heat loss due to vaporization of precipitation (Q_{rain}) (where applicable). The heat gains are due to
154 latent heat of crystallization (Q_{cryst}) and viscous dissipation (Q_{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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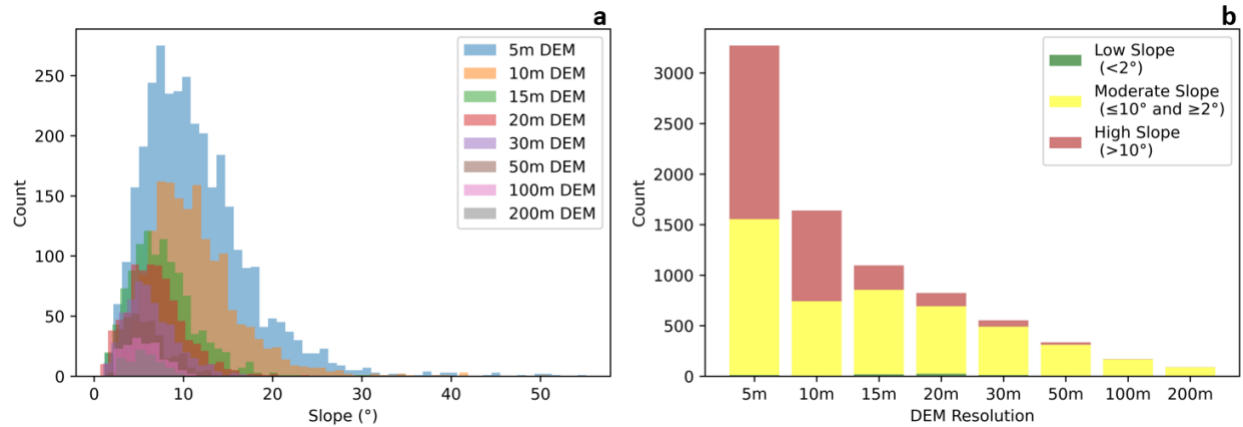
251

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

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



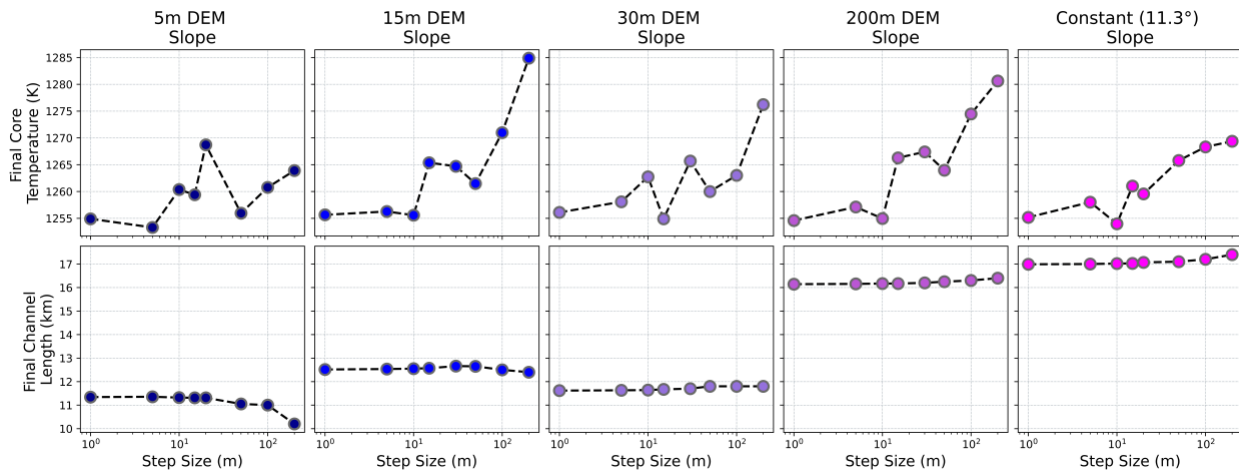
264
265 **Figure 2.** (a) Distribution of DEM slope values for the eight DEM resolutions. (b) Slope values distributed into
266 three groups of low (<2°), moderate (≤10° and ≥2°), and high (>10°) values. Low slope (<2°) values are so few that
267 they do not appear on the graph.

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°). 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 m³/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 m³/s), the MIROVA system (278
314 m³/s), and the TanDEM-X satellite (417 m³/s) (Dvigalo et al., 2013; Kubanek et al., 2017; Ramsey
315 et al., 2019). The effusion rate of 395 m³/s is closer to the values of ~400 m³/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.

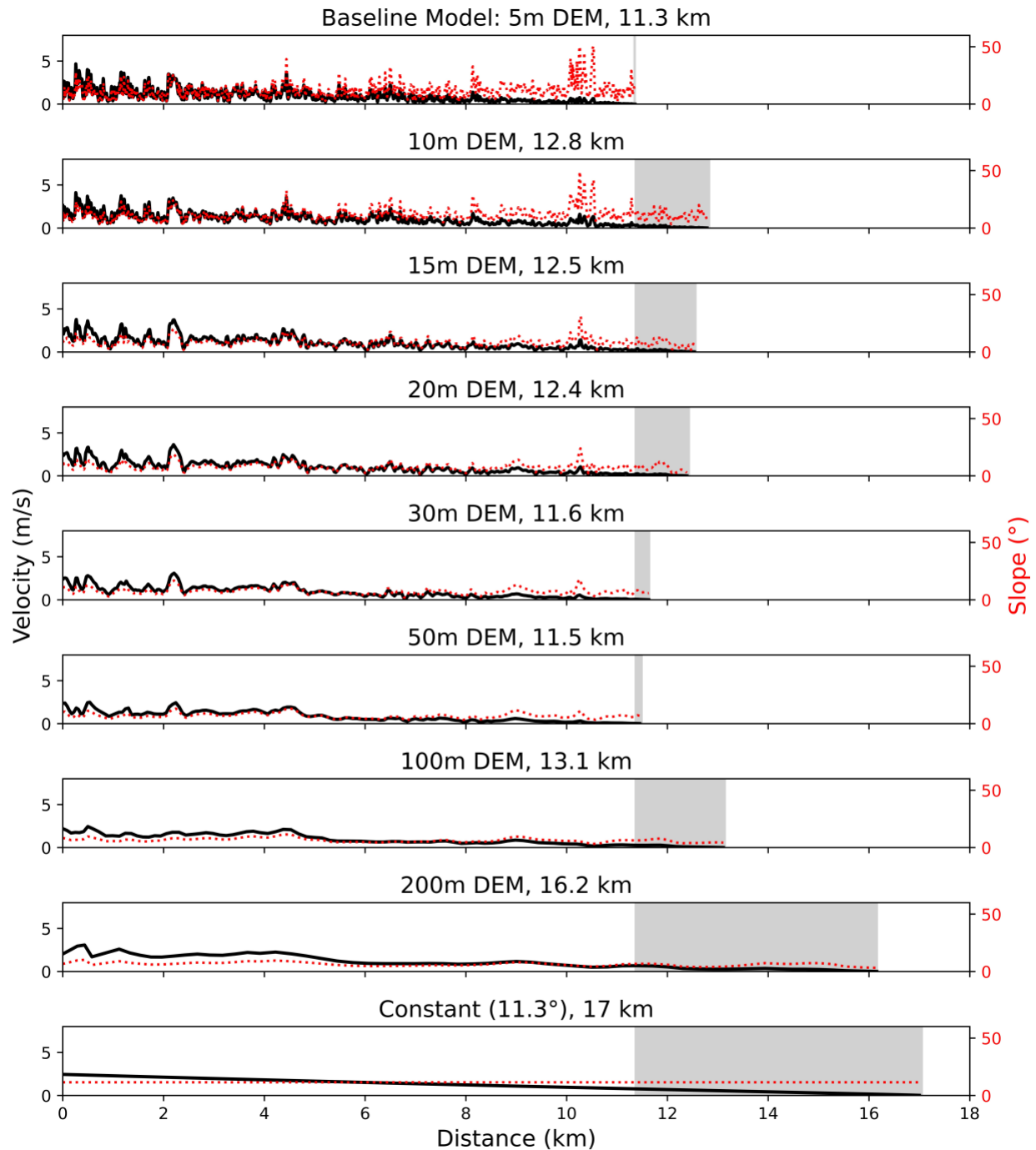
333 **4.4 Effusion Rate Variations**

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., ≥ 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 ~ 400 m³/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.

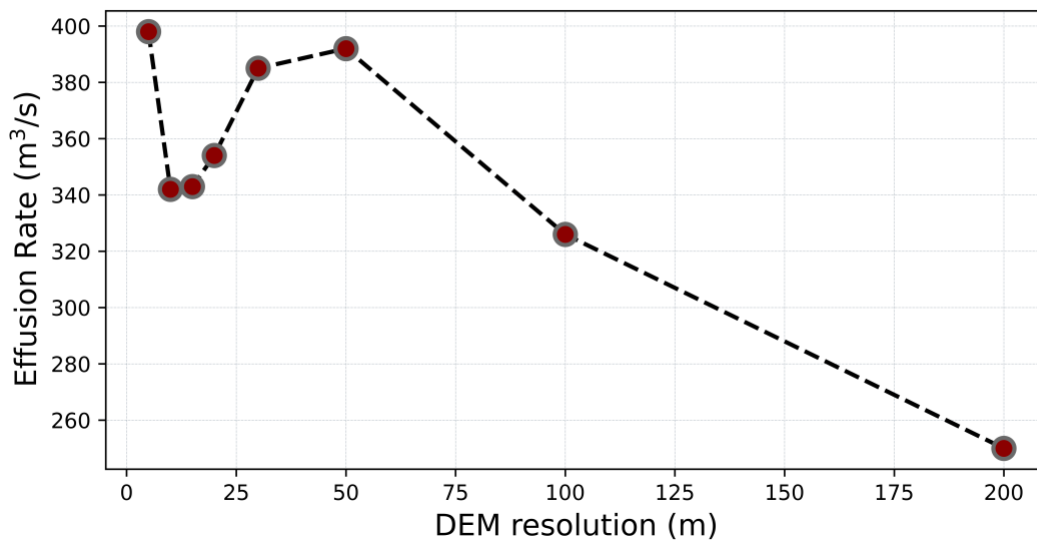


363

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 m³/s), 10 m (341 m³/s), 15 m (342 m³/s), 20 m (353 m³/s), 30 m (384 m³/s), 50 m (391 m³/s), 100 m (325 m³/s)
374 and 200 m (249 m³/s) model runs, and for the constant slope (11.3°) (248 m³/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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 391
 392

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.

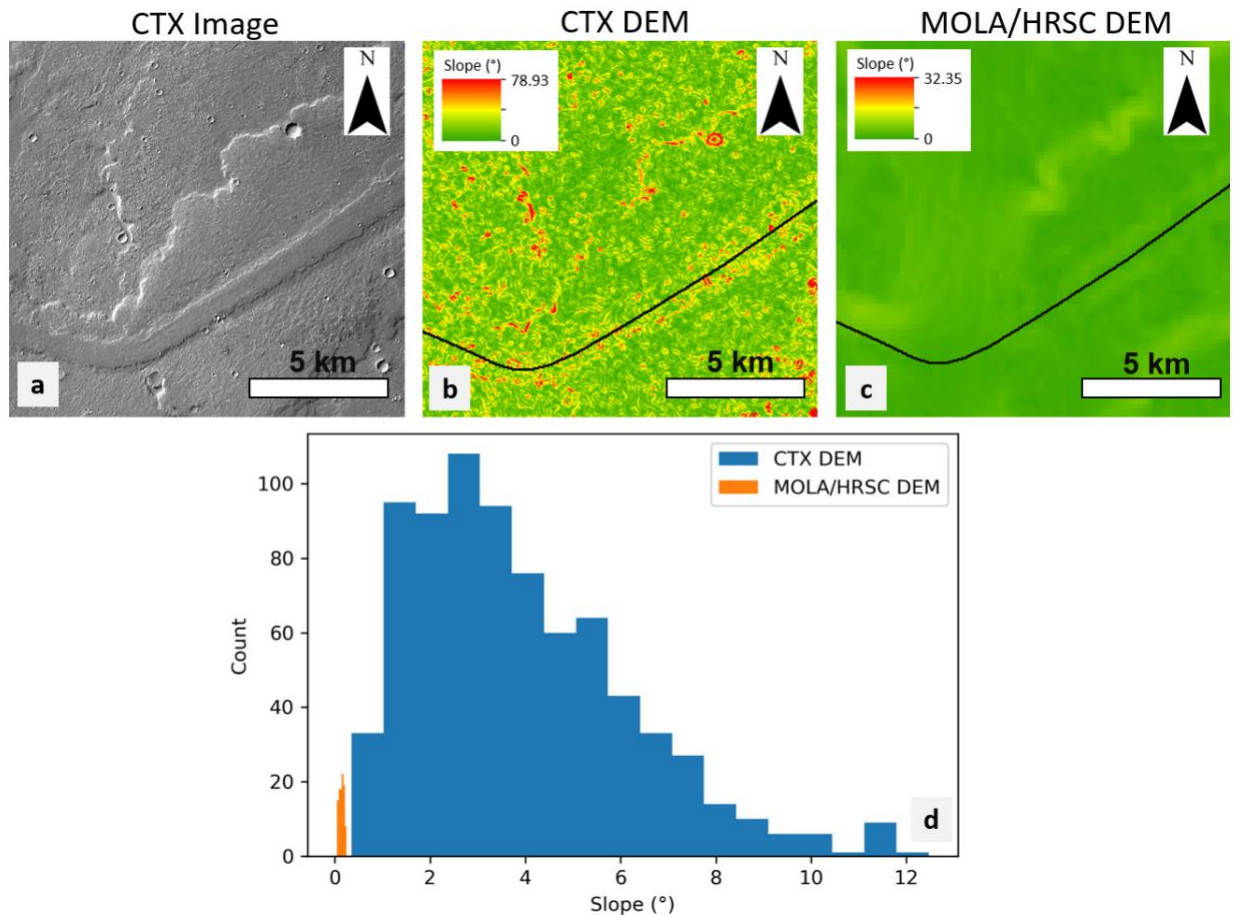
477 The impact of noise (and/or rapidly changing topography) in a DEM could be reduced
478 using the “slope smoothing” function in PyFLOWGO. This function interpolates a line over the
479 DEM based on a set number of points, defined by an individual before the model starts. It is
480 important to note 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-emplacement 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°) (Table 4). We compare the 5.6°
528 constant slope model run to those from the 11.3° constant slope results. Our results show that the
529 differences between a simulation on a constant slope of 5.6° or 11.3° 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 (~ 10 m to ~ 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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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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Models	Selection	References
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)
Heat Budget Modules	Selection	References
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)
Initial Input Parameters	Value	References
Step Size*	10 m	Harris and Rowland (2001)
Effusion Rate*	278 - 400 m ³ s ⁻¹	MIROVA (this study)
Width	30 m	Ramsey et al. (2019)
Depth	6.1 m	Ramsey et al. (2019)
Gravity	9.81 ms ⁻²	

Eruption Temperature	1355.15 K	Plechov et al. (2015)
Crystal Fraction	25%	Plechov et al. (2015)
Dense Rock Equivalent Density	2630 kg m ³	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 ⁻¹	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 ⁻³	Harris and Rowland (2001)
Air Specific Heat Capacity	1099 J kg ⁻¹ K ⁻¹	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 ⁻¹	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 ⁻¹	Harris and Rowland (2001)

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