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► **To cite this version:**

Magdalena Oryaëlle Chevrel, Andrew J.L. J.L. Harris. Monitoring Lava Flows. Modern Volcano Monitoring, inPress. hal-04160974

HAL Id: hal-04160974

<https://uca.hal.science/hal-04160974>

Submitted on 13 Jul 2023

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1 **Monitoring Lava Flows**

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10 **Table of content**

11	Abstract	3
12	1. Introduction	3
13	1.1. Preamble	3
14	1.2. Type of lava flows and associated hazard	4
15	1.2.1. Lava flow types	4
16	1.2.2. Associated hazard and risks	5
17	1.3. Spatial coverage, time scales and definitions	6
18	1.3.1. Spatial coverage, time scales	6
19	1.3.2. Definition of source terms	7
20	2. Parameters to be measured	9
21	2.1. Mapping spatial extent and dimensions	9
22	2.2. Lava temperature (core and surface)	12
23	2.2.1. Thermocouples (core temperature)	12
24	2.2.2. A first idea of the surface temperature	14
25	2.2.3. Thermal infrared thermometer (surface temperature)	14
26	2.2.4. Thermal camera (surface temperature)	17
27	2.2.5. A short note on temperature retrievals from satellite-based sensors	20
28	2.3. Lava surface velocity	21
29	2.4. Sampling	21
30	2.5. Lava flow morphology / typology	23
31	3. First Order derivatives	25
32	3.1. Building DEMs and volume estimation	25
33	3.2. Velocity profile and spreading rate	26
34	3.2.1. Velocity profile	26
35	3.2.2. Spreading rate	28

36	3.3. Heat Loss and Cooling rate	29
37	3.4. Sample chemical composition and texture	30
38	4. Second order derivatives	31
39	4.1. Volumetric discharge rate (volume flux)	31
40	4.1.1. Instantaneous effusion rate from lava velocity	31
41	4.1.2. TADR from volume, heat budget and flow area	32
42	4.1.3. Conversion to length	33
43	4.2. Rheological properties	33
44	4.2.1. Rheology from lava sample characteristics	33
45	4.2.2. Rheology of a lava flow	34
46	5. Hazard product: Lava flow modelling	35
47	6. Lava flow monitoring: The future	37
48	7. References	39
49		
50		

51 **Abstract**

52
53 During a volcanic effusive crisis, the active lava flow(s) need(s) to be monitored to best
54 anticipate the possible affected area. A number of measurements needs to be made on site either
55 from ground or from the air and by satellite imagery. These measurements need to be made
56 through established protocols so derivative parameters can be calculated and extracted to
57 contribute to forecasting and reduce possible impacts of lava flows. Measurements and
58 reporting parameters include lava flow setting, morphological description, dimensions
59 (thickness, length, width, underlying slope), temperature (core and surface), and sampling. First
60 order derivative are topographical changes, estimation of emitted lava volume, lava flow
61 velocity (channel and front), cooling rate, petrology and lava flow type characteristics. Second
62 order derivative then involve volumetric discharge rate estimation, heat budget models and
63 rheological constrains. These are essential parameters that are part of the monitoring duties and
64 feed lava flow hazard projection and numerical modelling. In this chapter, we review the
65 measurements that are typically made and the subsequent derivatives that all contribute to
66 monitoring activities during an effusive event.

67

68 **1. Introduction**

69 **1.1. Preamble**

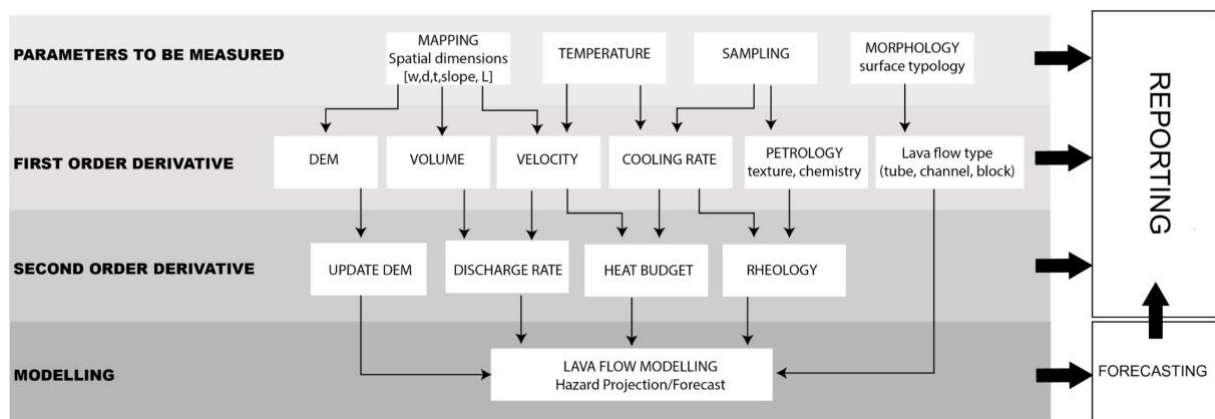
70 As a member of an agency, such as a volcano observatory, charged with monitoring and
71 surveillance during an effusive crisis you will have to focus on making observations and
72 measurements necessary to meet the reporting requirements of the agency. There is a lot to do,
73 and you will usually be understaffed and overworked. Worse, the lava flow may be in a remote
74 area, you will unlikely have helicopter support and will have to walk many kilometers over
75 difficult terrain carrying all that you need. To reach the lava, unlike when monitoring air fall or
76 a geophysical signal, you have to go to it and directly approach the source so as to collect
77 samples or measure the interior temperature. On return to the office you will need to work up
78 and process the data you have collected, enter them into a data base, use them for numerical
79 modelling and prepare a written or oral report. These can be daily, weekly and/or monthly for
80 both internal and external use, as well as press-release, depending on the eruption location,
81 threat and duration.

82 There will be a protocol in place for the measurements that need to be made as well as
83 the format and content of reporting duties. Thus, all you have time to carry out are the routine
84 measurements as there may be no time to set up experiments for scientific research paper. In

85 this chapter, we review the measurements that are typically made to contribute to monitoring
 86 activities during an effusive event. These fall into three linked categories which will provide
 87 information for reporting. They are (**Fig. 1**):

- 88 1. The measurements that need to be made in the field;
- 89 2. First order derivatives which are made from data collected in the field or via
 90 remote sensing; and
- 91 3. Second order derivatives that use the first order product to carry out calculations.

92 A fourth level of modelling can contribute to forecasting and will feed back into the reporting
 93 (**Fig. 1**). The objective of the chapter is to describe how to carry out these measurements,
 94 calculations and modelling.



95
 96 **Figure 1:** Flow chart tracking the series of measurements and calculation of derived parameters that are typically
 97 made to contribute to monitoring activities during an effusive event.

99 1.2. Type of lava flows and associated hazard

100 1.2.1. Lava flow types

101 Lava flows are gravity currents propagating downslope under their own weight, slowed by their
 102 own viscosity and core or crust yield strength, which increase with cooling and crystallization.
 103 Their velocity, morphology, shape, width and length mainly depend on the effusion rate (i.e.,
 104 lava volume flux leaving the vent), the total volume emitted, topography and the thermo-
 105 rheological properties of the lava. Typically, the higher the rate of effusion and the volume, the
 106 longer the flow, and the lower the viscosity and the steeper the slope, the faster the flow.
 107 Advance stops when the supply of lava stops (volume-limited flow) or when cooling is such
 108 that the flow viscosity reaches a limit that prevents further motion (cooling-limited flow).
 109 The different types of lava flows are distinguished according to their composition, their
 110 morphology and their dynamics. Silica-poor lava flows (e.g., basaltic flows) are the hottest and

111 most fluid and can reach speeds of several km/h. Flow types are generally pāhoehoe, with a
112 smooth, coherent surface, and ‘a‘ā with a surface comprising a layer of clinker. Flows can be
113 channel or tube-fed, and can extend from a few to a few tens of kilometers. The formation of
114 tubes insulates the lava, reducing cooling rates and allowing the lava to extend greater distances,
115 where the longest lava tubes on Earth are more than 100 km long. More siliceous compositions
116 (andesite, dacite and rhyolite) are colder, highly viscous and their advance rate does not exceed
117 a few m/day. Their thickness is generally several tens of meters and their surface is often made
118 of angular blocks which can reach sizes of the order of a meter. Silicic lava flows can, though,
119 extend several kilometers, where flow front collapses can depressurize the flow interior to feed
120 pyroclastic flows of the block-and-ash type.

121

122 **1.2.2. Associated hazard and risks**

123 Lava flows represent one of the most frequent volcanic hazards on a global scale, however, they
124 rarely pose a safety risk to populations (Harris, 2015). Deadliest examples are those of the
125 Nyiragongo eruptions in the Democratic Republic of Congo (DRC) in 1977, 2002 and 2021.
126 These eruptions involved extremely fluid lava flows that moved to high velocities and reached
127 the city of Goma, 10 km to the south of the source vents, in just a few hours. These eruptions
128 each accounted for several tens of fatalities, from direct contact with lava, collapse of buildings,
129 and explosion of fuel tanks.

130 Even if these human losses due to lava flows are minimal compared to those of other volcanic
131 phenomena (they represent only 0.2% of the total deaths linked to volcanic hazards – (Harris,
132 2015)), they nonetheless destroy, burn and bury all infrastructure and vegetation in their path,
133 having lasting effects on local economies. The 2018 effusive eruption of Kīlauea in Hawaii
134 which lasted only 4 months (Neal et al. 2019) caused no casualties, but destroyed more than
135 700 houses and caused losses estimated at over \$800 million (Meredith et al. 2022). The 2021-
136 2022 eruption of La Palma (Canaries), buried ca. 1200 ha of land including agricultural land
137 and villages, destroying more than 1600 houses and involved displacement of more than 7000
138 people (Cabildo de la Palma: <https://riesgovolcanico-lapalma.hub.arcgis.com>).

139 When the lava comes into contact with water, the instantaneous thermal expansion of the water
140 causes littoral explosions, while persistent entry of lava into the ocean can build benches.
141 Benches are unstable, and often collapse so that ocean entry sites are extremely dangerous for
142 visitors. Vaporization of sea water at ocean entries also feeds persistent gas plumes named
143 lazes. Lazes and gas plumes associated with emission of lava at the vent, that could create a
144 volcanic smog (called vog), are loaded with fine particles, as well as harmful gases and aerosols

145 (HCl, CO₂, NO₂ and SO₂; (Hansell and Oppenheimer 2004). At source, they are also extremely
146 hot so that inhalation of the gas can be fatal due to extreme lung damage. Wet and dry acid
147 deposition can also damage infrastructure, through corrosion, as well as crops, forest and habitat
148 (Blong 1984; Schmidt 2015).

149 Flow front collapses at silicic lava flows can represent a significant hazard due to the associated
150 block and ash flow (Harris et al. 2002; Pallister et al. 2019). A flow front collapse at Santiaguito
151 (Guatemala) in 1922 resulted in thousands of casualties (Rose 1987). Locations immediately
152 down slope of an active silicic lava flow are thus extremely dangerous places to live and work.

153

154 **1.3. Spatial coverage, time scales and definitions**

155 **1.3.1. Spatial coverage, time scales**

156 Lava flow monitoring can be carried out from three locations: ground, air and space. Each
157 vantage point allows differing spatial coverage and time scales of change to be tracked,
158 increasing from local and rapid for the ground observer to synoptic and regular from satellites.
159 Ground-based surveys include rapid, real-time measurements usually focusing around the key
160 elements of the flow field: the vent, the channel or tube, and flow front. Ground-based surveys
161 are usually carried out on foot, and have increasingly been supported by surveys by unmanned
162 aerial systems or vehicles (UAS-UAV, see review from James et al., 2020). UAS surveys cover
163 a few m² to hundreds of m² and are performed from heights of typically ≤ 120 m. This allows
164 greater coverage, is faster than walking the edge of a flow unit, and allows access to zones that
165 are impossible to approach on foot. However, like measurements made on foot, UAS surveys
166 cover only a restricted time slot of a few minutes.

167 Airborne surveys are measurements performed from helicopter or any aircraft and cover area
168 of few km² to 10's km². These operations are performed from higher altitudes, between 300
169 and 1500 m, than via UAS and may be ran over several hours.

170 From space, satellite acquisition includes geostationary as well as polar orbiting satellites allow
171 synoptic coverage at wavebands spanning the ultraviolet, through the visible to the infrared and
172 radar. However, only the latter is not affected by cloud cover.

173 Effective monitoring of lava flows combines observations from each of these three locations.
174 Observations are complementary, and ideally data acquired from all three locations are
175 combined to ensure a full coverage of the advancing lava flow. For example, while the ground-
176 based observation can be used to ground truth a point satellite-sensor-derived image, the image
177 can be used to extend knowledge beyond the isolated points visited on the ground.

178

179

1.3.2. Definition of source terms

180 A key source term for lava flow monitoring is the volumetric discharge rate (in m^3/s). This has
181 three definitions depending on the temporal scale (**Fig. 2a**; see also [Harris et al., 2007](#)) over
182 which the flux is measured:

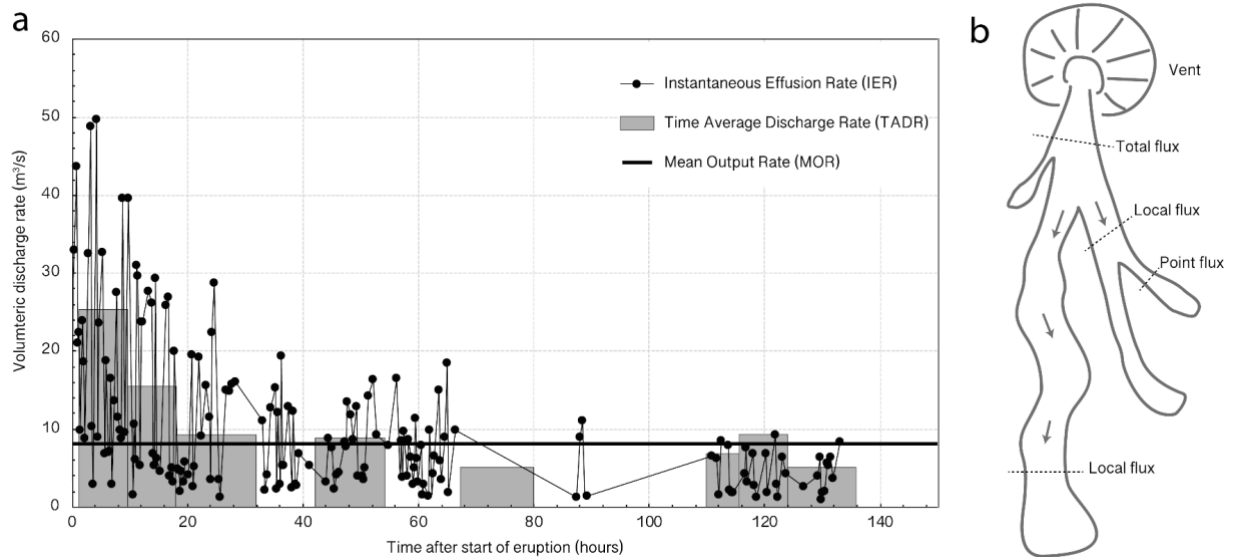
183 1. *Instantaneous effusion rate (IER)*: This is the volume flux measured at any single point
184 in time. It is primarily measured in the field at the master channel or tube using the
185 cross-sectional area (in m^2) of the flow multiplied by the flow velocity (in m/s).

186 2. *Time-averaged discharge rate (TADR)*: This is the volume flux averaged over some
187 period of time. This is the value typically obtained from satellite infrared data or from
188 DEM subtraction during the eruption. In this latter case, the volume of the lava flow
189 field on day n is subtracted from the volume of day $n+i$. The residual volume is then
190 divided by the time between the two measurements to give the time-averaged flux
191 during the i days between the two volume measurements.

192 3. *Mean output rate (MOR)*: This is the total volume erupted during the entire eruption, or
193 eruptive phase (usually obtained by DEM difference pre and post eruption), divided by
194 the duration of the eruption.

195 As we move through these measurements, the discharge rate curve becomes increasingly
196 smoothed and we lose the detail of peaks and troughs (**Fig. 2a**). The absolute value of the peak
197 fluxes also becomes reduced, which is something to bear in mind when applying predictive
198 models that relate volume flux to flow runout.

199 The flux measurement will also be influenced by where in the system the measurement is
200 made (**Fig. 2b**). The *total* volume flux can only be measured at the master channel or tube
201 exiting the vent and through which all of the volume passes. In a bifurcating system, after the
202 first branch a *local* flux will be measured which will be less than the total flux as the volume is
203 beginning to be partitioned between different distributaries. Finally, a *point* flux will be
204 obtained at a flow front lobe when many lobes are active (**Fig. 2b**).



205
206 **Figure 2:** Discharge rate definition over a) time and b) in space.
207

208 We need also to distinguish between bulk and dense rock equivalent (DRE) volume fluxes.
209 The bulk measurement includes all fluid and bubbles, as well as cracks in solid lava and spaces
210 between clasts in an 'ā'a flow. In contrast, the DRE only consider the solid phase and does not
211 include bubbles. This can be quite an issue as at the vent bubble contents can be very high and
212 usually decreases toward flow front. Worse, bubble contents can vary with time during an
213 eruption. Thus, volume flux measurements should always be taken, or reported, in tandem with
214 a measure of the bubble content. As a result of all inherent sources of error, the uncertainty on
215 a volume flux measurement is often at least 50 %. Thus, numbers (below) 100 should be
216 reported to the nearest decimal place; above 100 to the nearest round number; and above 500 to
217 the nearest ten, etc.

218 Finally, the spatial position of flow velocity also needs to be well defined because it changes
219 down system depending slope, flow depth and rheology. Fundamentally, channel velocities
220 measured at any given point down system will be higher than the flow front advance rate, with
221 the latter measurement being fundamental for hazard assessment. Velocity will also vary
222 between the flow surface and base, and from bank to bank. Velocity across and within a channel
223 may present a gradient from zero at the sides and base to a maximum at the channel center and
224 surface. Thus, the spatial point at which the velocity measurement is made must be noted so
225 that it can be related to the maximum and mean velocity (see later), both in terms of position
226 across a channel, as well as down system.

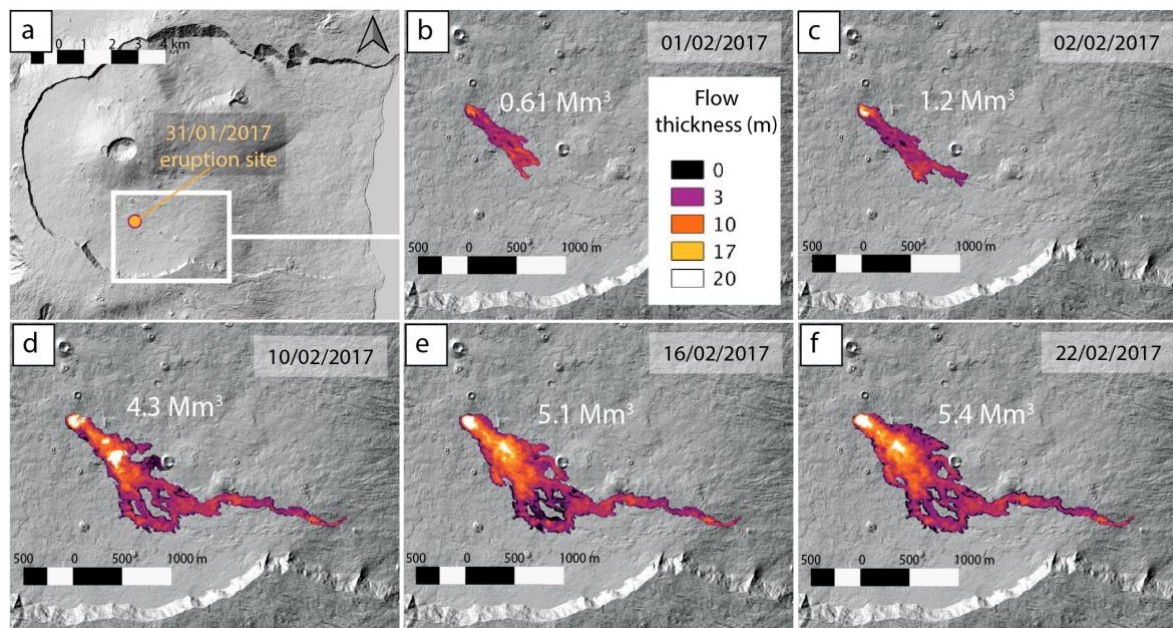
227 **2. Parameters to be measured**

228 **2.1. Mapping spatial extent and dimensions**

229 Monitoring lava flows starts with acquiring the geographic coordinates of the eruption site. This
230 requires location on the Global Navigation Satellite System (GNSS) for the ends of the eruptive
231 fissures and thus also their length and strike, or point a location of single vent. Then, the flow
232 front position is next most important point to locate, and hence the flow length at time t . The
233 evolving length should then be tracked. Ideally flow outline and thickness should be obtained
234 too, and terrain and topography updated so as to allow the trajectory of any new flow to be
235 assessed. As the eruption goes on, repeat surveys ensure an effective monitoring.

236 Fissure or vent location can be obtained by hiking to the eruption site with handheld GNSS.
237 However, lava flows often occur in remote places, and it is often difficult, or impossible to get
238 close to the vent (in particular due to the ongoing eruption). In this case, airborne survey is the
239 most efficient manner to get rapidly to the eruption site and acquire the required coordinates
240 with a handheld GNSS. Since the 1980's, reconnaissance overflights have been mainly operated
241 using helicopters. Satellite images are also consulted to detect the associate hotspot and may
242 rapidly provide localization of the lava flow extent although at low resolution (1–4 km/pixel),
243 or more precisely if there is an overpass by a satellite with higher (30–60 m/pixel) spatial
244 resolution ([see chapter Monitoring Heat](#)). Final options for an approximate location of the
245 eruptive vent are localization of the tremor sources and use of webcam networks.

246 During an eruption, the extent of an expanding flow can be tracked in multiple ways, depending
247 on the flow setting and dimensions, and on observatory funds where some techniques are
248 cheaper than others. Ground-based mapping can either be done by walking the flow limits with
249 handheld differential GNSS, using webcam networks or by UAS. UAS-derived photographs
250 can be stitched together to construct georeferenced orthomosaics from which the outline of the
251 flow can be drawn. Additionally, these photographs can be used for photogrammetry applying
252 structure from motion (SfM) technique to reconstruct the 3D topography and produce an
253 updated DEM. Over the past decade, this technique has become common. As mentioned above,
254 UAS-based survey is ideal for areas of less than 1 km², to map parts of the flow, local features,
255 vent areas, breakouts and flow fronts. However, flow-field-wide orthomosaics and DEMs
256 exceeding a few km squares need airborne surveys to be carried out by ultralight aircrafts or
257 helicopters (**Fig. 3**).

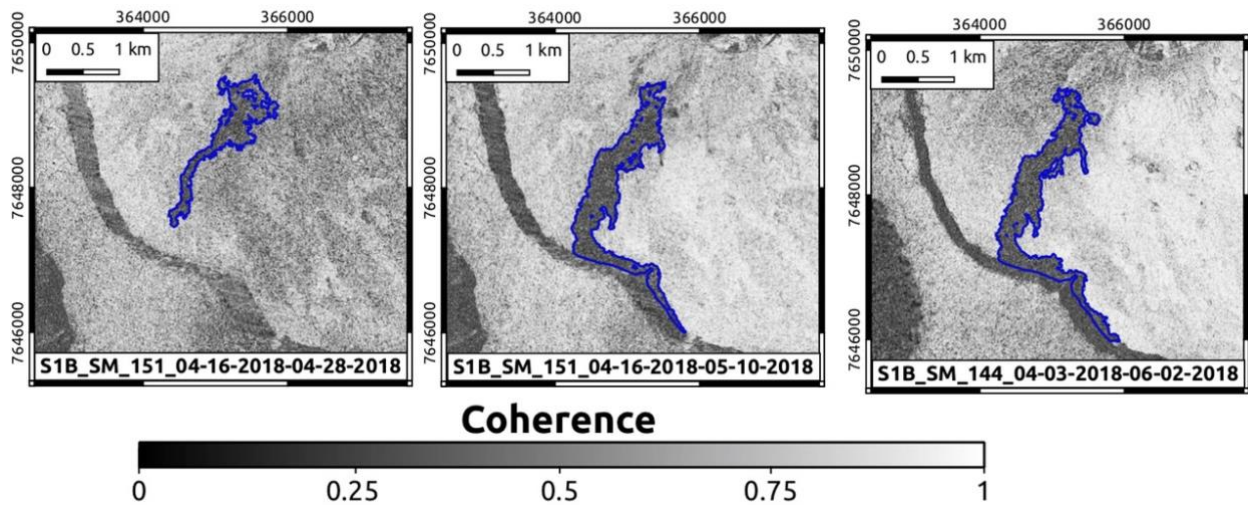


258
259 **Figure 3:** a) Location of the eruptive site of the 31 January – 27 February 2017 eruption of Piton de la
260 Fournaise volcano, La Réunion. b-f) evolution of the flow area, thickness (in m) and corresponding volume
261 (in Mm³) has obtained from repeat photogrammetric campaigns. Figure modified from Derrien (2019).
262

263 Moving to the airborne perspective, in Hawaii, helicopter overflights were carried out every
264 week, and sometimes daily, during the last recent effusive crises. Images acquired by a
265 handheld thermal camera from an airborne survey were used for mapping (Patrick et al. 2017).
266 Airborne Light Detection and Ranging (LiDAR) surveys can also be conducted to allow
267 millimeter topographic precision, though coverage is usually local and costs are high. At Piton
268 de la Fournaise (La Réunion), lava flow field evolution is usually mapped via ultralight aircraft
269 surveys, which have a more reasonable price (**Fig. 3**).

270 Satellite imagery is the most efficient way to track the evolution of lava flow fields and to map
271 entire lava flow fields with a single image. Satellite thermal and visible imagery (see chapter
272 Monitoring Heat) have spatial resolutions of 30 m (TM, ETM+, ALI, ASTER) to a meter
273 (Planet Labs) and to sub-meter (e.g., GeoEye and WorldView satellites). In case of cloud
274 coverage that hinders thermal and visible imagery to be exploited, satellites equipped with radar
275 sensors are used to monitor ground surface displacement through Interferometry Synthetic
276 Aperture Radar (InSAR) (e.g., ERS, Envisat, TerraSAR-X, Sentinel 1, ALOS-2). InSAR allows
277 the newly emplaced lava flow to be seen by comparing an image before and after lava flow
278 emplacement because it results in a highly incoherent surface (low coherence, **Fig. 4**).
279 Coherence maps can therefore be used to define the lava flow outline under any weather
280 conditions (Bato et al. 2016; Dieterich et al. 2012; Harris et al. 2019; Poland 2022).

281



282 **Figure. 4.** Evolution of the flow outlines and extension of the lava flow field as mapped from
283 InSAR coherence maps obtained via the Oi² platform (<https://opgc.uca.fr/volcanologie/oi2>)
284 during the April 2018 eruption of Piton de la Fournaise (modified from Harris et al., 2019).

285

286 Recent advances in mapping from satellite imagery is the Copernicus Emergency Management
287 Service (Copernicus Emergency Management Service 2022; <https://emergency.copernicus.eu/>)
288 that since 2012 has provided geospatial information through a no-cost mapping service in cases
289 of natural disasters within hours or days from the activation of support via their emergency
290 management activities for a natural disaster, including volcanic eruptions. In 2021, the
291 Copernicus Emergency Management Service provided a first precise lava flow map just two
292 days after the May 2021 Nyiragongo (RDC) eruption. The service was also extremely effective
293 in mapping the flow expansion during the 2021 eruption of La Palma (Spain) (Fig. 5).

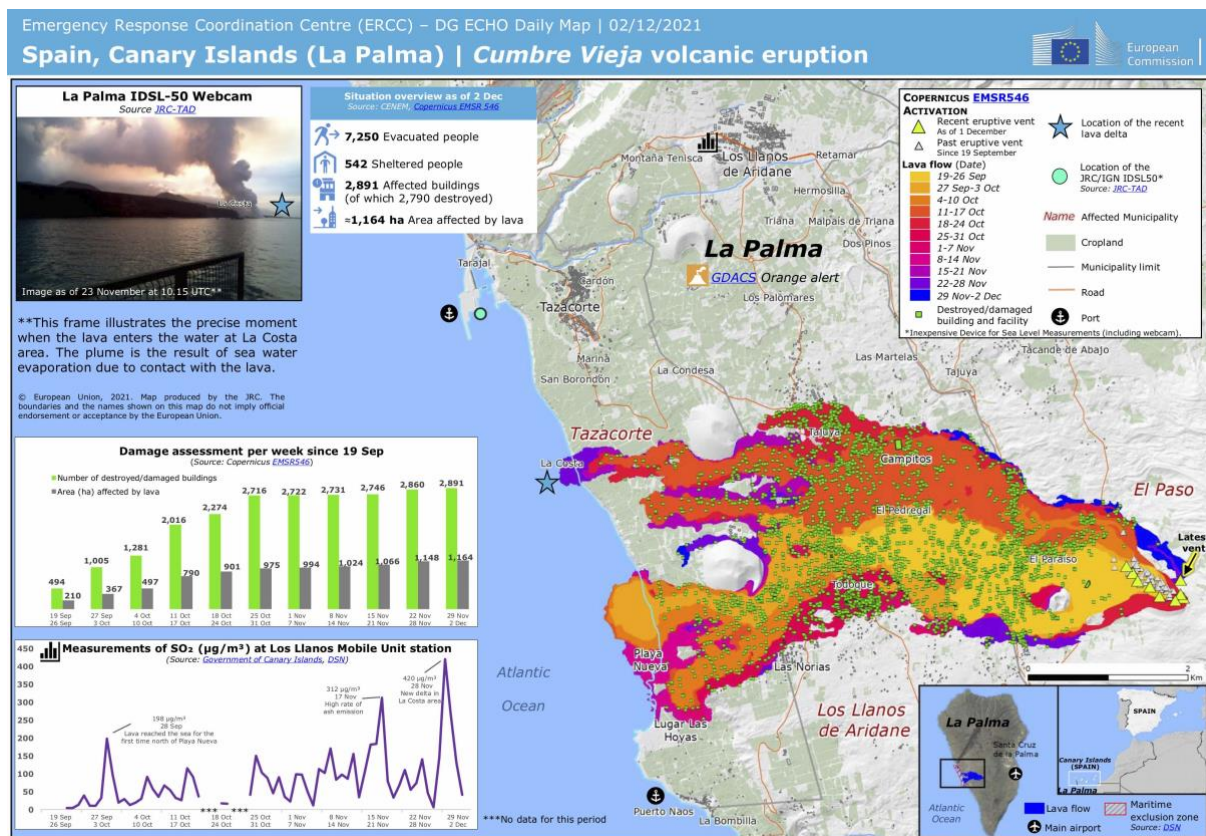


Figure 5: The Emergency Response Coordination Centre (ERCC) daily map for the La Palma (Spain) lava flow field for 2 December 2021. Copernicus Emergency Management Service (© European Union, 2012-2022)

2.2. Lava temperature

2.2.1. Thermocouples (core temperature)

The optical approach gives the surface temperature, so that a thermocouple is required to obtain the temperature of the flow interior. The interior temperature is typically hundreds of degrees centigrade hotter than the surface, and is the temperature we need to assess the eruption temperature, for down-flow cooling estimates and modelling. The thermocouple consists of two wires with a junction at the penetrating end, protected by a metal sheath and connected to a read out. The thermocouple is plunged into the lava and the temperature recorded. There are five concerns involved when making this measurement:

1. Probes are typically less than a meter long (e.g., **Fig. 6**). The measurement thus requires close approach to the active lava, which will have surface temperatures greater than 800 °C and interior up to more than 1100 °C. This can be at best uncomfortable and at worst painful, and means that all exposed flesh needs to be well covered to avoid radiation burns. Approach, if possible, should also be made up-wind, which is especially true at skylights where heated air blowing

314 out of the tube system can have temperatures >600 °C. A fire suit and helmet is
315 one solution (**Fig. 6**), and metal shield can be used to stop the radiation from the
316 lava. A longer probe could be another solution, but as they become longer than
317 1 m they tend to bend upon insertion, making penetration difficult. In that case,
318 the probe can be attached to a rigid steel pole of 1 to 2 meters that would ease
319 penetration into the lava flow although this adds to bulk and weight.

320 2. To protect the delicate wires, a metal sheath is used. To allow protection in the
321 extreme environment of a lava flow, as well as during penetration, this needs to
322 be at-least 3 mm thick. Such a thickness is also required to allow the probe to
323 penetrate the surface crust without it bending thus failing to pass through the
324 crust. However, the sheath needs time to heat up which means that the reading
325 will pick up over a period of one to two minutes. Once the reading is stable the
326 measurement is made. This means that the operator needs to stay in-situ for the
327 same amount of time.

328 3. Due to lava quenching to the probe, the sensor may not heat to the same
329 temperature as the surrounding probe. In such a case, the reading will begin to
330 decline without stabilizing and the probe should be withdrawn. It is thus best to
331 (i) have a pre-measurement idea of what the maximum temperature should be,
332 and (ii) continue to make the measurements until you have at-least three that are
333 the same and stable.

334 4. Care should be taken not to ground the probe on the cooler flow base, and to
335 ensure that the probe is well within the fluid interior and not in the cooler visco-
336 elastic crust. Finally, you should constantly stir to (i) avoid lava quenching to
337 the probe, and (ii) avoid the probe becoming entrapped in the constantly forming
338 surface crust.

339 5. If the flow is too large, close approach will not be possible due to the radiative
340 heat and/or the measurable zones maybe inaccessible in a sea of hot, recently
341 emplaced lava. Also, flow margins may be too unstable to allow safe approach
342 and/or flow surface in a channel may be too far below the levee rim. The easiest
343 targets to measure are small pāhoehoe lobes (**Fig. 6a**), or slowly moving
344 breakout through lava front or margin (**Fig. 6b**) but these are not always
345 available or accessible.

346 Remember, that upon completion of the measurement the metal sheath of the thermocouple will
347 remain extremely hot for several minutes, so on no account touch it. Also, after each

348 measurement, the lava attached to the probe will need to be quenched into water, so it will break
349 off, or need to be knocked off with a hammer.



350
351 **Figure 6:** Measuring lava temperature with K-type thermocouple that is 60 cm long and 3 mm thick sheath and
352 linked to a data logger at a) pāhoehoe lava lobe (Hawaii, 2016; photo: © USGS, M. Patrick) and b) at an 'a'a flow
353 front (Piton de la Fournaise, La Reunion, 2019) (photo: © L. Perrier).

354

355 **2.2.2. A first idea of the surface temperature**

356 The simplest way to have a ball-park estimation of lava surface temperature was visually
357 by matching surface color to the temperature associated with that color. This best done at night
358 and is the principle of the optical pyrometer. The pyrometer consists of a heated wire, the
359 temperature of which is adjusted to change the color at which it glows. When viewed against
360 active lava, the point at which the color of the wire is not distinguishable from the surface in
361 the background gives the temperature of that background. Alternatively, the assessment can be
362 made visually and a lookup table used to assess the temperature (Harris, 2013).

363

364 **2.2.3. Thermal infrared thermometer (surface temperature)**

365 The thermal infrared thermometer uses radiated heat to obtain surface temperature and
366 usually work with spectral radiant exitance in the 8–14 μm range. Note that, given the high
367 temperature range of active lava surfaces, a model which can measure the full range of expected
368 temperatures from ambient out to 1200 $^{\circ}\text{C}$ needs to be used. Being a remotely sensed
369 measurement, no close approach is necessary. However, the measurement is of a radiative
370 temperature (T_{rad}) and will not be the kinetic temperature. That is, most bodies are not perfectly
371 emitting blackbodies, so that spectral radiant exitance from a surface emitting at kinetic
372 temperature T_{kin} and wavelength λ will be reduced by a factor described by the emissivity (ϵ).
373 Spectral radiant exitance, $M(\lambda, T)$, varies as a function of wavelength and temperature (T), and
374 is given by the Planck function:

375
$$M(\lambda, T) = c_1 \lambda^{-5} \left[\exp \frac{c_2}{\lambda T} - 1 \right]^{-1} \quad (1)$$

376 Here c_1 and c_2 are constants with the values of $3.741 \times 10^{-16} \text{ W m}^2$ and $1.4393 \times 10^{-2} \text{ m K}$,
 377 respectively. Note that units of wavelength and temperature thus need to be in units of meters
 378 and Kelvin, and $M(\lambda, T)$ has the units W m^{-2} , i.e., $\text{J s}^{-1} \text{ m}^{-2}$.

379 Equation (1) applies to a black body, however for a body with an emissivity less than
 380 one the function becomes:

381
$$M(\lambda, T_{rad}) = \varepsilon M(\lambda, T_{kin}) = \varepsilon c_1 \lambda^{-5} \left[\exp \frac{c_2}{\lambda T} - 1 \right]^{-1} \quad (2)$$

382 Thus, to obtain the actual surface temperature, we need to convert the radiative temperature to
 383 a spectral exitance using Equation (1), divide the value of $M(\lambda, T)$ by emissivity, and convert
 384 the result back to temperature, which will now be a kinetic temperature. This can be done by
 385 inverting the Planck function:

386
$$T_{kin} = \frac{c_2}{\lambda \ln \left(\frac{\varepsilon c_1 \lambda^{-5}}{M(\lambda, T_{rad})} + 1 \right)} \quad (3)$$

387 Emissivity will vary with wavelength and surface type, where those for typical lava surface
 388 types are given in **Table 1**. There is debate over what the emissivity for a molten surface should
 389 be and evidences that emissivity changes with temperature and composition ([Biren et al. 2022](#);
 390 [Burgi et al. 2002](#)), but the values of **Table 1** are those typically used.

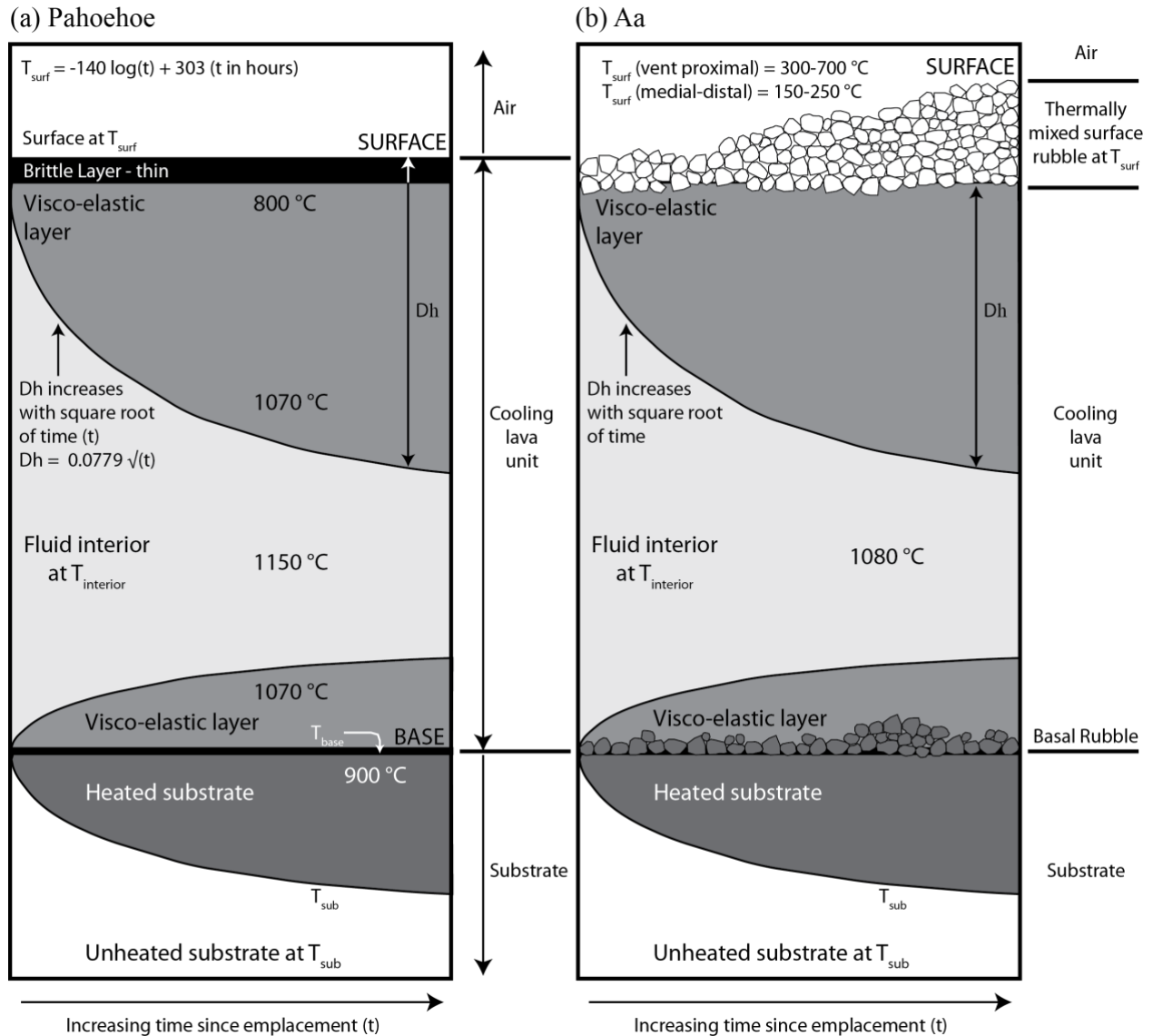
391

Location	Type	Emissivity
		8 – 14 μm
Kilauea	Basalt: pāhoehoe	0.901
Kilauea	Basalt: ‘ā’a	0.954
Kilauea	Basalt: pāhoehoe – ropey	0.943
Kilauea	Basalt: pāhoehoe – smooth, glassy	0.900
Kilauea	Basalt: pāhoehoe – vesicular	0.909
Etna	Basalt: pāhoehoe – slabby (spiny)	0.957
Etna	Basalt: ‘ā’a	0.971

392 **Table 1:** Band-averaged emissivity values for a range of lava flow types (from [Harris, 2013b](#)). Note that these are
 393 values for cold sample, and not molten material.

394

395 It must be recalled that the measured surface temperature will likely not be that of the
 396 interior. Rapid radiative cooling, for example means that the surface temperature can cool by
 397 50–300 °C within a second of exposure. On well-crustured, mature flows, the difference between
 398 the surface and interior temperature will be greater still (**Fig. 7**).



399
 400 **Figure 7:** Thermal layers in (a) a pāhoehoe lobe and (b) an ‘ā’a flow and their evolution with time (after
 401 [Harris, 2013](#)). For pāhoehoe, cooling and layer depth relations are from [\(Hon et al. 1994\)](#), as are the temperatures
 402 at the contact between each thermo-rheological layer and are for Hawaii. T_{base} is the basal contact temperature and
 403 is from [\(Keszthelyi 1995\)](#). For ‘ā’a surface temperatures for the proximal and medial-distal are for ‘ā’a channels
 404 at Piton de la Fournaise, La Réunion [\(Harris et al., 2019\)](#) and Lonquimay, Chile [\(Oppenheimer 1991\)](#), respectively;
 405 and interior temperature is for ‘ā’a erupted during Etna (Italy) 1991–1993 eruptions [\(Calvari et al. 1994\)](#). As
 406 apparent, while the thermocouple temperature must be made for the interior, and measurement can only be made
 407 when the crust is thin and immature, the surface temperature will not reflect the interior temperature. However,
 408 the surface temperature is what we need in order to estimate heat fluxes and cooling rates.

409

410

2.2.4. Thermal camera (surface temperature)

411

412

413

414

415

416

417

Thermal cameras tend to operate in the same waveband as thermal infrared thermometers (7.5–14 μm), but provide temperatures for an array of pixels providing a thermal image rather than a single spot temperature. Focal plane arrays tend to comprise 640×480 pixels, with each pixel having an instantaneous field of view (IFOV) of 0.65 mrad and the image having a standard total field of view (TFOV) of $24 \times 18^\circ$ (or $45 \times 12^\circ$ wide lens are also used). For a perfectly horizontal or vertical field of view the pixel or image diameter field of view (D_{FOV}) can be calculated using the simple trigonometric relation:

418

$$D_{\text{FOV}} = 2 [D \tan(\beta/2)] \quad (4)$$

419

420

421

Where D is the distance to the pixel or image center and β is the angular IFOV or TFOV. If the ground within the field of view is sloping, then D_{FOV} should be corrected for slope (α):

$$D_{\text{FOV}} (\text{slope corrected}) = D_{\text{FOV}} / \cos(\alpha) \quad (5)$$

422

423

424

For oblique viewing angles calculating the field of view dimensions requires the viewing angle (θ) to be taken into account (**Fig. 8**). For a viewing position h meters above the surface, this can be done using:

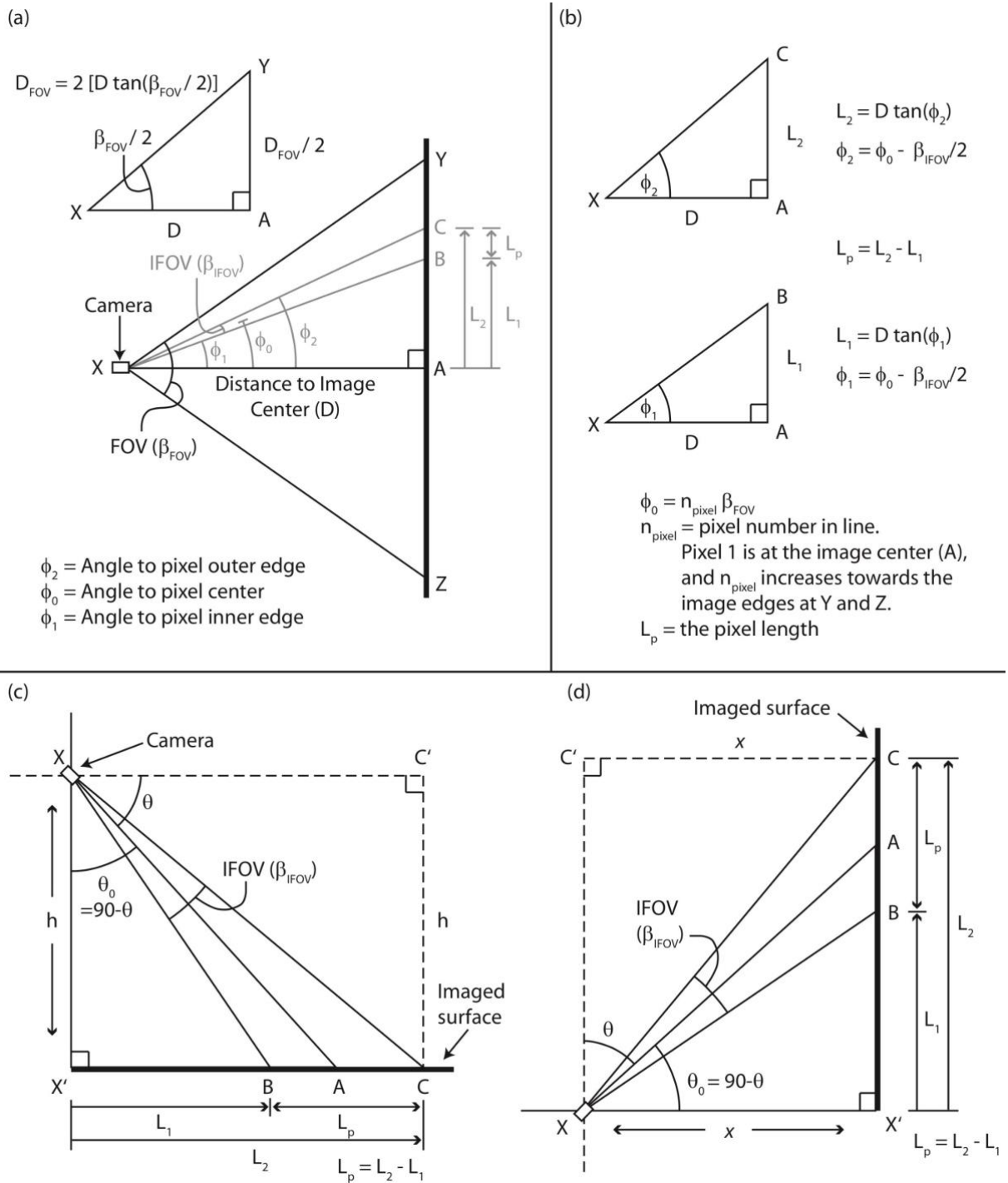
425

$$D_{\text{FOV}} = h \left[\tan \left(\theta + \frac{\beta}{2} \right) - \tan \left(\theta - \frac{\beta}{2} \right) \right] \quad (6)$$

426

427

For a viewing position x meters away from the surface (**Fig. 8**), calculation of D_{FOV} requires replacing h with x in Equation (6).



428

429 **Figure 8:** (a) Geometry for calculating pixel and image size for a flat surface viewed at right angles, with triangle
 430 required to calculate the total image field of view (FOV). Geometry for calculating the size of pixels within this
 431 FOV is in gray. (b) Triangles required to calculate the pixel size (L_{pixel}) using the detector instantaneous field of
 432 view (IFOV). (c) Geometry for calculating pixel size for a surface imaged by looking downwards, and (b) looking
 433 upwards. Modified from (Harris, 2013).

434

435 Pixel temperatures will be subject to the same emissivity correction as described for the
 436 thermal infrared camera, but also (for distances greater than 100 m) requires atmospheric
 437 correction. The signal received by the camera will be a mixture of the surface emitted radiant

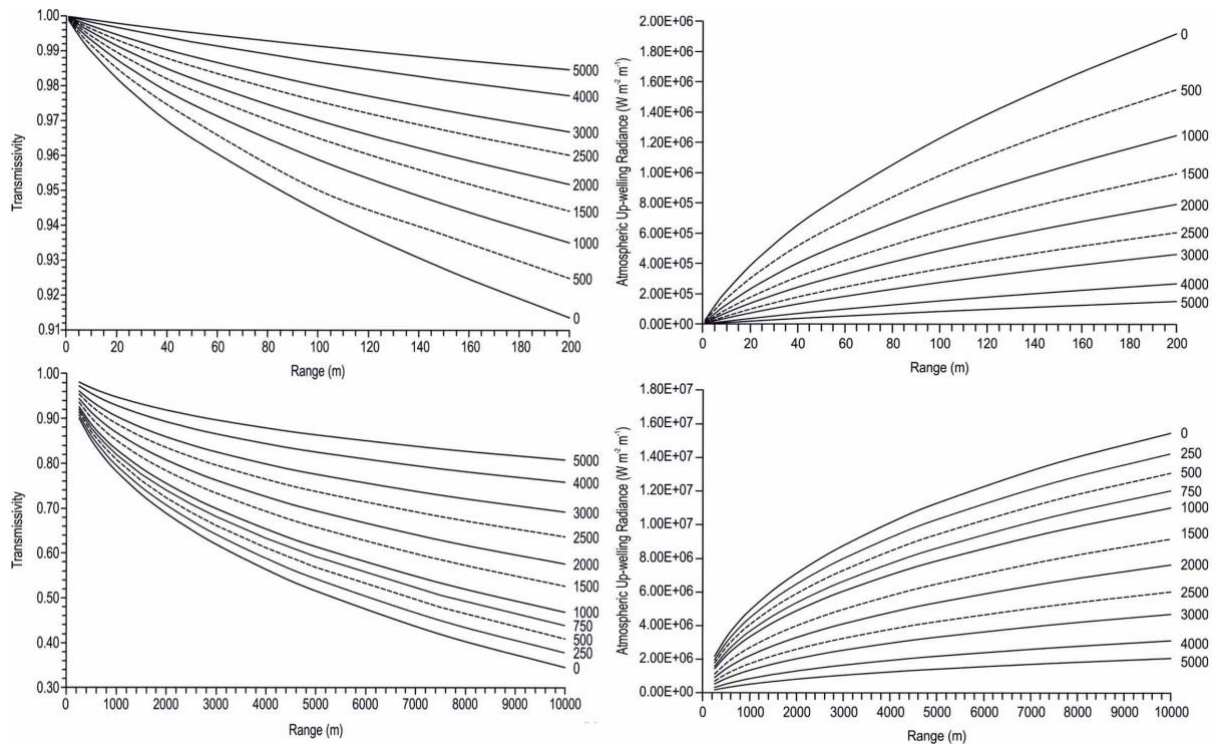
438 exitance, $\varepsilon M(\lambda, T)$, as well as that emitted by the atmosphere, termed atmospheric upwelling
 439 radiance (M_U). This will all be modified by atmospheric absorption (ψ) as described by
 440 transmissivity ($\tau = 1 - \psi$). Thus, the temperature recorded by the camera is a brightness
 441 temperature (T^*) given by (**Fig. 9**):

$$M(\lambda, T^*) = \tau(\lambda)\varepsilon(\lambda)M(\lambda, T_{kin}) + M_U(\lambda) \quad (7)$$

443 Atmospheric correction thus needs to take into account the emissive and absorptive properties
 444 of the atmosphere. This can be done by re-arranging Equation (7) so that,

$$M(\lambda, T_{kin}) = [M(\lambda, T^*) - M_U(\lambda)] / [\tau(\lambda)\varepsilon(\lambda)] \quad (8)$$

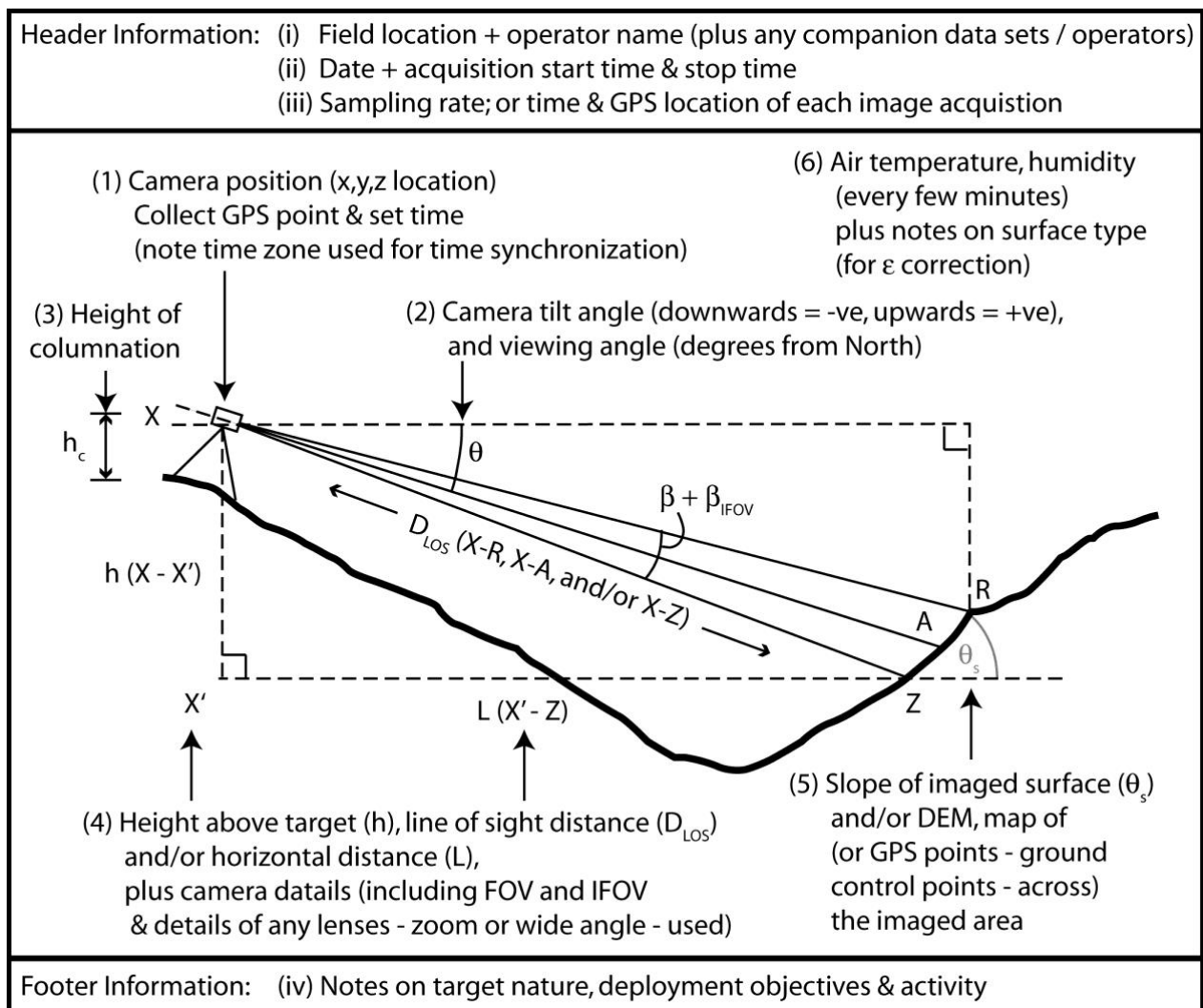
446 The atmospheric properties will vary with altitude and distance. The atmosphere becomes
 447 thinner with increased altitude and decreased path length, so that M_U and τ decrease
 448 accordingly. Typical correction values, as a function of altitude and distance are given in **Figure**
 449 **9**.



450
 451 **Figure 9:** Increase in atmospheric transmissivity (left graphs) and upwelling radiance (right graphs) for a
 452 horizontal line of sight (LOS) at a range of altitudes between 0 to 10 000 m (top graphs is a zoom over 0 to 200
 453 m). Values were obtained using MODTRAN with 1976 US Standard atmosphere and a CO₂ mixing ratio of 380
 454 ppm·v across the 7.5 μm to 13 μm waveband and are plotted from Tables in (Harris, 2013). Note that, after a
 455 distance of only 10 m transmissivity begins to become an issue, with almost 6 % of the signal being absorbed by
 456 a distance of 100 m at sea-level. At ambient temperatures, upwelling radiance is an issue, accounting for 20 % of
 457 the signal received by an infrared detector placed at a distance of 1 km from a surface at 0 °C. However, for higher
 458 temperatures the contribution is not so important, being 3 % for surfaces at 200 °C and 1 % for surfaces at 500 °C

459 (see [Harris, 2013](#)). Thus, at higher temperatures the correction for upwelling could be ignored for an approximation
 460 of kinetic temperature, but transmissivity cannot.

461
 462 The thermal camera measurement principles are the same whether the camera
 463 measurement is made from the ground or an aircraft. For airborne measurements the craft needs
 464 to fly sufficiently slowly so as to avoid image smearing, and doors must be off otherwise the
 465 temperature of the glass, which is 100 % opaque in the infrared, will be recorded. However, for
 466 both cases to solve the relations given above the acquisition conditions need to be recorded, as
 467 given in **Figure 10**.



468 **Figure 10:** Parameters to be recorded in a field note book at the time of image acquisition to allow adequate
 469 geometric and atmospheric correction of a thermal image. DEM: Digital Elevation Model, GPS: Global
 470 Positioning System, (I) FOV: (Instantaneous) Field of View, LOS: Line of sight. From [\[Harris, 2013\]](#).

471
 472 **2.2.5. A short note on temperature retrievals from satellite-based sensors.**
 473 *See [Chapter Monitoring Heat](#) for more details.*

474 Satellite-based sensors operating in the thermal infrared (typically 10–14 μm), as well as in the
475 mid (typically 3.5–4.0 μm) and shortwave (around 2.5 μm) infrared, can provide temperature
476 information. However, these are pixel-integrated temperatures meaning that the measurement
477 is an integration of all surfaces emitting within the pixel, which is typically 30 to 4000 m in
478 dimension. Mixture modelling can retrieve this surface structure, but are subject to the need to
479 oversimplify a complex problem. That is, a continuum of temperatures between background
480 (ambient) and eruptive (ca. 1000 $^{\circ}\text{C}$) need to be described by a mixture model with, typically,
481 two components (hot and cold). Such mixture models are also blighted by issues of saturation,
482 where gain settings were not designed for measurements over high temperature surfaces.
483 Instead, the maximum temperature that can be recorded in the thermal infrared is typically 60–
484 120 $^{\circ}\text{C}$, and in the shortwave around 250 $^{\circ}\text{C}$. Thus, over active lavas, all pixels are saturated
485 with the only information available being that the temperature is greater than the saturation.
486 Worse, the point spread function and optical reflection can cause intense smearing in the
487 imagery and halo effects. Thus, satellite-based sensors are not, currently, well suited for
488 retrieval of absolute temperature, but can be used for mapping as previously shown (also see
489 Chapter Monitoring Heat).

490

491 **2.3. Lava surface velocity**

492 Surface velocity can be obtained from a series of images (or video) collected in the field either
493 using a camera from the ground or flown on an UAS using particle velocimetry. This consists
494 of tracking a particle, or a set of particles, on scaled digital photos or thermal images acquired
495 at a known frequency. Velocity is then calculated from the distance travelled by the particle
496 and the time between two images. For this, pixel size must be well known and calculated
497 according to distance from the target and IFOV following the trigonometric relations as
498 described in the in section 2.2.4. In case of images or video footage acquired via UAS, a nadir
499 view is ideal as obtained with a stationary hover (to avoid image smearing) and a FOV covering
500 the whole channel including the stationary levée to anchor the velocity, with a few minutes of
501 recording being ideal (James et al. 2020). Particle velocimetry software, such as PIVlab
502 (Thielicke and Stamhuis 2014), can be used to analysis a few minutes of video. Surface velocity
503 may then be exploited to derive velocity profile over the full width and depth of the channelized
504 lava (see section 3), and integrated with depth to estimate IER and its local fluctuation (e.g.,
505 Bailey et al., 2006; Harris et al., 2007; 2019; Dieterich et al. 2021) .

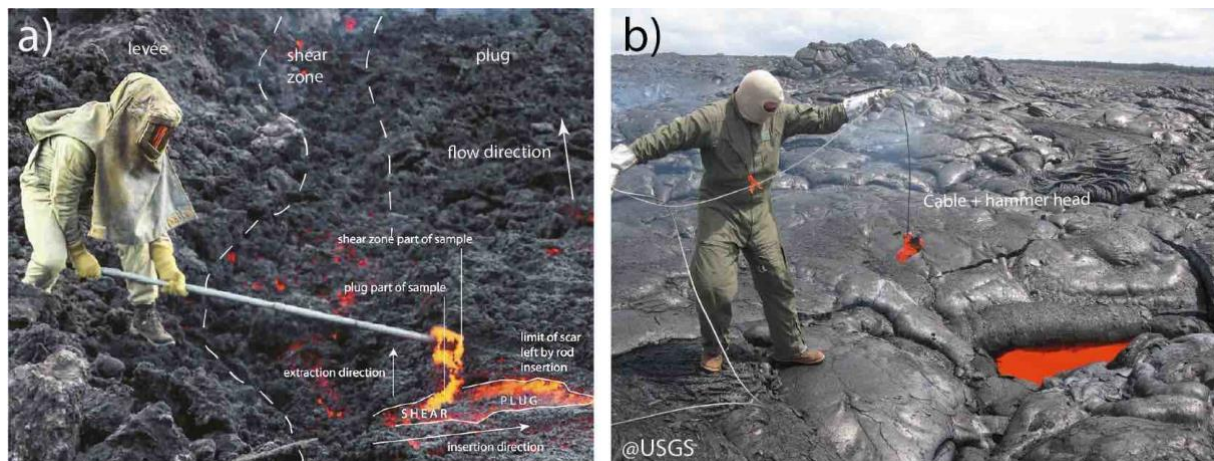
506

507 2.4. Sampling

508

509 Collecting rock samples during an eruption has the main objective of providing time-series for
510 the physio-chemical properties of the magma so as to provide insights into the eruption source
511 and allow initialization of lava flow simulation models (Harris et al., 2017, 2019). Following
512 the evolution of texture, chemical composition and petrology during an eruption has come to
513 be known as petrological monitoring and involves collection and distribution of data as soon as
514 samples have been received, prepared and analyzed (e.g., Pankhurst et al. 2022; Gansecki et al.
515 2019).

516 Effective petrological monitoring involves sampling volcanic deposits as soon as, and as often
517 as, possible during an ongoing eruption because the emitted lava may be buried the next day by
518 a new flow or fallout. The main problem with sampling an active lava flows is often, though,
519 not an easy task, as with thermocouple measurements, the access to the lava flow. Lava channels
520 and tubes are often inaccessible, and recently emplaced and hot lava usually needs to be crossed
521 to access the flow. Due to the heat, appropriate protective clothing, shield and sampling
522 instrument need to be selected (Fig. 11). Sampling the cooler lava flow front or lateral levées
523 is often easier, although falling blocks need to be watched for.



524

525 **Figure 11:** a) Sampling a lava channel from the stable levées (Piton de la Fournaise, La Réunion, 2015; (Harris et
526 al. 2020); a heat suit and 2-m stainless steel pole welded to a scoop is used; all the same the boots of the operator
527 began to flame just after this photograph was taken, sending toxic smoke up into the helmet. b) Sampling a lava
528 tube through a skylight (Kilauea, Hawaii, 2010) (courtesy of USGS Hawaiian Volcano Observatory); here a
529 hammer head on a cable is being used, and the operator has a fire-resistant balaclava and flight suit.

530

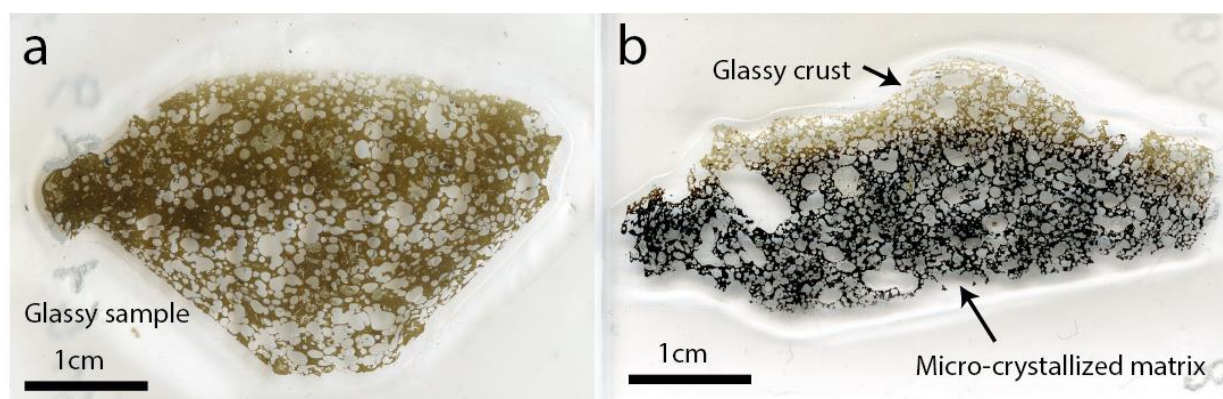
531 Samples of molten lava must be rapidly quenched in water, which will freeze-in the lava
532 physico-chemical and textural properties before any modifications due to a slower cooling

533 (micro-crystallization, element diffusivity, post-sampling vesiculation and expansion). Such a
534 sampling method ensures retrieval of the proportion of phases, glass composition and redox
535 state (**Fig. 12**). But requires the following to be carried to the sampling site:

- 536 1. A sampling instruments (hammer, chain, shovel ...);
- 537 2. Several liters of water for quenching;
- 538 3. A container to quench the sample in.

539 All of this is added equipment volume and weight to carry to the sampling site. Waiting for the
540 sample to cool in air will result in complete crystallization of the groundmass as well as re-
541 organization of the vesicle population. The resulting interstitial melt phase will therefore not be
542 glassy and representative of the flow state, but instead will display a microcrystalline
543 groundmass typical of slow, in-situ (post-emplacment) cooling (**Fig. 12**). This means that the
544 microlite and bubble contents present during flow will be overprinted and lost. If the surface
545 layer is naturally quenched so as to preserve a thin glassy rind, then this layer can be
546 representative of the molten lava (**Fig. 12b**). However, such layers are hard to find, and the
547 sample must be ‘in-place’; meaning it is not a rafted fragment that was quenched closer to the
548 vent and then transported down flow.

549



550

551 **Figure 12:** Polished thin sections (30 μm) of basaltic lava collected at Kilauea (Hawaii) in 2016. a) Lava sampled
552 with a stainless-steel tool and quenched in water. The sample is entirely glassy and preserved crystal and bubble
553 size distribution. b) Sample collected on a naturally cooled pāhoehoe lobe showing a glassy crust (brown color)
554 and a non-glassy micro-crystallized interior (black color).

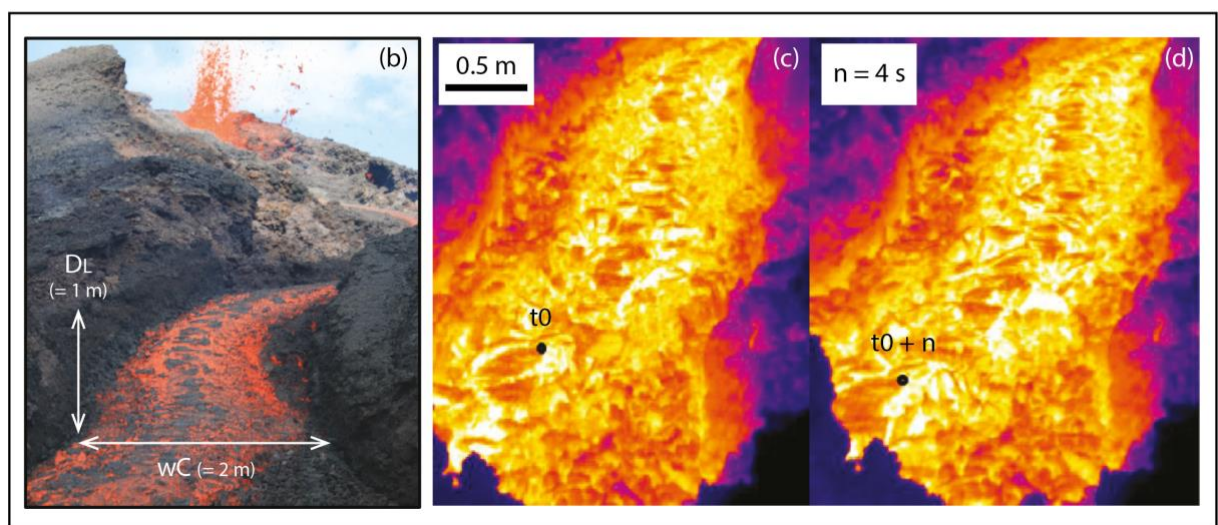
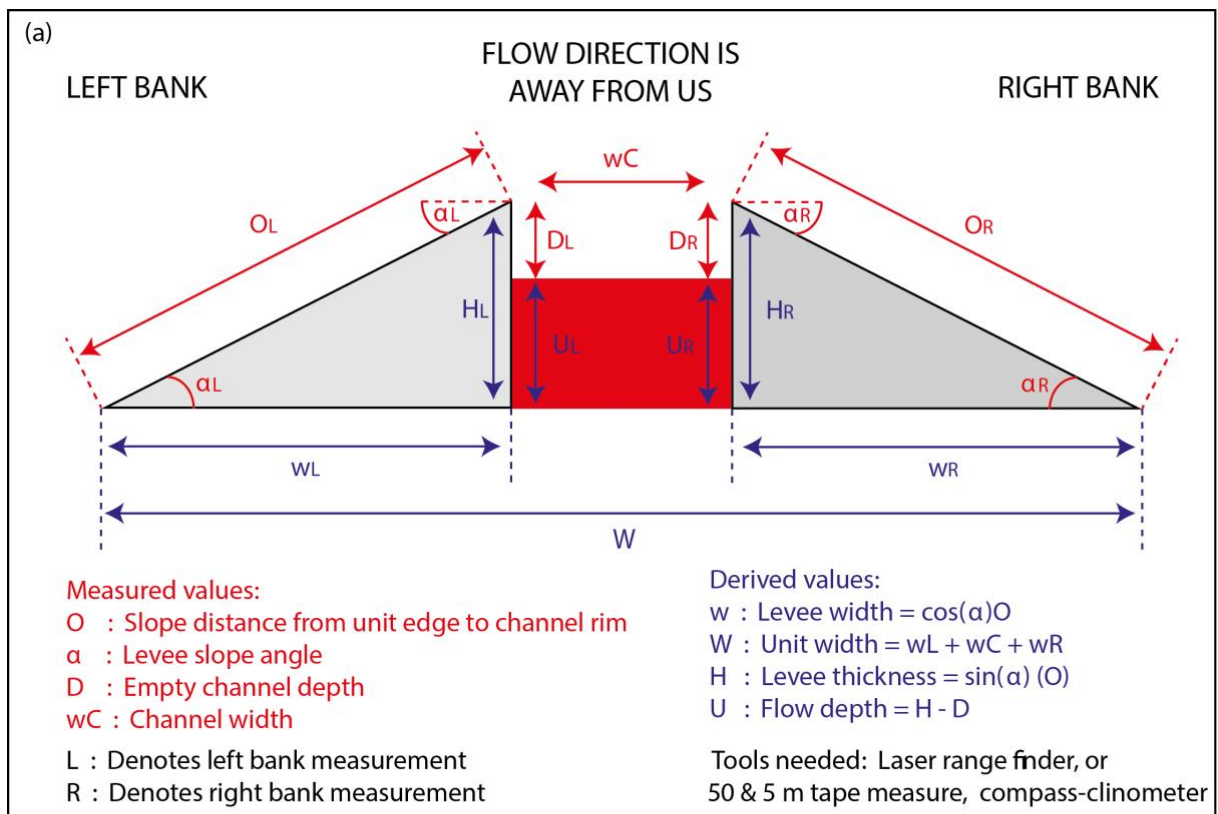
555

556 2.5. Lava flow morphology / typology

557 Lava flow type and associated morphology should be reported as it gives indication on the lava
558 flow dynamics. The field book entries for the day include descriptions and sketches of:

- 559 1. The lava flow distribution system (channel, tube-fed, dispersed ...);

- 560 2. Lava flow type (juvenile ‘ā’a, mature ‘ā’a, blocky, transitional, s-type / p-type
 561 pāhoehoe);
 562 3. Flow field form (ridged, ogives, ramped, ropey, hummocky, sheet ...);
 563 4. Any associated features (skylights, lava balls, lava boats, tumuli, shatter rings ...).
 564 Included for each entry should be estimates of approximate dimensions, especially length,
 565 width, height and depth of any feature, as well as underlying and new surface slope (**Fig. 13**).
 566 Of course, titles, orientations, scales, dates, times, GNSS points and tags for photos taken
 567 should be entered.



568

569 **Figure 13:** (a) Channel dimensions to be measured and calculated (modified from [Harris et al., 2022](#)), with (b)
570 example of below bank flow from a channel active at Piton de La Fournaise (La Réunion) on 1 August 2015
571 ([Harris et al., 2020](#)). Dimensions were obtained from scaled digital photos and thermal images, with velocities
572 also being obtained from the latter. Thermal images were collected at a frequency of 4 s. In (c) a piece of cooler
573 crust has been selected as close to the channel center as possible, and found again in the next image obtained, as
574 given in (d). The object has travelled 0.2 m in 4 s, so velocity is 0.05 m/s. Given the channel width and a square
575 channel that is as wide as it is deep gives an effusion rate of $[2 \times 2 \times (0.67 \times 0.05 \text{ m/s})] = 0.13 \text{ m}^3/\text{s}$. Of course,
576 the type of measurements and notes taken will depend on the feature under observation, where this case focuses
577 on a vent-proximal lava channel.
578

579 3. First Order derivatives

580 3.1. Building DEMs and volume estimation

581 To assess future lava flow paths, the constantly changing topography needs to be updated.
582 Commonly, DEMs are constructed by stereophotogrammetry that consists of merging
583 georeferenced images taken from different angles. Images are either acquired by UAS ([James
584 et al., 2020](#)), during flight surveys ([Derrien 2019](#)) or from satellite sensors operating in the
585 visible ([Bagnardi et al., 2016](#)). To build a DEM either multiple ground control points (GCPs)
586 are used to allow scaling and georeferencing (the larger the study area, the more GCPs are
587 required) or, as usually done during an eruption, the images are geotagged which allows a
588 decent georeferencing. Indeed, during an eruption establishing ground markers on foot is often
589 impossible. Data processing then involves applying SfM technique using software such as
590 Pix4D, Metashape or MicMac. Space agency services also produce DEMs from optical data
591 (e.g., ASTER GDEM), from multi-view optical systems of high spatial resolution (e.g., tri-
592 stereo Pleiades), or from radar data (e.g., SRTM and tanDEM-X; [Zink et al. 2014](#)).

593 LiDAR technique involves a pulsed laser that returns three-dimensional information of the
594 pointed surface. After filtering, the obtained point cloud generates topography of bare earth and
595 can be used in geographic information systems. LiDAR may either be mounted on an airplane
596 or on an UAS and provide very high-spatial resolution DEMs (of a few cm), although at high
597 cost ([Cashman et al. 2013](#)).

598 Reconstruction of new topography has two main goals. First, if topography is obtained rapidly
599 and regularly it can be used to update and improve numerical simulations of lava flow paths.
600 Second, the new DEM can be used to quantify the lava flow field volume. This is achieved by
601 calculating the difference between the topography in the new DEM and that of the pre-existing
602 DEM. Alternatively, first estimation of the lava volume can be calculated from the average
603 thickness of the flow and the flow area. Thickness can either be measured on the field from
604 flow edge height, from shadow in visible imagery if sun geometry can be calculated or from
605 DEM. For long lasting eruptions, regular volume estimation and DEM acquisition is part of the

606 monitoring routine as it serves to produce volume time series. This can then be compared with
607 volumes calculated by integrating TADR obtained via satellite over the period of time in
608 between two images to check on the effects of data loss due to cloud cover in the latter data set.
609

610 **3.2. Velocity profile and spreading rate**

611 In the field, as well as in thermal and visible video, flow velocities can be obtained by timing
612 the transit time of a marker, such as a piece of crust or the flow front, over a known distance
613 (section 2.2.4).

614 **3.2.1. Velocity profile**

615 For surface measurements, the velocity will increase from a minimum velocity at the channel
616 margins to a maximum (v_{\max}) towards the center. The reverse will reply with depth where the
617 velocity will decrease from a maximum at the surface to a minimum at the flow base (**Fig. 14**).
618 It is thus important to place the measurement in terms of where it is taken within this velocity
619 profile, where the safest approach is to take the maximum value at the channel center.

620 Considering a simple wide channel geometry and that lava rheological behavior is Newtonian,
621 the calculated velocity profile can then be used, through the [Jeffreys' \(1925\)](#) equation, to set
622 this measurement in spatial context. For a channel of width or depth h , velocity as a function of
623 distance from the bank or base (z) can be calculated from:

$$624 \quad v(z) = \frac{\rho g \sin(\alpha)(h^2 - z^2)}{4\eta} \quad (9)$$

625 Here, ρ is density, g is gravity, α is slope and η is viscosity, and the coordinate system is:

626 $z = 0$ at the channel center and/or surface (where $v(z) = v_{\max}$), and

627 $z = t$ at the channel margins and/or base (where $v(z) = 0$).

628 Other measurements taken at different points on the profile can be compared with the derived
629 profile, and if the fit is good the average velocity (v_{mean}) can be obtained. The average is
630 typically $0.67 \times v_{\max}$ (**Fig. 14a**) and can be calculated from,

$$631 \quad v_{\text{mean}} = \frac{\rho g \sin(\alpha)h^2}{8\eta} \quad (10)$$

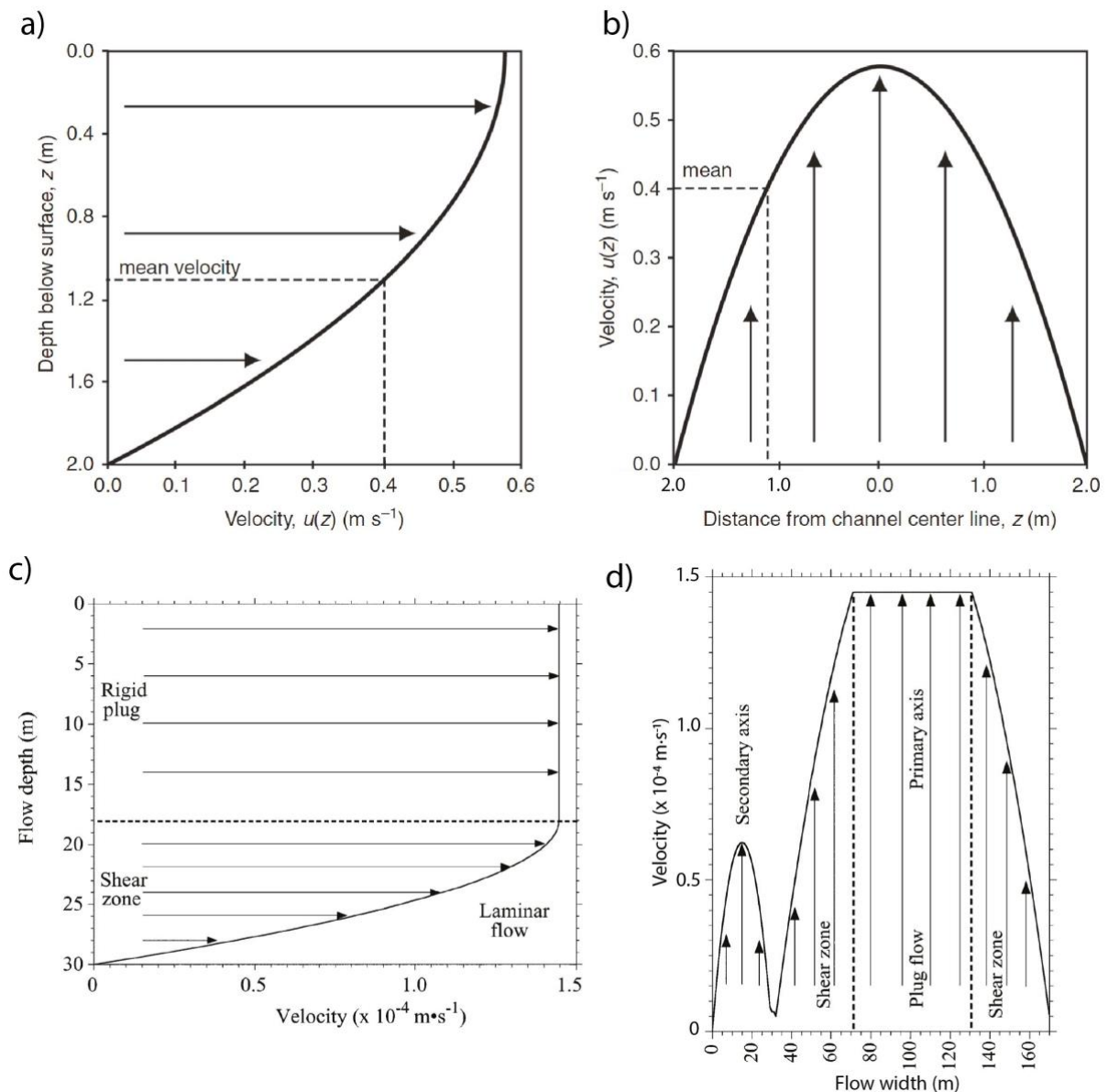
632 Thus, an important measurement to help constrain the velocity system is flow width and, if
633 possible, depth where the equations can also be used to constrain flow velocities in now inactive
634 channels or tubes. Another important piece of information is whether there is a plug, or zone of
635 non-shearing lava, at the channel center. Across this plug all flow is at the same velocity, so
636 that the velocity profile needs be calculated with the limits

637 $z = 0$ at the plug edge and/or base (where $v(z) = v_{\max}$), and

638 $z = t$ at the channel margins and/or base (where $v(z) = 0$),
 639 as done in (**Fig. 14c**). The mean velocity can also be re-calculated to take into account that a
 640 yield strength needs to be overcome to flow so we are dealing with a Bingham, rather than
 641 Newtonian, fluid using ([Moore 1987](#)):

$$642 \quad v_{mean} = \left(\frac{\rho g \sin(\alpha) h^2}{8\eta} \right) \left(1 - \frac{4}{3} \frac{\tau_0}{\tau} + \frac{1}{3} \left(\frac{\tau_0}{\tau} \right)^4 \right) \quad (11)$$

643 Here, τ_0 is the yield strength, which can be calculated from plug width (w_p), where $w = w_p \rho g \sin(\alpha)$,
 644 and τ is the basal shear stress [$\tau = \rho g h \sin(\alpha)$]. Thus, width of the plug is a final
 645 measurement that needs to be made.



646
 647 **Figure 14:** Vertical (a) and horizontal (b) velocity profiles for Newtonian flow (Etna, Italy, 2001, modified from
 648 [Bailey et al., 2006](#)), and for Bingham flow (c and d) (Santiaguito, Guatemala, 2001, modified from [Harris et al.,](#)
 649 [2002](#))

650

651

3.2.2. Spreading rate

652 If pixel dimension is known, feature width, length, perimeter and area (A) can be measured

653 from airborne thermal imagery. Changes from overflight to overflight can then be used to derive

654 advance velocities and spreading rates. Application of thermal chronometry can also be used to

655 estimate the time since emplacement (t). This is based on the characteristic cooling that the flow

656 surface experiences and which can be empirically described by (Hon et al. 1994):

$$657 \quad T_{\text{surf}} = -140 \log(t) + 303 \quad (12)$$

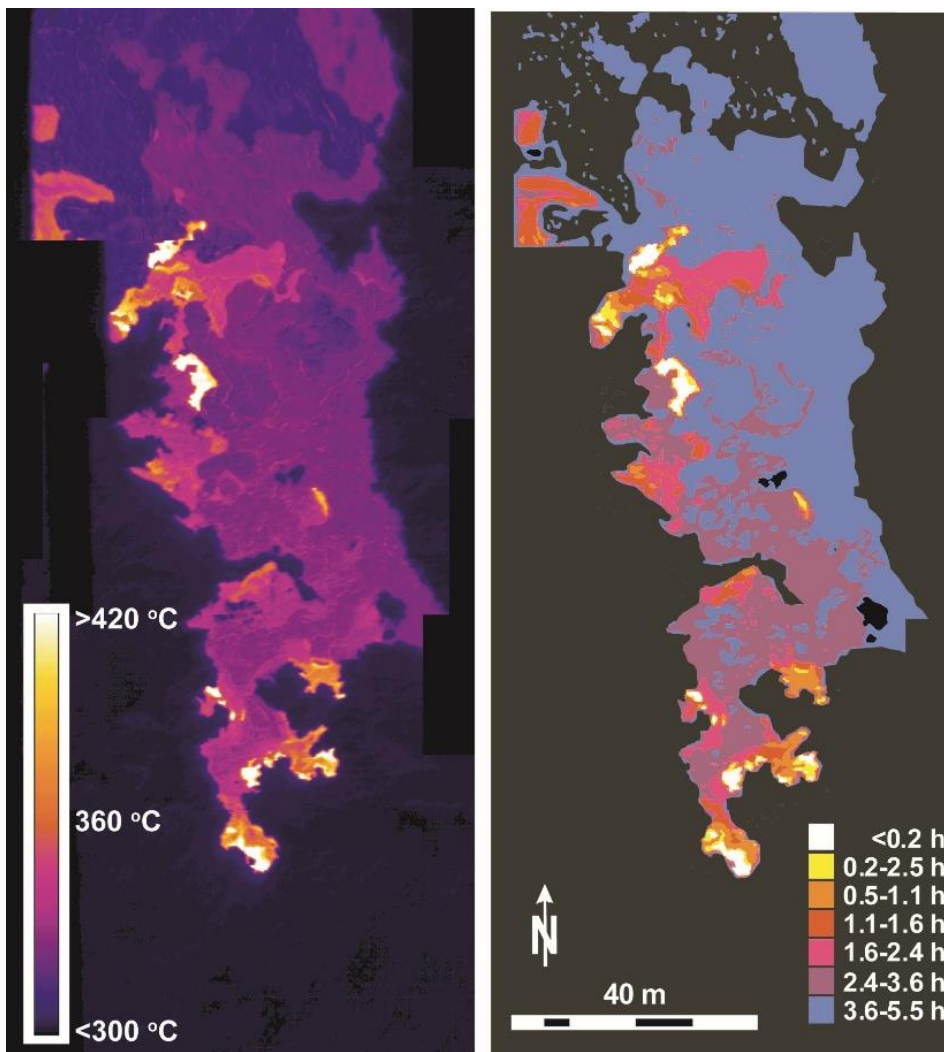
658 Where t is time since emplacement in hours. Now the surface temperature can be used to

659 derived age from:

$$660 \quad t = 146 \exp(-0.0164 T_{\text{surf}}) \quad (13)$$

661 Lava area emplaced over given periods can now be mapped, and the area (ΔA) and time (Δt)

662 differential used to obtain spreading rate ($\Delta A/\Delta t$) (Fig. 15).



663 **Figure 15:** Surface temperature and conversion to time since emplacement for a pāhoehoe flow active on Kilauea
664 (Hawaii) in 2006. This is a mosaic of five images acquired vertically from a helicopter hovering 200 m above the

665 surface. An area of 2400 m² was covered in the 1.1 hour prior to acquisition to give a spreading rate of 2181 m²
 666 per hour. From (Harris et al., 2007).

667 **3.3. Heat Loss and Cooling rate**

668 The two main heat losses from a lava flow surface are due to radiation (Q_{rad}) and convection
 669 (Q_{conv}). In terms of heat flux density (W m^{-2}) these can be written:

$$670 \quad Q_{\text{rad}} = \varepsilon \sigma (T_{\text{surf}}^4 - T_{\text{amb}}^4) \quad (14a)$$

$$671 \quad Q_{\text{conv}} = h_c (T_{\text{surf}} - T_{\text{amb}}) \quad (14a)$$

672 In which σ is the Stefan Boltzmann constant ($5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$), T_{surf} is the ambient
 673 temperature (in Kelvin) and h_c is the convective heat transfer coefficient ($35\text{--}50 \text{ W m}^{-2} \text{ K}^{-1}$). If
 674 flow depth (t) and mean velocity (v_{mean}) have also been measured, we can also estimate heat (in
 675 W m^{-2}) supplied from advection (Q_{adv}) and generated by crystallization (Q_{cryst}) from:

$$676 \quad Q_{\text{adv}} = v_{\text{mean}} t \rho c_p (dT/dx) \quad (15a)$$

$$677 \quad Q_{\text{cryst}} = v_{\text{mean}} t \rho C_L (d\phi/dT) (dT/dx) \quad (15b)$$

678 Here, c_p is lava specific heat capacity, dT/dx is the cooling rate per unit distance, C_L is latent
 679 heat of crystallization and $d\phi/dT$ is fraction crystallized per degree cooling. This can be solved
 680 if $d\phi/dT$ and/or dT/dx have been measured through sampling and/or thermocouple
 681 measurements at different down flow points. On the other hand, if $d\phi/dT$ can be assumed, then
 682 cooling rate can be estimated.

683 We have a system whereby heat is supplied by Q_{adv} and Q_{cryst} , and lost by Q_{rad} and Q_{conv} .
 684 In our case, the heat lost is not balanced by heat supplied, i.e.,

$$685 \quad Q_{\text{rad}} + Q_{\text{conv}} > Q_{\text{adv}} + Q_{\text{cryst}} \quad (16a)$$

686 As a result, the system is in disequilibrium and cools as a response. Given this imbalance, if we
 687 write the condition of 16a out in full and rearrange to isolate cooling, we arrive at:

$$688 \quad \frac{dT}{dx} = \frac{Q_{\text{rad}} + Q_{\text{conv}}}{v_{\text{mean}} t \rho (c_p + C_L \frac{d\phi}{dT})} \quad (16b)$$

689 This approximation can be used to estimate potential flow length (e.g. Harris and Rowland,
 690 2001). For example, if the cooling rate is 10 °C per kilometer, and the difference between
 691 eruption temperature and solidus or the point at which the flow can no longer move is 300 °C,
 692 then the maximum likely cooling-limited run out is 30 km. This back of the envelop calculation
 693 is very much an approximation, but can give a first order idea as to likely flow run out.

694

695 **3.4. Sample chemical composition and texture**

696 Analyzing rocks take time because samples need to be cut, powdered and polished thin sections
697 produced. For a long time, sample analyses were therefore only used for scientific research
698 purposes. However, in the last decades, analytical analyses have become faster and more
699 sophisticated petrological monitoring has now becoming more efficient (Re et al. 2021). Early
700 petrological monitoring of effusive activity at Kilauea by HVO during the 1980s and 1990s
701 contributed to assessing the state of the magmatic system feeding effusion (e.g. Thornber et al.,
702 2003). Analyzes also allowed tracking of eruption temperatures and cooling rates through
703 application of glass geothermometry (Helz et al. 2003).

704 More recently, the response system such as DYNVOLC (operated by OPGC,
705 <http://dx.doi.org/10.25519/DYNVOLC-Database>) allows providing petrological and textural
706 analyses during any and all eruptions of Piton de la Fournaise (La Réunion) within a few days
707 of sample collection to contribute to monitoring efforts by the observatory (Harris et al., 2017).
708 This response system was also applied recently during the submarine eruption offshore of
709 Mayotte where Berthod et al. (2021) performed detailed petrographic analyses of collected
710 submarine samples to contribute to monitoring efforts. Another recent example is Pankhurst et
711 al. (2022) who published a report on the rapid petrological monitoring response to the La Palma
712 eruption with combined field, petrographic and geochemical analyses conducted within 10 days
713 of sampling.

714 Petrological monitoring is therefore a key action to best constrain and understand an ongoing
715 effusive event (Gansecki et al. 2019). The first information to obtain is bulk rock chemistry,
716 including major and trace elements. Major element composition is essential to rock type
717 classification and for physical properties to be determined such as density and viscosity,
718 essential for lava flow modeling. Trace elements, although not used for lava flow modelling,
719 inform on the magma storage conditions and ascent path, and evolution of these conditions
720 through time. This is essential to track any changes of the magma properties and track whether
721 there is a new magma recharge, or an evolution toward more silica-rich or volatile-rich
722 composition that could force the eruption to transition toward explosive behavior.

723 Bulk chemistry is always accompanied with a detailed description of petrology and texture.
724 Petrological analyses give information on the magmatic plumbing system including magma
725 source depth, residence time, ascent rates, and ascent path. Magma reservoir depth (pressure)
726 and temperature are essentially obtained through thermobarometry (see Putirka, (2008) for
727 reviews). Crystal size distribution, crystal zonation, reaction rim features and presence of
728 xenoliths are also used to constrain ascent rate and path. Textural analysis includes

729 consideration of density, porosity, percentage of crystals and vesicles, as well as, crystal and
730 bubble size distributions (Guilbaud et al. 2007; Rust and Cashman 2011; Gurioli et al. 2008;
731 Harris et al. 2020; Robert et al. 2014; Chevrel et al. 2018). These inform also on the ascent
732 dynamics, especially for explosive phases because the texture is preserved at magma
733 fragmentation. In the case of monitoring lava flows, sample analysis provides information on
734 the lava temperature using geothermometry, cooling rate, crystallization history, phase (melt,
735 crystal, bubble) proportion, redox state and, hence, thermo-rheological properties. Care should
736 be taken when analyzing a sample given the sampling strategy (see section about sampling).

737 In essence, when monitoring a lava flow, it is essential to:

- 738 1. measure the bulk rock major element composition and volatile content,
- 739 2. estimate the temperature at the eruptive vent and then down flow to constrain cooling
740 rate,
- 741 3. measure density including dense rock equivalent and derive porosity,
- 742 4. estimate the vesicle content and size distribution, and
- 743 5. estimate the crystal (phenocrysts and microlite) content and size distribution.

744 To constrain cooling and crystallization rate down flow, one must look for the phases that
745 crystallized once the lava had been erupted, and not before, i.e., we need to use the microlites.
746 Usually the phenocryst content does not change down a lava flow, but the microlite content
747 often increases as the lava cools (Cashman et al. 1999).

748

749 **4. Second order derivatives**

750 **4.1. Volumetric discharge rate (volume flux)**

751 **4.1.1. Instantaneous effusion rate from lava velocity**

752 The effusion rate (E_r) of lava is given in m^3/s and defined as:

$$753 \quad E_r = v d w \quad (17)$$

754 In which v , d and w are flow velocity, depth and width, respectively. The “simplest” way to
755 calculate this is by measuring v , d and w in the field at an active channel or via a view of an
756 active stream in a tube through a skylight. If measured at a point location over a short time this
757 will provides the IER (see section 1.3.2).

758 Velocity can be estimated using the time needed for a surface marker, such as a
759 distinctive piece of crust, to move across a known distance (section 2.2.4). Because velocity
760 will vary across the channel (see **Fig. 14**), care must be taken to make this measurement at the
761 channel center where velocity will be at a maximum. This can then be converted into an average

762 velocity, which for a parabolic flow profile will be $v_{mean}=0.67v_{max}$, taking care to correct for the
763 presence of any plug. The average velocity needs to be used in Equation (17). Alternatively,
764 velocity can be obtained from sequences of scaled digital photographs or thermal images
765 (section 2.2.4; **Fig. 13c,d**). The same scaled images can be used to estimate channel width.
766 Width can also be obtained using a laser range finder, although these are often very unreliable
767 for low reflectance values, as is the case for basalt, as well as in heat haze or gassy conditions.
768 Radar guns have also been used for flow in tubes.

769 Channel depth is more difficult to estimate, but may be approximated from levee or flow
770 front height, taking into account a basal layer of crust and the level of flow below bank (**Fig.**
771 **13**). Alternatively, a channel shape (square, rectangular, or semicircular) can be assumed, and
772 the width used to approximate depth for the given shape. Another option is to wait for the
773 activity to stop, and directly measure the channel geometry directly with tape measure (**Fig.**
774 **10**), and then use this to estimate an average velocity using Equation (10). Attempts have been
775 made by plunging iron bars into the flow, but this is remarkably difficult to do due to difficulty
776 of access down into the channel for below bank flow, instability of levees, presence of
777 impenetrable surface crusts, and extreme high temperature conditions. Attempts have been
778 made to throw javelins into channel flow, but to no avail (Lipman and Banks 1987).

779

780 **4.1.2. TADR from volume, heat budget and flow area**

781 Volume flux can also be estimated by subtracting sequential volumes of the flow field and
782 dividing by the time between each volume estimate, which give a TADR. Volume estimates
783 are either obtained from geometry of the flow estimated and DEM subtraction (see section 3.1),
784 or via satellite-based hot spot detection systems (see [Monitoring heat chapter and \(Coppola et](#)
785 [al. 2010\)](#)) or via petrological method ([Verdurme et al. 2022](#)).

786 Another method to estimate TADR that has been well-used for satellite sensor-derived data in
787 the thermal infrared, such as AVHRR and MODIS, is the heat budget approach (See Chapter
788 Heat Monitoring). This assumes a direct relation between heat flux and TADR where heat flux
789 increases linearly with TADR, i.e.,

$$790 \quad \text{TADR} = x A (Q_{\text{rad}} + Q_{\text{conv}}) \quad (18)$$

791 where A is the flow area and x is a scaling factor that depends on local conditions such as slope,
792 degree of insulation and rheology. The basis of this relation is that the greater the TADR, the
793 greater the potential for the lava to spread (under any given slope, insulation and rheological
794 condition) and so attain a greater area. Hence, TADR scales with $A \times (Q_{\text{rad}} + Q_{\text{conv}})$. Means of

795 scaling this relationship are reviewed in [Harris \(2013\)](#). A simpler approach is to assume that
796 flow area increases with TADR, so that

$$797 \quad \text{TADR} = x A \quad (19)$$

798 In both cases (18 and 19) the solution becomes an empirical one which can be solved using
799 historical data for TADR and A , or theoretically ([See Chapter Heat Monitoring](#)).

800

801 **4.1.3. Conversion to length**

802 In a hazard sense, the most useful way to use effusion rate or TADR is to quickly assess how
803 far a lava may extend at a given supply rate. This follows on from [Walker \(1973\)](#) who showed
804 that lava flow length (L) tends to scale with TADR, so that

$$805 \quad \text{TADR} = x L \quad (20)$$

806 As with the relations of (18) and (19), this scaling needs to be carried as a function of flow
807 composition, rheology, slope and insulation. Again, this can be done using field data for TADR
808 and L or theoretically, if all flow characteristics (type, cooling rate, petrology and rheology)
809 and topography are approximately the same for all cases.

810

811 **4.2. Rheological properties**

812 Defining the rheological behavior of lava flows can be considered from two angles. The first is
813 the lava material itself that depends on its chemical and petrographically characteristics. The
814 second from the flow as viewed as a whole, considering the flow's dimension, velocity and
815 morphology.

816 **4.2.1. Rheology from lava sample characteristics**

817 Defining the lava rheology from sample characteristics must be performed on samples collected
818 along the flow, either collected after emplacement or from in situ sampling. Lavas are made up
819 of three phases. The molten phase, the solid phase (the crystals) and the gas phase (bubbles).
820 Most lavas are defined as suspensions, meaning that their crystal cargo is less than the random
821 maximum crystal packing. Although multiple models exist to constrain the viscosity of a
822 magmatic suspension (see review by [Kolzenburg et al. 2022](#)), a simple protocol can be applied
823 to define it. In this case, the viscosity of a lava is set as the apparent viscosity (η_{app}) of a mixture
824 of polydisperse particles (crystals and bubbles of various shapes and sizes) in a liquid phase
825 (that is a silicate melt):

$$826 \quad \eta_{app} = \eta_{melt} \cdot \eta_r = \eta_{melt}(T, X) \cdot \eta_{rc}(\phi_c, r, \dot{\gamma}) \cdot \eta_{rb}(\phi_b, Ca) \quad (21)$$

827 where the viscosity of the melt phase (η_{melt}) is Newtonian and depends on temperature (T) and
828 composition (X). The relative viscosity due to the presence of a crystalline phase (η_{rc}) depends
829 on the volumetric abundance of crystals (ϕ_c) and their aspect ratio (r), as well as the strain rate
830 ($\dot{\gamma}$). The relative viscosity due to the presence of bubbles (η_{rb}) depends on the volumetric
831 abundance of bubbles (ϕ_b) and their capillarity number (Ca), which depends on the melt
832 viscosity, bubble surface tension and strain rate.

833 The melt phase viscosity can be estimated from its composition and temperature using an
834 appropriate model or directly by measuring the temperature-viscosity relationship (Russell, et
835 al 2022). This is straightforward to obtain for lava samples where the glass phase has been well
836 preserved (*in situ* sampling and quenching or naturally quenched crust, Fig. 12) because the
837 chemical composition (major elements) can be easily measured using a microprobe and
838 temperature can be calculated from crystal-liquid geothermometer (Putirka 2008). In turn, if
839 the glass has not been preserved, but instead there is a microcrystalline groundmass, an
840 alternative method is to calculate the bulk groundmass composition obtained by subtracting the
841 crystal chemical composition using the average mass density of the crystals to the bulk chemical
842 composition (Ramírez-Uribe et al. 2021).

843 To estimate the effect of crystals and bubbles a detailed textural analysis is necessary. This
844 must include quantification of the particle sizes, shapes (aspect ratio) and abundances (volume
845 fraction). These characteristics are usually obtained from image analyses of thin sections and
846 microphotographs (obtained from microscope or electronic microscope, backscattered electron
847 images) and using software such as ImageJ (see for example Chevrel et al., 2018). There are a
848 large number of models to calculate the effect of crystals and bubbles on lava viscosity
849 considering either two phases (melt + crystals or bubbles) or three phases (melt + crystals +
850 bubbles). All details of the models found in the literature for magmatic suspensions are
851 presented in the review of Kolzenburg et al. 2022).

852 **4.2.2. Rheology of a lava flow**

853 The apparent rheological characteristics of a lava flow may be obtained from its dimension,
854 velocity and morphology. During an eruption, the viscosity of the flowing lava is estimated
855 from using either the velocities of lava flowing within channels or at flow fronts (e.g. (Harris
856 et al. 2004; deGraffenried et al. 2021; Harris et al. 2020). Under the simplification that lava
857 flows in a laminar fashion with no inflation following a Newtonian or Bingham behavior,
858 modification of the Jeffreys' (1925) equation and derivative equations (e.g., Eq. 9, 10, 11) can

859 be rearranged to estimate viscosity (Nichols 1939). Apparent lava flow yield strength can also
860 be obtained from its geometric parameters (Hulme, 1974).

861 A third alternative method to monitor the rheological properties of lava as it flows are direct *in*
862 *situ* viscosity measurement (see Chevrel et al. (2019) for review). However, this method is still
863 far from being possible as an operational tool as there is currently ongoing research on reliable
864 instrument that could be deployed easily in the field to perform such measurement (Chevrel et
865 al. 2023). In over 60 years (1948 to 2019), only eleven studies on field lava rheology
866 measurements have been published. Research is still ongoing and it would perhaps in the future
867 lead to a routine measurement.

868 **5. Hazard product: Lava flow modelling**

869 Monitoring lava flows is often accompanied by forecasting most probable flow paths and areas
870 of flooding. The simplest way of forecasting lava flow path is to define the line of steepest
871 descent (Kauahikaua 2007). In Hawaii, the lines of steepest descent lines have been defined for
872 use as a general guide in forecasting the initial lava flow path in future eruptions and to assist
873 emergency managers (Kauahikaua et al. 2017; Rowland et al. 2005). Long-term hazard maps
874 are a way of visualizing areas that will most likely be impacted (Chevrel et al. 2021; Favalli et
875 al. 2012; Del Negro et al. 2013). However, when an eruption starts, a short-term hazard map is
876 more appropriate because it focuses on the location of the ongoing eruption and represents a
877 given scenario for that is restricted to the downhill area from the vent or the fissure and the zone
878 at risk. These numerical models have been developed to include satellite-derived source terms
879 (mostly TADR) and to allow real-time lava flow emplacement simulations and projections
880 [Wright et al. 2008; Ganci et al. 2012; Harris et al. 2019]. This is routinely produced at highly
881 eruptive center like at Hawaii (Neal et al. 2019), Mt Etna in Italy (Vicari et al. 2011) and Piton
882 de la Fournaise, La Réunion (see below). Such maps are good communication tools between
883 scientists and stakeholders. Nonetheless to ensure a common understanding and to help in
884 building effective mitigation strategies they need to be designed based on the needs of the end
885 users (Chevrel et al. 2022). For this, engagement with end users (local authorities) is absolutely
886 necessary.

887 Lava flow forecasts usually involve provision of model-derived the

- 888 1. area that will potentially be covered
- 889 2. most likely path and
- 890 3. distances the flow could reach.

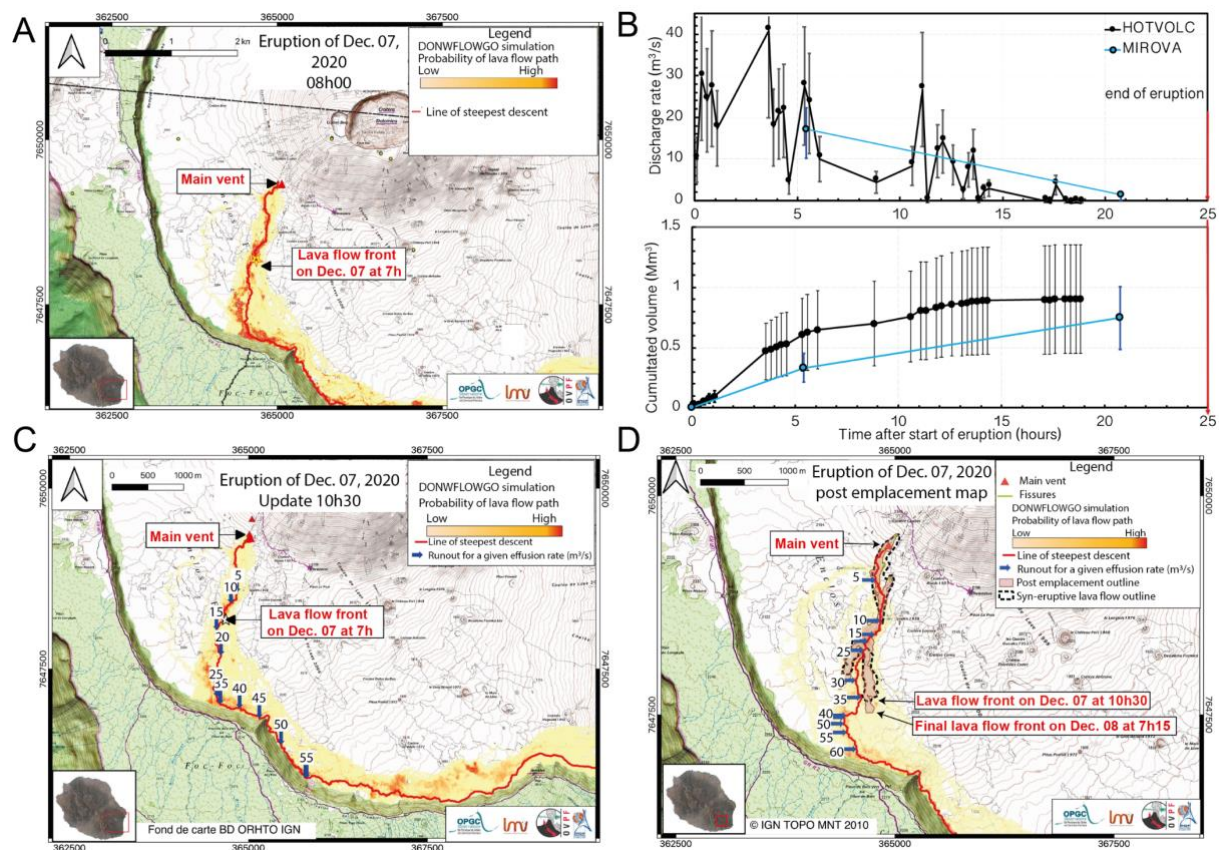
891 Output depends on topography, slope, volume flux at the vent, total volume of lava emitted
892 and/or the thermo-rheological properties of the lava. There are a number of numerical models
893 that can be used to build short-term hazard maps but they all need to be pre-initialized,
894 calibrated and validated (Harris et al. 2019; Vicari et al. 2011; Hyman et al. 2022). Pre-
895 initialization and calibration rely on collection of source term data collected, and validation
896 involves retrospective analyses on previous eruptions to best fit the model to reality.

897

898 *Exemple at Piton de la Fournaise (La Réunion)*

899 At Piton de la Fournaise (Le Réunion), since 2014 lava flow inundation areas have been forecast
900 and modeled through a satellite-data driven protocol (Chevrel et al. 2022; Harris et al. 2017;
901 2019). This protocol determines the flow probable inundation area and runout distances via the
902 DOWNFLOWGO numerical code that is a combination of DOWNFLOW, that provides the
903 path of steepest descent and probable lava flow inundation area (Favalli et al. 2005), and
904 FLOWGO that calculates the maximum cooling-limited runout of lava for a given effusion rate
905 (Harris and Rowland, 2001). To operate this protocol on an operational basis and routinely
906 during each eruption, a retrospective analysis of the April 2018 eruption was carried out (Harris
907 et al., 2019). As part of the protocol, an initial hazard map is shared within the first hours of
908 any eruption and, when needed, the map is updated as eruption conditions evolve (new vents
909 opening, extension of tubes, change in TADR) (**Fig. 16**). The maps provide a fundamental
910 communication tool between scientists and the authorities and have been developed through
911 iterative dialog with local civil protection. Figure 16 shows the delivered products during the
912 operational forecasting and monitoring response at Piton de la Fournaise for the December 2020
913 eruption. This includes a map of the eruption site where the line of steepest descent and the
914 probable lava flow paths is shown together with the location of the eruptive vent and of the
915 flow front. This short-term hazard map can be better constrained and updated through the
916 eruption as TADR is known (**Fig 16b, c**) and lava flow outline is obtained either from satellite
917 imagery or airborne photogrammetry (**Fig 16d**).

918



919
 920 **Figure 16:** Operational forecasting and monitoring response at Piton de la Fournaise (La Réunion) for the
 921 December 2020 eruption. a) First map produced on December 7, 2020 at 8h00 local time. b) TADR and cumulative
 922 volume since the start of the eruption obtained by the HOTVOLC (Gouhier et al. 2016) and MIROVA (Coppola
 923 et al. 2016) systems. c) Updated map produced on December 7, 2020 at 10h30 local time. d) Final map showing
 924 the probability of lava flow invasion, runoff distances as function of TADR and the final lava flow field outline
 925 obtained by airborne photogrammetry (post emplacement outline). The lava flow outline drawn from the Sentinel-
 926 2 image acquired on December 7 at 10h30 is also shown (syn-eruptive outline). This map was produced after the
 927 end of the eruption. Taken from Chevrel et al. (2022).
 928

929 6. Lava flow monitoring: The future

930 Technology and capabilities are constantly improving and evolving, as new techniques come
 931 on line, existing instrumentation becomes improved, and prices become lower. One could argue
 932 that for monitoring, in some areas, we do not need any further reductions in images spatial
 933 resolution. For example, for lava flow modelling a DEM spatial resolution of 10 m is more than
 934 adequate. However, what we do need is the best possible temporal resolution, so as to track
 935 lava flow spreading, topographic trends and update lava flow models with the most up-to-date
 936 topography. Given the time scales involved in emplacement of a rapidly extending lava flow a
 937 few minutes may be ideal, whereas for longer lasting events (years) a few weeks may be
 938 sufficient. The key is, to balance data loads with scenario, processing capabilities and need. In
 939 populated areas the need for regular and fast processing of data and product, as developed by
 940 Copernicus Emergency Management Service (e.g., Fig. 4), is more pressing than in remote and
 941 unpopulated areas.

942 Of course, lava is hot so one of the best means of tracking it is remotely and in the infrared.
943 However, no terrestrial sensor is currently perfect for tracking active lava. Fundamentally, gain
944 settings are usually too low so that saturation occurs at 60–120 °C, meaning all pixels over an
945 active lava flow are saturated. A pressing need is for at-least one waveband in the mid and/or
946 thermal infrared with a saturation of at-least 800 °C.

947 UAS, that has literally taken off over the last decade, represent a promising tool to monitor lava
948 flows but onboard thermal imaging payloads need to be developed for lava flow temperatures.
949 To track entire lava flow fields, advances in UAS portable thermal imaging technology needs
950 to be associated with improving capabilities for UAS to cover larger areas and for building
951 orthomosaics from potentially thousands of images covering tens of km².

952 Seismicity and infrasound have been shown, especially during the 2018 eruption of Kilauea,
953 capable of tracking the evolution of an effusive event, especially in terms of vent location and
954 effusive intensity (Patrick et al. 2019). Further advances would be to track flow advance, where
955 ground-based radar and/or InSAR could be used. This would be ideal for cloudy regions, but
956 frequencies of current InSAR acquisitions could be increased to a few hours or so. Then, there
957 are all sorts of possibilities we could think of, such as placing transmitting temperature and
958 position sensors on and/or within an advancing lava flow.

959 While fast and simple non-physical 1 and 2D models are systematically used forecast lava flow
960 paths and velocities and preparation of short term hazard maps (Harris et al. 2019; Chevrel et
961 al. 2022; Ganci et al. 2012; Herault et al. 2009; Neal et al. 2019), more sophisticated physics-
962 based models are still not well developed to produce a timely forecast (Dietterich et al. 2017).
963 Development of faster numerical models accompanied by automatization of satellite data
964 integration are needed so that such capabilities can be used as operational tools for efficient
965 near real time lava flow propagation forecasting (Hyman et al. 2022; Cappello et al. 2022).

966 Automated reporting systems and observatory data bases are advancing all the time, so as to
967 ease the administrative and reporting load of volcano observatory staff. Crowd sourcing through
968 social media, will without a doubt, continue to develop as a resource to contribute to following
969 effusive eruption impacts as well as distributing warning, alerts and information (Wadsworth
970 et al. 2022). Conversely, there are other basics that are slow to improve: funding is never
971 enough, staff numbers are always too low, helicopter time is frequently non-existent, and
972 working hours are long, as are kilometers walked. The field practitioner will thus always argue
973 in terms of reducing equipment weight and bulk along with a well-defined duty and task list,
974 plus developing better means to keep drinking water cold and boots from burning.

975

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