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## A multi-decadal view of the heat and mass budget of a volcano in unrest: La Soufrière de Guadeloupe (French West Indies)

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1 **A multi-decadal view of the heat and mass budget of a volcano in**  
2 **unrest: La Soufrière de Guadeloupe (French West Indies)**

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20

21 **Abstract** Particularly in the presence of a hydrothermal system, many volcanoes output large  
22 quantities of heat through the transport of water from deep within the edifice to the surface. Thus,  
23 heat flux is a prime tool for evaluating volcanic activity and unrest. We review the volcanic unrest  
24 at La Soufrière de Guadeloupe (French West Indies) using an airborne thermal camera survey, and  
25 in-situ measurements of temperature and flow rate through temperature probes, Pitot-tube and  
26 MultiGAS measurements. We deduce mass and heat fluxes for the fumarolic, ground and thermal  
27 spring outputs and follow these over a period spanning 2000–2020. Our results are compared with  
28 published data and we performed a retrospective analysis of the temporal variations in heat flux  
29 over this period using the literature data.

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We find that the heat emitted by the volcano is  $36.5 \pm 7.9$  MW, of which the fumarolic heat flux is dominant at  $28.3 \pm 6.8$  MW. Given a total heated area of  $26\,780\text{ m}^2$ , this equates to a total heat flux density of  $1366 \pm 82\text{ W/m}^2$ , which is amongst the highest established for worldwide volcanoes with hydrothermal systems, particularly for dome volcanoes. A major change at La Soufrière de Guadeloupe, however, is the development of a widespread region of ground heating at the summit where heat output has increased from  $0.2 \pm 0.1$  MW in 2010 to  $5.7 \pm 0.9$  MW in 2020. This change is concurrent with accelerating unrest at the volcano, and the emergence of two new high-flux fumaroles in recent years. Our findings highlight the importance of continued and enhanced surveillance and research strategies at La Soufrière de Guadeloupe, the results of which can be used to better understand hydrothermal volcanic systems the world over.

**Keywords** Heat and mass flux · airborne thermal imagery · Pitot tube · MultiGAS · fumarole

## Introduction

Hydrothermal systems in active island-arc andesitic volcanoes are produced by the interaction of hot magmatic fluids, essentially gaseous water,  $\text{CO}_2$ ,  $\text{H}_2\text{S}$  and/or  $\text{SO}_2$  and  $\text{HCl}$ , produced by magma degassing at depth with marine or meteoric water at shallower depths and the host-rock (Sigurdsson et al 2015; Hedenquist and Lowenstern 1994). Cooling through interaction with water (dissolution and/or absorption into deep ground waters and mixing with meteoric and sea-water) and the host-rock causes chemical species to be reduced (Giggenbach 1975; Moretti and Stefansson 2020). Hence, the geochemical profile of fluids discharged at the surface of a hydrothermal system is typically different to that at depth. Volcanic hydrothermal systems can undergo sudden and catastrophic changes in behaviour. Two events in recent years, in particular, have highlighted the importance of understanding all aspects of volcanic hydrothermal systems and their hazardous behaviour: the September 2014 Ontake (Japan) and December 2019 Whakaari (White Island, New Zealand) eruptions, both of which resulted in the tragic loss of human life.

The boiling of geothermal liquids liberates a fraction of the dissolved gases, which fractionate into the vapour phase that ascends to the surface through steam-dominated fumaroles. Partial condensation of these vapours into ground waters may generate steam-heated waters likely to disperse laterally where they can further mix with external waters and discharge as thermal springs (Hedenquist and Lowenstern 1994; Sigurdsson et al 2015). Therefore, significant amounts of heat are emitted as the super-heated steam, generated by these interactions, rises towards the surface through networks of cracks, fissures and more porous rock within the edifice. The super-heated steam either condenses near the surface or escapes to the atmosphere through fumaroles (Chiodini et al 2001; Fischer and Chiodini 2015; Stimac et al 2015). Heat emission can occur in several

forms. First, where resistance to flow is low (high permeability subsurface) the steam may reach the surface without condensing and, second, where resistance to flow is high (low permeability subsurface) the steam may condense near the surface. In the first scenario, the fumarolic output is high and significant amounts of heat and mass are transferred to the environment. In the second scenario, fumarolic output is correspondingly lower and heat is brought to the surface by forced convection and liberated to the environment by radiation and conduction (Harris 2013; Gaudin et al 2016). This leads to thermal anomalies (ground heating) and small, low-flux fumaroles typically distributed over quite large areas (cf. Aubert et al 1984; Aubert 1999; Harris and Stevenson 1997; Harris and Maciejewski 2000, for example). In many cases, ground heating far exceeds the fumarolic output in terms of energy transfer (Matsushima et al 2003; Mannini et al 2019). Due to the high heat capacity of water, direct fumarolic degassing and diffuse small fumarole/soil degassing are generally the two major components of heat loss at hydrothermal volcanic systems (Aubert 1999; Chiodini et al 2001). The final component of heat transfer in hydrothermal volcanic systems is through a network of thermal springs which typically appear along the flanks or base of the system, taking advantage of structural discontinuities. These springs discharge water, initially heated by volcanic gases, that has either condensed deep within the edifice or nearer the surface when it has come into contact with the water table (Fischer and Chiodini 2015; Stimac et al 2015).

Whilst the relative importance of the different heat loss mechanisms will vary from volcano to volcano and may vary in time at a given site, volcanic heat flow in general is indicative of (e.g. Hardee 1982; Lardy and Tabbagh 1999; Harris et al 2009): 1. The state and position of the magma body; 2. The porosity/permeability of the edifice or dome; 3. The extent of infiltration of external water into the system. As such, spatio-temporal variations in heat flow are of particular importance for both monitoring and fundamental research and allow us to greatly constrain numerical models of the magmatic and plumbing systems (Di Renzo et al 2016).

Hydrothermal systems play a fundamental role in providing and enhancing the physico-chemical conditions that promote rock alteration, as well as the pressurisation of hydrothermal fluids. These processes act as strong forcing and triggering agents on the dynamics of volcanic activity by mechanically weakening edifice-forming volcanic rock (Pola et al 2012; Wyering et al 2014; Heap et al 2015; Mordensky et al 2019) and, therefore, promoting recurrent partial flank collapses (López and Williams 1993; de Vries et al 2000; Reid et al 2001; Reid 2004; John et al 2008), as observed at La Soufrière de Guadeloupe (Komorowski et al 2005; Rosas-Carbajal et al 2016). Escalating pressurisation of hydrothermal systems, as a result of permeability loss due to hydrothermal alteration, can also lead to explosive activity (Heap et al 2019) that can reach paroxysmal levels with non-magmatic laterally-directed turbulent pyroclastic density currents or blasts (e.g. Bandaisan, Japan, in 1889). Hydrothermal alteration has also been observed to reduce the thermal conductivity

98 and thermal diffusivity of andesite for a given porosity (Heap et al 2020). Finally, the hydrother-  
99 mal system is a strong modulator of geophysical and geochemical signals of magmatic unrest and  
100 can generate a plethora of non-magmatic unrest signals that render monitoring, as well as their  
101 interpretation and forecasting, very challenging (Pouget et al 2015).

102 In this paper, which spans the past 20 years with particular emphasis on the 2010–2020 pe-  
103 riod, we concentrate on the use of thermal measurements to infer the state of unrest of a major  
104 hydrothermal volcanic system, that of La Soufrière de Guadeloupe (Lesser Antilles). We present  
105 the first study for this volcano that fully integrates measurements of all the heat sources over such  
106 a long period of time. La Soufrière de Guadeloupe is a good target for such a study due to the  
107 wealth of geochemical, geological and geophysical data acquired on the volcano. As such, it is often  
108 considered a natural laboratory representative of andesitic hydrothermal systems.

## 109 Context

110 La Soufrière de Guadeloupe (16.0446° N, 61.6642° W, alt. 1467 m, hereby referred to as La Soufrière)  
111 is an andesitic dome volcano situated in the south of the Basse-Terre island of Guadeloupe (French  
112 West Indies), which is part of the Lesser Antilles volcanic arc and is the most recent edifice of  
113 the Grande Découverte complex (445 ka, Komorowski et al 2005). La Soufrière is amongst the  
114 most active and potentially deadly of the volcanoes in the Lesser Antilles Arc (Komorowski et al  
115 2005). Hydrothermal activity is sustained by gas and heat transfer from a 6–7 km deep andesitic  
116 magma reservoir to shallower aquifers (Pichavant et al 2018). Owing to an extensive hydrothermal  
117 system, La Soufrière has undergone a series of six phreatic and hydrothermal explosive eruptions  
118 since the last major magmatic eruption in 1530 C.E. (Komorowski et al 2005). The most recent,  
119 and probably most famous, eruption was in 1976–77 (Feuillard et al 1983; Hincks et al 2014).

120 The present edifice dates back at least 9150 years (Komorowski et al 2005; Legendre 2012),  
121 during which time several major magmatic eruptions have occurred, the latter in around 1530 C.E.,  
122 when the current dome was emplaced (Komorowski et al 2005; Boudon et al 2008). Since this last  
123 magmatic event, there have been a number of phreatic and/or hydrothermal explosive eruptions.  
124 The last eruption occurred in 1976–77, following which the volcano became essentially dormant  
125 until 1992 when seismic activity and steam emissions from summit fumaroles recommenced (OVSG-  
126 IPGP 1999-2001<sup>1</sup>; Zlotnicki et al 1992; Komorowski et al 2001, 2005). Summit degassing has  
127 gradually increased concomitantly with other observables (seismic, gas flux and concentration,  
128 ground and fumarole temperatures, deformation, emissions of chlorine-rich acid gases), over the  
129 past ~ 30 years. This has included the appearance of two new high-flux fumaroles (Napoléon Nord  
130 and Napoléon Est, labelled NAPN and NPE on Fig. 1; OVSG-IPGP 2014-2016; Komorowski et al

<sup>1</sup> <http://www.ipgp.fr/fr/ovsg/bulletins-mensuels-de-lovsg>

2005; Villemant et al 2014; Moretti et al 2020a), extensive zones of substantial surface heating and scalding of vegetation. Several fumarolic sites on the flanks characterised by a low state of activity since 1976, gradually vanished. At the summit, Tarissan (TAS), Cratère Sud (CS), la Fente du Nord, Gouffre 56 (G56) and the Lacroix fumaroles had all become inactive by 1984 (Komorowski et al 2005; Boichu et al 2011; Feuillard 2011; Ruzié et al 2013).

An increase in activity in 2018 raised speculation that the volcano is in a state of growing unrest and is likely to undergo another eruptive episode in the near future (Moretti et al 2020a). Until 2014, ground thermal anomalies and accompanying soil degassing had likely been limited to the areas directly surrounding the major fumaroles, as well as the Faille de la Ty/Ravine Claire/Matyлис structure (Fig. 1; OVSG-IPGP 2014-2020; Komorowski et al 2005; Lesparre et al 2012; Brothelande et al 2014). In recent years, however, a number of thermal anomalies and altered zones have been observed such as at the Zone Fumerolienne Napoléon Nord (ZFNN) at the summit, delimited by NAPN, Cratère Dupuy (DUP) and TAS, adjoining the Breislack fault (BLK) and in the upper Matyлис ravine (Fig. 1, OVSG-IPGP 2014-2020 and this work). Increasing fluxes and acidification of the water and gas rising up within the volcano has led to significant alteration and weakening of the edifice, leaving it vulnerable to flank collapse during even moderate seismic activity or extreme rainfall (Komorowski et al 2005; Rosas-Carbaljal et al 2016).

The summit vents are located near major fractures and fault zones, i.e. zones of high vertical permeability (see Fig. 1 and Zlotnicki et al 1992; Komorowski et al 2005). These are likely to have acted as a magma ascent route for the dome (Brombach et al 2000). The deepest part of these fractures act as a zone of preferential input of magmatic gases into the hydrothermal aquifer, and its shallowest part behaves as a zone of preferential discharge for the hydrothermal aquifer (Brombach et al 2000). The horse-shoe shaped scar of recurrent partial edifice collapses over at least the last 3000 years, including the major Amic Crater (1370 BCE) and the 1530 CE events, forms a listric clay-rich low-permeability south-sloping surface. This allows preferential outflow of heated ground waters through a number of thermal springs (Brombach et al 2000; Villemant et al 2005; Ruzié et al 2013; Villemant et al 2014). Here, fluids are heated within the hydrothermal system and then partially cooled by mixing with meteoric water before escaping to the environment.

## Materials and methods

Aerial thermal surveys were carried out in 2010 and 2019, MultiGAS and Pitot-tube measurements have been carried out monthly since 2017 and the thermal springs have been sampled monthly since 2000. Our measurements are effectively contemporaneous even though sampling times differ between different methods and sites. We calculated errors on our estimations using standard error propagation formulae (Ku 1966; Gibbings 1986). Examples of how to apply these formulae and

165 a table of relative standard errors for all the parameters used in this study can be found in the  
 166 Supplementary Material.

### 167 Ground thermal anomaly flux

168 We used airborne thermal imagery to measure the extent and distribution of thermal anomalies  
 169 over the entire volcano using an InfraTec VarioCam HD thermal camera (8–14  $\mu\text{m}$ ) with  $640 \times 480$   
 170 pixel resolution. A 15 mm focal length lens ( $56.1 \times 43.6^\circ$  FOV), gave an instantaneous field of view  
 171 (IFOV) of 1.65 mrad. The distinguishable temperature difference between neighbouring pixels,  
 172  $\text{NE}\Delta T$ , was 0.5 K. We used an Isotech Calisto calibration oven with black-body source to calculate  
 173 the drift of the camera’s temperature measurements and applied this to our thermal images.

174 The airborne thermal survey was conducted on 22 November 2019 with helicopter support  
 175 provided by the local Civil Protection Service and in pre-dawn conditions (05:40–06:05 local time,  
 176 sunrise was at 06:14 local time). Ground heating due to incoming solar radiation was hence min-  
 177 imised. GPS locations were recorded at 1 Hz using a Garmin 64st. Images were acquired through  
 178 the open door of the helicopter from heights of about 50–300 m. The sky was cloudless with only a  
 179 very light wind from the North (cf. the predominant trade winds, les Alizées, blow from the East).  
 180 Little rain had fallen in the week prior to the survey, so the ground surface was dry.

181 We georeferenced and orthorectified our thermal images using a DEM calculated from Institut  
 182 Geographique National (IGN) aerial photography, processed using MicMac (Rupnik et al 2017)  
 183 and IGN orthophotos (Fig. 2). Georeferencing of the thermal images was performed in QGIS using  
 184 a thin-plate spline transform when the images were taken obliquely, or a Helmert transform for  
 185 vertically-oriented images. Pixel to physical distance conversions were computed as per Bombrun  
 186 et al (2018).

187 A schematic of the various fluxes seen by the camera is shown in Fig. 3. The effective brightness  
 188 temperature is a function of the incoming fluxes which are functions of the temperature of the  
 189 objects in the field of view through the Stefan-Boltzmann law. Brightness temperature is also  
 190 affected by reflection of incoming radiation (e.g.  $L_{\text{sol}}$  and  $L_{\text{atm}}$  in Fig. 3). Thus the true temperature  
 191 of the ground can be expressed as

$$192 \quad T = \left( \frac{T_{\text{cam}}^4 - T_{\text{atm}}^4 - (1 - \tau)T_g^4}{\epsilon\tau} \right)^{1/4}, \quad (1)$$

193 where  $T_{\text{cam}}$  is the brightness temperature seen by the camera,  $T_{\text{atm}}$  is the brightness temperature  
 194 of the upper atmosphere,  $T_g$  is the temperature of gases between the object and camera,  $\tau$  is the  
 195 transmissivity of an atmospheric and volcanogenic gas mixture between camera and the ground  
 196 and  $\epsilon$  is the emissivity of the ground (Fig. 3).

197 We converted at-camera (brightness) temperature to absolute temperature by applying Eq. 1.  
 198 Fumarole plumes and areas outside the region of interest were masked. We calculated  $\tau$  using a  
 199 radiative transfer model, with the surface-camera distance given by the georeferenced images and  
 200 GPS location of the camera (Kochanov et al 2016; Berk and Hawes 2017). We took the surface  
 201 emissivity to be constant for all the heated areas with  $\epsilon = 0.95$  in line with that found for other  
 202 studies on andesitic systems (Sekioka and Yuhara 1974; Gaudin et al 2016).

203 We note that not all of the steam condenses before reaching the surface. Condensed liquids  
 204 drain away to be discharged elsewhere in the system (i.e. through thermal springs, in which case  
 205 the heat transported is accounted for in the thermal springs heat budget) and any residual heat  
 206 transferred to the ground, where it is accounted for in the soil heat budget. Hence we do not  
 207 consider heat transported by condensed water here (cf. Gaudin et al 2015). Our heat balance is  
 208 thus (Sekioka and Yuhara 1974; Matsushima et al 2003; Harris 2013; Mannini et al 2019),

$$Q_{\text{soil}} = Q_{\text{soil,rad}} + Q_{\text{soil,conv}}, \quad (2)$$

$$Q_{\text{soil,rad}} = A_{\text{heated}} \epsilon_{\text{soil}} \sigma (T^4 - T_{\text{amb}}^4), \quad (3)$$

$$Q_{\text{soil,conv}} = A_{\text{heated}} h_c (T - T_{\text{amb}}), \quad (4)$$

209 where  $Q_{\text{soil}}$  is the soil heat flux and subscripts rad and conv refer to radiative and conductive  
 210 components of  $Q_{\text{soil}}$ , respectively,  $T$  is the ground temperature,  $T_{\text{atm}}$  is the ambient temperature  
 211 and  $A_{\text{heated}}$  is the heated area,  $\epsilon_{\text{soil}}$  is the soil emissivity. The heat transfer coefficient,  $h_c$ , depends  
 212 on several factors, particularly the local wind speed,  $w$ . We use the Schlichting-Neri model (Neri  
 213 1998; Gaudin et al 2013)

$$214 \quad h_c = 1500w(z) (1.89 + 1.62 \log(z/z_0))^{-2.5}, \quad (5)$$

215 where  $z$  is the height above the surface and  $z_0$  is a measure of the surface roughness. Eq. 5 has been  
 216 shown to produce results that are consistent with the surface heat balance at La Soufrière (Gaudin  
 217 et al 2013), such that the heat conducted to the surface equals  $Q_{\text{soil}}$ . We note that the surfaces  
 218 on the volcano where heat transfer occurs consist typically of centimetric blocks and thus we take  
 219  $z_0 = 0.01$  m as our roughness scale. We determined  $w$  from measurements at the Sanner weather  
 220 station (cf. Piton Sanner in Fig. 1) at the time of thermal image acquisition. The anemometer at  
 221 Sanner is approximately 2 m above ground level, so we take  $z = 2$  m in our calculations. For wind  
 222 speeds between 5–10 m/s, as seen on the 22 November, we find  $h_c$  between 21.1 and 42.3 W/(m<sup>2</sup> K).  
 223 Considering error propagation, we estimate a relative standard error of about 10% on the radiative  
 224 and convective flux measurements, and thus about 15% for the total flux.



225 Fumarole heat and mass fluxes

226 *In-plume fumarole steam flux via MultiGAS traverses*

227 The OVSG<sup>2</sup> MultiGAS consists of an IR spectrometer for CO<sub>2</sub> determination and electro-chemical  
228 sensors for SO<sub>2</sub>, H<sub>2</sub>S and H<sub>2</sub>. The atmospheric pressure is determined with the sensor installed  
229 on the CO<sub>2</sub> spectrometer card. The MultiGAS also includes an externally-fitted relative humid-  
230 ity (RH) sensor (Galltec, range: 0–100% RH, accuracy:  $\pm 2\%$ ) and temperature sensor (range:  
231  $-30$ – $70$  °C, resolution: 0.01 °C), to determine water vapour concentration (Moussallam et al 2017).  
232 H<sub>2</sub>O determination with these external sensors reduced the risk of underestimating the measured  
233 water/gas ratios due to steam condensation in the inlet. An onboard GPS receiver tracked the  
234 location of the instrument at 1 Hz. Data were visualised on an external tablet in real time. More  
235 detailed information about the OVSG MultiGAS, its design and performance characteristics can  
236 be found in Tamburello et al (2019) and Moretti et al (2020a,b).

237 Fumarolic gas fluxes were determined for the three main vents that generate plumes (CS, TAS  
238 and G56, Figure 1) following Allard et al (2014) and Tamburello et al (2019). The horizontal  
239 and vertical distributions of gas species in the plume cross-sections were measured a few meters  
240 downwind from the vents during orthogonal traverses on foot. Gas concentrations were measured  
241 at two different heights (typically 0.9 and 2 m) as the volcanic gas plumes are generally flattened  
242 to the ground by strong trade winds (2–14 m/s) and have a maximum height of ca. 3–4 m above  
243 the ground at each measuring site with a maximum gas density centred at between  $\sim 1.5$  and  
244 2 m above the ground (our visual observations; Gaudin et al 2016; Tamburello et al 2019). For  
245 each site, we interpolated the concentration measurements using a 2D spline function and then  
246 integrated over the plume cross section to obtain integrated concentration amounts (ICAs) using  
247 RatioCalc (Tamburello 2015). The CO<sub>2</sub> fluxes are derived by multiplying the CO<sub>2</sub> concentration  
248 integrated over the plume cross section with the wind speed measured during the gas survey with  
249 a hand-held anemometer. We use CO<sub>2</sub> as the volcanic marker as, due to its more conservative  
250 behaviour compared to H<sub>2</sub>S and due to the faster response of the IR CO<sub>2</sub> sensor compared to  
251 the electro-chemical H<sub>2</sub>S sensor, this way the MultiGAS is able to detect rapid concentration  
252 changes during plume transects. This avoids flux underestimations and leads to more accurate  
253 gas flux measurements (Tamburello et al 2019). Due to the high atmospheric background for  
254 H<sub>2</sub>O and CO<sub>2</sub>, our walking profiles start and end in pure atmospheric background in order to  
255 characterise and then subtract the ambient air composition from our recorded data. Steam fluxes,  
256  $\dot{m}$ , are derived from the CO<sub>2</sub> flux by multiplying it by the weight ratio of H<sub>2</sub>O/CO<sub>2</sub>. Steam flux  
257 estimates were possible only when water was successfully determined via the external RH sensor.

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<sup>2</sup> Observatoire Volcanologique et Simologique de la Guadeloupe

258 It is important to note that some temporal variability of steam fluxes could be due to: i) different  
 259 ambient humidity and weather condition at the summit between field measurements; ii) occasional  
 260 partial steam condensation on the external sensors. Indeed, particularly for tropical volcanoes such  
 261 as La Soufrière, water vapour in the plume rapidly condenses upon contact with the atmosphere.  
 262 However, this condensed water is not taken into account by the MultiGAS measurements. It has  
 263 been shown that, in such tropical conditions, properly accounting for the condensed water adds  
 264 approximately 35% to the steam flux estimations (Gaudin et al 2016), an increase which we consider  
 265 in our analysis. Lastly, wind speed is the main source of error in quantifying volcanic gas fluxes,  
 266 leading to typical standard errors on steam flux estimation of about 40%.

### 267 *At-vent fumarole fluxes via Pitot-tube measurements*

268 Measurements of the steam exit speed at the vent of several fumaroles were made using a Pitot-  
 269 tube instrument based around Freescale MPX2200AP and MPX2010DP temperature compensated  
 270 pressure sensors that measured the dynamic pressure in the moving stream and ambient (stagna-  
 271 tion) pressure. Pressure readings were taken at 3.75 Hz and the median of 10 measurements was  
 272 recorded by an Arduino Due. Uncertainty in the pressure readings was 3.1 Pa, meaning that the  
 273 minimum detectable speed was about 5 m/s. From these values, the speed of a moving stream of  
 274 gas,  $u$ , of density  $\rho$  was calculated as (Massey and Ward-Smith 1998),

$$275 \quad u = \sqrt{\frac{2\Delta p}{\rho}}, \quad (6)$$

276 where  $\Delta p = p_0 - p$  is the dynamic pressure,  $p_0$  is the stagnation pressure and  $p$  is the free-stream  
 277 pressure. Vent temperatures were simultaneously measured using a PT1000 resistance temperature  
 278 sensor with an instrumental error of  $\pm 1$  K. These measurements, along with the pressure readings,  
 279 were used to calculate the steam density,  $\rho(p, T)$ , using numerical codes based on IAPWS ther-  
 280 modynamic calculations (Wagner and Pruß 1993, 2002). Measurements were taken repeatedly at  
 281 different points across the vent in order to build up an idea of the velocity distribution. Typically,  
 282 7 measurements were taken and the median velocity from these measurements was used in the  
 283 calculations that follow.

284 From vent speed, we deduce the mass flux from the fumaroles which, as water vapour contributes  
 285 up to 98% of the total mass (OVSG-IPGP bulletins 2017-2020; Allard et al 2014; Tamburello et al  
 286 2019; Moretti et al 2020a), is equivalent to the steam flux,

$$287 \quad \dot{m} = \rho \bar{u} A \approx \rho_{\text{steam}}(T) \bar{u} A, \quad (7)$$

where  $\bar{u}$  is the mean vent speed (equivalent in this case to median vent speed) and  $A$  is the area of the vent. Whereas in Moretti et al (2020a), vent area was estimated by eye by the Pitot-tube operator, here we calculate  $A$  by analysing thermal images. We repeatedly took thermal images looking straight into the vents throughout the period when Pitot-tube measurements were made, from which we manually traced around the vent perimeter and, using the on-camera laser distance measurements, then converted to physical area (i.e. in  $\text{m}^2$ ) via a pixel-to-physical length conversion as per Bombrun et al (2018). We estimated the relative standard error on mass flux measurements to be 10%.

### Heat flux estimations

The heat released through fumarolic activity is essentially due to cooling and condensation of the volcanic steam. Fumarole heat flux can generally be decomposed into two contributing factors: radiation by the heated vent surface,  $Q_{\text{rad}}$ , and the specific and latent heat carried by the gas phase,  $Q_{\text{gas}}$ , so that  $Q_{\text{fumarole}} = Q_{\text{rad}} + Q_{\text{gas}}$  (Harris, 2013; Gaudin et al, 2016). Heat lost to the surroundings through the walls of the fumarole pipes is not considered as part of this heat budget, but are accounted for through the geothermal heating of the surrounding ground (Stevenson 1993; Mannini et al 2019), as shown in the previous section. Following Harris (2013), Allard et al (2014) and Gaudin et al (2016) we write these as

$$Q_{\text{rad}} = A\epsilon\sigma (T^4 - T_{\text{amb}}^4), \quad (8)$$

$$Q_{\text{gas}} = \dot{m} (c_{p,v}(T) (T - T_{\text{boil}}) + L(T) + c_{p,l} (T_{\text{boil}} - T_{\text{amb}})), \quad (9)$$

where  $\epsilon$  is the ground emissivity,  $\sigma$  is the Stefan-Boltzmann constant,  $c_p$  is the specific heat capacity,  $L \approx 2260 \text{ kJ}/(\text{kg K})$  the latent heat of condensation,  $T$  is the temperature of the steam,  $T_{\text{boil}} \approx 96.7^\circ\text{C}$  is the boiling temperature of water at the dome altitude and  $T_{\text{amb}} \approx 17^\circ\text{C}$  is the ambient temperature at the summit. The subscripts  $v$  and  $l$  refer to the vapour and liquid states of water, respectively, with  $c_{p,v} \approx 2.015 \text{ kJ}/(\text{kg K})$  and  $c_{p,l} \approx 4.200 \text{ kJ}/(\text{kg K})$  for summit temperatures and pressures. During the survey period,  $T$  ranged from  $96.9$  and  $108.6^\circ\text{C}$  for CS and has been measured in the water lake at TAS to be approximately  $97.5^\circ\text{C}$  (OVSG-IPGP 2016-2020). Since the G56 vent is in a c. 30 m deep cavity within the volcano, requiring specialised equipment to access, it is impractical to measure it directly. Thus we estimate that the temperature at the G56 vent is at the boiling temperature of water. We note that  $c_p$  and  $L$  are functions of  $p$  and  $T$  and were solved for using similar numerical routines as for density.  $T_{\text{boil}}$  is a function of pressure only and is also deduced from the IAPWS formulations (Wagner and Pruß 1993, 2002).

317 Given the instrumental and measurement errors summarised in the text above, and using error  
 318 propagation techniques (see Supplementary Material for details), we estimated the standard errors  
 319 on the flux estimation using the Pitot-tube and MultiGAS instruments. In the case of the Pitot-  
 320 tube instrument, the standard error in estimating  $Q_{\text{fumarole}}$  is dominated by the mass flux and  
 321 radiative flux terms and, overall, is around 10%. The standard error for our estimates based on  
 322 MultiGAS measurements is dominated by the uncertainty in the mass flux measurements alone  
 323 and so is about 40%.

### 324 Thermal Springs

325 The nine thermal springs situated around the base of the current dome have been monitored  
 326 regularly by the OVSG since 1978 by manually measuring temperature and flow rate. The majority  
 327 of sites have been visited on a 1–3 month basis to take manual temperature readings as well as  
 328 physico-chemical parameters such as pH and conductivity, and to take samples for future chemical  
 329 analysis (Villemant et al 2005, 2014). During these outings, and when it was possible, volumetric  
 330 flow rate,  $\dot{V}$ , was deduced from the time taken to fill a container of known volume. This process  
 331 was repeated 6–10 times and we report here the mean value of these measurements. From this, we  
 332 calculate the mass flow rate,  $\dot{m}_{\text{spring}} = \rho\dot{V}$ , which then allows us to calculate the heat flux as the  
 333 sum of specific, evaporative and radiative heats,

$$334 \quad Q_{\text{spring}} = Q_{\text{spec}} + Q_{\text{evap}} + Q_{\text{rad}} \approx Q_{\text{spec}} = \dot{m}c_{p,l}(T)(T - T_{\text{amb}}). \quad (10)$$

335 We drop evapotransport,  $Q_{\text{evap}}$ , and radiative heat losses,  $Q_{\text{rad}}$ , in Eq. (10) as these contribute  
 336 negligibly to the heat budget. The relative standard error on these measurements is about 5%.

## 337 Results

### 338 Ground heat flux

339 We show our results from the analysis of the thermal images (Fig. 2) in Table 1. For each site  
 340 with detected thermal activity (summit, lower Ravine Matylis, Ravine Claire and FTY), we have  
 341 determined radiative and convective fluxes as well as the flux density,  $q_i = Q_i/A$ . As large fluxes  
 342 can be observed by low intensity emissions over a large area, we also calculate the total heat flux  
 343 density,  $q = (Q_{\text{rad}} + Q_{\text{conv}})/A_{\text{heated}}$ , as a metric for comparing intensity between sites (Table  
 344 1) with a relative standard error of about 6%. At the summit we found the radiative flux to  
 345 be 0.74 MW and the convective flux to be 4.94 MW, with  $h_c = 41.6 \pm 2.3 \text{ W}/(\text{m}^2 \text{ K})$  (Eq. 5) and  
 346  $A_{\text{heated}} = 14070 \text{ m}^2$  (areas were calculated by counting the number of heated pixels, e.g. Fig. 2).

347 These values were by far the largest in magnitude of all the sites, and larger than the total heat  
 348 fluxes for the other sites combined. This finding is supported by the heat flux density, which is  
 349 considerably greater than any other site.

350 We calculated  $h_c = 37.8 \pm 2.3 \text{ W}/(\text{m}^2 \text{ K})$  for wind speeds during acquisition of the images of  
 351 the flank sites, which was used in the calculations for all sites. Owing to a relatively large emitting  
 352 surface of  $8010 \text{ m}^2$ , the heat flux at Ravine Claire (RC) is second to the summit with radiative and  
 353 convective fluxes of  $0.13 \text{ MW}$  and  $0.75 \text{ MW}$  (Table 1), respectively. In the lower Matylis ravine, a  
 354 strong thermal anomaly leads to high flux densities ( $42$  and  $238 \text{ W}/\text{m}^2$  for radiation and convection,  
 355 respectively), although a low heated area ( $1630 \text{ m}^2$ ) keeps the overall fluxes low. We identified two  
 356 sites along FTY (FTY0 and FTY1 in Figs. 1 and 2) which have similar results for flux density  
 357 and had a total flux of  $0.58 \text{ MW}$  and a mean flux density of  $227 \text{ W}/\text{m}^2$ . We note that all these  
 358 sites (Matylis, RC, FTY) are linked to the Ty N-SE and Galion N-S faults that cut the dome  
 359 (Komorowski et al 2005; Rosas-Carbajal et al 2016).

#### 360 Fumarole heat and mass flux

361 The mass and heat fluxes are shown in Fig. 4 a) and b), respectively. Steam fluxes estimated  
 362 from MultiGAS traverses show that, using  $\text{CO}_2$  as a marker, the min/mean/max values are:  
 363  $0.35/0.52/0.86$ ,  $0.15/0.30/0.47$  and  $0.29/0.44/0.67 \text{ kg/s}$  for CS, G56 and TAS, respectively. Heat  
 364 flux estimates based on these data give  $0.93/1.36/2.28$ ,  $0.40/0.79/1.25$  and  $0.78/1.12/1.79 \text{ MW}$  for  
 365 CS, G56 and TAS, respectively. Considering the relative standard error of 40%, we find that the  
 366 MultiGAS fluxes have remained stable since regular estimates began in mid 2018, subsequent to  
 367 the M4.1 earthquake.

368 The temporal variations in fumarole steam flux calculated by the Pitot-tube for the three vents  
 369 at CS are also shown in Fig. 4a. These data indicate that the fluxes can show a large degree of  
 370 variation in short time periods, which is especially true during periods of accelerated unrest such as  
 371 from March–May 2018 (Moretti et al 2020a). We find steam fluxes to have min/mean/max values of  
 372  $0.01/0.12/0.31 \text{ kg/s}$  at CSC,  $0.22/0.70/1.50 \text{ kg/s}$  at CSN and  $0.82/2.71/3.85 \text{ kg/s}$  at CSS (Fig. 4a),  
 373 which equate to heat fluxes of  $0.03/0.29/0.75 \text{ MW}$  for CSC vent,  $0.53/1.69/3.64 \text{ MW}$  for CSN vent,  
 374 and  $1.99/6.56/9.31 \text{ MW}$  for CSS vent (Fig. 4b). Fluxes were also measured at NAPN and we found  
 375 mean mass and heat fluxes of  $0.03 \text{ kg/s}$  and  $0.07 \text{ MW}$  with very little variation over time, including  
 376 during the 2018 unrest. The contribution of NAPN to the total heat and mass budget is thus  
 377 negligible. We note that, due to an improved method for estimating vent area based on head-on  
 378 thermal images compared to visual estimation during measurements (see Methods), the vent heat  
 379 fluxes presented here are quantitatively lower than reported in Moretti et al (2020a), although the  
 380 qualitative temporal variation is the same. The Pitot-tube data show that vent fluxes at CS were

381 strongly affected by and decreased during the 2018 unrest phase, but have since settled to around  
382 4 kg/s and 10 MW for mass and heat flux, respectively. Fig. 4b) shows, although the coefficients for  
383 heat capacity and latent heat vary with temperature, the same trends as per Fig. 4a), indicating  
384 that the heat flux depended much more strongly on the variations in mass flux than temperature  
385 changes during this period.

386 At CS, we have overlap in the Pitot-tube and MultiGAS instrument data that allows us to  
387 compare the data collected by these instruments from closely spaced outings. For example, in  
388 terms of steam flux the Pitot-tube data from 15 June 2020 show that CSC+CSN+CSS emitted  
389 around  $4.6 \pm 0.5$  kg/s. The flux estimated from MultiGAS measurements at CS on 22 May 2020  
390 were  $0.7 \pm 0.3$  kg/s. These MultiGAS estimates are almost an order of magnitude times lower than  
391 those from the Pitot-tube, and Fig. 4 indicates that this is systematically the case. Whilst we have  
392 attempted to correct for the quantity of condensed vapour that is undetectable by the MultiGAS,  
393 additional errors in this calculation are likely to be primarily responsible for the difference be-  
394 tween these two values, although they agree to within an order of magnitude and appear to show  
395 qualitatively the same temporal variations.

#### 396 Thermal springs heat and mass fluxes

397 In Fig. 5, we present the mass flow rate and temperature measured over the period between 2000  
398 and 2020 (Figs. 5a and b, a subset of the entire data set, see Villemant et al 2005), as well as  
399 the heat flux calculated from this via Eq. (10) (Fig. 5c). Whilst the flow rate and temperature  
400 measurements have continued until the present day, there are gaps in the mass and heat flux data  
401 during 2014–2016 due to instrument failures. The GA, Tarade spring (TA), Bains Jaunes (BJ)  
402 and Pas du Roy (PR) springs are amongst the most accessible and this is reflected in both the  
403 abundance and persistence of the measurements in the OVSG database. They are also the most  
404 representative of acid-sulphate thermal springs linked to La Soufrière’s hydrothermal activity. This  
405 record does not reflect the absolute total mass/thermal output of the thermal springs as i) other  
406 sites are known but are far less accessible or impractical to measure and ii) some sites may not yet  
407 have been discovered. However, particularly as GA and TA have the largest known flow rates, it  
408 is likely that these calculations are nonetheless representative of the total budget for the thermal  
409 springs. We fitted linear trends to the data for TA, GA and PR, and extrapolated where necessary  
410 to project the values to the current date.

411 Overall, we see that both mass flow rate and water temperature have slowly and steadily  
412 increased over time in an approximately linear fashion. For example, the flow rate at TA increased  
413 from around 1.1 kg/s in 2010 to 2.1 kg/s at present whilst its temperature rose from 309 to 318 K.  
414 Only the TA and PR sites have data that cover the whole data range and manual measurements

415 stopped at GA in 2014. Historically, GA dominates the heat budget for the thermal springs, and  
416 has almost double the output of TA. Summing over these three sites, we find that the total heat  
417 flux from the thermal springs is around 0.57 MW (including the extrapolated trend for GA).

## 418 Discussion

419 Comparison of steam and heat flux estimation methods

### 420 *Fumarole flux*

421 Our measurements (Fig. 6) show that the plume mass and heat fluxes have not undergone extensive  
422 evolution since August 2005. With this in mind, we must consider that the fumarole plume heat  
423 and mass flux estimates of Gaudin et al (2016) to be excessively high. In their discussion, the order  
424 of magnitude discrepancy with the estimations from MultiGAS traverses (Allard et al 2014) was  
425 mostly attributed to the MultiGAS studies not accounting for condensed water vapour. However,  
426 we note several key assumptions in Gaudin et al (2016) that may have led to systematic errors in  
427 their estimations:

- 428 – Plume thickness grows linearly with distance from the vent,  $x$ , and not  $x^{2/3}$ . This latter scaling  
429 is for the height of the plume axis from the ground (Slawson and Csanady 1967), so that their  
430 mass flux integral overestimated the plume area.
- 431 – The plume section was assumed to be axisymmetric though this is generally not true: wind-  
432 blown plumes from smoke stacks, cooling towers and in laboratory experiments have been  
433 shown to be more broad (horizontally) than thick (vertically, Contini and Robins 2001). Thus,  
434 transmissivity calculated looking horizontally through a plume would have been lower than was  
435 actually the case, leading to an overestimation of plume temperature. Consequently the plume  
436 density and thus the mass flow rate should be lower than the given estimations.
- 437 – The vapour-carrying capacity of the plume was assumed to be equal to that of the atmosphere.  
438 However, as plume temperatures are higher than atmospheric and more water vapour can  
439 therefore be carried without condensing, this relationship does not hold.

440 Overall, this suggests that a more realistic plume flux for 2010 would be more in line with the  
441 MultiGAS (taking into account condensed vapour) and Pitot-tube measurements, that is a steam  
442 flux of 5.3 kg/s for CS. Thus, taking the Pitot-tube measurement at CS as the ground truth, we  
443 found a scaling factor for the MultiGAS measurements. Using this to scale the MultiGAS flux for  
444 TAS gives 6.2 kg/s. Likewise, we find heat fluxes of 13.0 MW for CS and 15.2 MW for TAS (see  
445 Table 2) for the 2010 study.

446 Moreover, despite some assumptions, our results show that MultiGAS traverses and Pitot-tube  
447 measurements provide qualitatively coherent vent flux estimations, yet quantifying the steam flux  
448 using MultiGAS is a challenge, particularly in a tropical atmosphere with 100% RH. To over-  
449 come this shortcoming, we take the  $\text{H}_2\text{O}/\text{CO}_2$  ratio determined from Giggenbach bottle sampling  
450 (OVSG-IPGP 2017-2020; Moretti et al 2020b) and multiply this by the  $\text{CO}_2$  flux estimated from  
451 the MultiGAS data. As this ratio is measured at the vent, it is not subject to a loss of matter due  
452 to condensation contrary to measurements within the plume. The resulting fluxes at CS resemble  
453 much more closely the Pitot-tube-derived fluxes (see “Reworked CS MG data” in Fig. 4). This  
454 correlation starkly indicates the difficulties in accounting for condensed volcanogenic vapour in the  
455 MultiGAS steam-flux estimations. Nevertheless, in a monitoring context, either or both methods  
456 could be applied in various volcanoes worldwide to estimate their mass and heat fluxes.

#### 457 *Ground flux*

458 Although we have used the same model for  $h_c$  as Gaudin et al (2016), we obtain slightly different val-  
459 ues simply due to differing weather conditions (compare  $h_c = 41.6 \pm 2.3$  and  $37.8 \pm 2.3 \text{ W}/(\text{m}^2 \text{ K})$   
460 for the summit and flanks, respectively, with  $h_c = 30.5 \pm 0.3 \text{ W}/(\text{m}^2 \text{ K})$  as derived from data in Ta-  
461 ble 2 of Gaudin et al 2013). Thus, similarity between the results of our study and those of Gaudin  
462 et al (2013) at the same site would be suggestive of a decrease in temperature at that site. A good  
463 comparison can be made at the FTY sites. We note in particular that the mean total heat flux  
464 density for FTY0 + FTY1 sites combined,  $228 \pm 14 \text{ W}/\text{m}^2$ , is in strong agreement with the heat  
465 flux density calculated from temperature gradient measurements of  $265 \pm 45 \text{ W}/\text{m}^2$  (Gaudin et al  
466 2013), which suggests that, on average, temperatures have not decreased (the ambient temperature  
467 during the 2010 survey and ours was approximately  $17^\circ\text{C}$  in both cases). It is somewhat unclear  
468 precisely where Gaudin et al (2016) defined the boundary of FTY and, indeed, their Fig. 1 suggests  
469 that this might extend into what we define as Ravine Claire, so judging the evolution of the extent  
470 of this site is difficult. However, taking uncertainties into account, the present-day total flux for  
471 FTY0+FTY1+RC of  $1.46 \pm 0.23 \text{ MW}$  is not too dissimilar to the  $1.0 \pm 0.2 \text{ MW}$  reported in 2010.

472 Given the angular resolution of the thermal camera and camera-to-ground distances of 50–  
473 300 m, this equates to a pixel length of between 0.08 and 0.5 m. As fumaroles are often of smaller  
474 dimension than the resolution length-scale, especially when recording from greater distances, their  
475 temperature is integrated over the pixel area along with the cooler surroundings and so pixels  
476 that cover thermally non-homogeneous ground will display temperatures lower than the true tem-  
477 perature of the hotter object. Harris et al (2009) demonstrated that, at a distance of 100 m, the  
478 pixel-integrated temperature of a  $6 \times 13 \text{ cm}$  ( $78 \text{ cm}^2$ ) fumarole in a  $169 \text{ cm}^2$  pixel was lower than  
479 the actual temperature by  $40^\circ\text{C}$ . Taking this into account, we must consider that the fluxes that



480 we calculate are minimum estimates, emphasising the importance of the ground heat flux for La  
481 Soufrière.

482 Total heat budget

483 As noted by Gaudin et al (2016), some heat loss may be undetectable by the methods described  
484 in this work, due to either vegetation cover (e.g. on the flanks) or temperature changes that are  
485 below the instrument resolution (summit in the region of CS). As per their work, we note that the  
486 “background” heat conducted through the system, as deduced from borehole measurements and  
487 extrapolated to the scale of the dome adds only an additional 0.013 MW. Furthermore, we note  
488 that some heat will be transported by gases other than steam, notably CO<sub>2</sub> in the plume and, in  
489 particular, CO<sub>2</sub> soil degassing which is a widely-used proxy for heat flux (cf. Chiodini et al 1998;  
490 Bloomberg et al 2014; Harvey et al 2015). A detailed study is beyond the scope of this present  
491 work but we may make progress under the following assumptions:

- 492 1. Passive CO<sub>2</sub> degassing occurs in the same areas and to the same extent as the ground thermal  
493 anomalies.
- 494 2. The ground heat flux,  $Q_{\text{soil}}$ , calculated above equals the underground convection of steam,  
495  $\dot{m}_{H_2O} c_{p,H_2O} (T - T_{\text{amb}})$ .
- 496 3. The CO<sub>2</sub>/H<sub>2</sub>O ratio in areas of soil degassing is the same as in the fumaroles (Chiodini et al  
497 2001).

498 Under assumption (2),  $\dot{m}_{H_2O} = 135.5$  kg/s based on a typical anomaly temperature of 80 °C  
499 and ambient temperature of 15 °C. Based on a H<sub>2</sub>O/CO<sub>2</sub> ratio of 43.5 (OVSG-IPGP, 2020), this  
500 gives  $\dot{m}_{CO_2} = 3.2$  kg/s and thus  $Q_{CO_2} = \dot{m}_{CO_2} c_{p,CO_2} (T - T_{\text{amb}}) = 0.19$  MW.

501 Clearly, although comparable to the contributions of certain thermal springs, heat transport by  
502 CO<sub>2</sub> does not add significantly to the total budget. Nevertheless, due to the accelerating spread of  
503 the altered zones and the fact that the area over which CO<sub>2</sub> degassing occurs may be far greater  
504 than that involved in degassing of water vapour (cf. Chiodini et al 2005), the OVSG has begun to  
505 carry out joint surveys of soil temperature profiling and CO<sub>2</sub> flux (via the accumulation chamber  
506 technique, Chiodini et al 1998) in order to further constrain ground heat losses and we will return  
507 to this issue in a forthcoming paper.

508 Summing the fumarole ( $28.3 \pm 6.8$  MW, 77.5%), ground ( $7.6 \pm 1.1$  MW, 20.8%) and the thermal  
509 springs fluxes (0.56 MW 1.5%), we obtain a total heat output of  $36.5 \pm 7.9$  MW (see Table 2).

## 510 Temporal evolution

511 La Soufrière has undergone a significant evolution of its activity during 2010–2020 as described and  
512 analysed in detail by Moretti et al (2020a) and OVSG-IPGP (2014-2020). This can be summarised  
513 as follows:

- 514 1. The appearance of new fumarolic vents and the reactivation of pre-existing fumaroles with local  
515 high-velocity degassing.
- 516 2. Vegetation die-off near-to and far-from active vents (see Supplementary Material Fig. 1).
- 517 3. The enlargement of a major extensive area of heated ground on the summit areas that progresses  
518 towards the North from the Fracture Napoléon (see below and Supplementary Material Fig. 2).
- 519 4. More frequent and stronger seismic events including felt events (M4.1, April 2018).
- 520 5. An acceleration in the opening rate of several summit fractures.
- 521 6. The appearance of new mineralised water springs at the base of the volcano as a result of the  
522 rapid cooling of hydrothermal fluids.

523 Undoubtedly the greatest phenomenological change at the summit of La Soufrière is indeed the  
524 appearance and spread of the ground thermal anomaly in the ZFNN region. For example, the  
525 heated area at the summit has gone from an estimated 610 m<sup>2</sup> in 2010 (Gaudin et al 2016) to  
526 14 070 m<sup>2</sup> for the present study. Indeed, whereas Gaudin et al (2016) identified thermal anomalies  
527 along the Napoléon fracture and in the craters containing the CS fumaroles, they calculated that  
528 the associated heat losses were 0.01 MW by radiation and 0.2 MW by forced convection. The  
529 present-day radiative and convective fluxes of  $0.74 \pm 0.07$  and  $4.94 \pm 0.49$  MW have increased by  
530 an order of magnitude between 2010 and 2020, which is in large part due to this increase in heated  
531 area and also, to a lesser degree, because of increased temperatures. The total heat flux density  
532 presently estimated at  $403 \pm 26$  W/m<sup>2</sup> is greater than the 2010 estimate of  $326.3 \pm 68.5$  W/m<sup>2</sup>,  
533 and thus suggests an increase in thermal intensity at the summit, though these values are within  
534 the bounds of measurement uncertainty.

535 Apart from the pulse of unrest around March–April 2018, the fumarolic fluxes have not changed  
536 considerably since early 2018 (Fig. 4), and the thermal spring fluxes have increased only slightly  
537 (Fig. 5) over the past 20 years. To gain a greater perspective of the overall temporal evolution  
538 of plume fluxes over a similar period, we plot in Fig. 6 our data along with the steam and heat  
539 fluxes taken from Allard et al (2014), Gaudin et al (2016) and Tamburello et al (2019). This figure  
540 shows that our current data are, given the natural variability of these fluxes, consistent with the  
541 previous MultiGAS measurements of Allard et al (2014) and Tamburello et al (2019). They are far  
542 more consistent with the Pitot-tube measurement for the CSN + CSC vents cited in Gaudin et al

(2016) than the fluxes that they derived from analysis of thermal images (compare their value of  $5.3 \pm 1.6$  kg/s to the contemporary sum for CSN + CSC of around  $1.4 \pm 0.3$  kg/s, Fig. 4a).

Three major swarms of volcano-tectonic (VT) earthquakes occurred from 1 February to 28 April 2018, with the third swarm initiated by the off-volcanic axis M4.1 earthquake which struck at 00:32 UTC on 28 April and was widely felt throughout Guadeloupe. In particular, as reported in Moretti et al (2020a), a short-lived increase in plume flux occurred concurrently with temperature increases before the earthquakes, but both observables had returned to background levels before the M4.1 event on 27 April 2018. Hence, our Pitot-tube flux results illustrate the importance of these flux estimations for close monitoring of volcanic activity.

Our results combined with published data indicate that plume flux has decreased overall since these records began. Thus, given a lack of increased VT seismicity or other signs of sudden evolution in 2010, and in the light of the errors discussed above, it seems that the values reported in Gaudin et al (2016) are anomalous. In order to provide a better comparison with the present study, we attempt to re-estimate the 2010 plume fluxes given the available data from this period (Allard et al 2014; Gaudin et al 2016). We suppose that, despite possible overestimation, the ratio of CS/TAS fluxes was correctly established in 2010 and that the relative standard error will remain unchanged. Thus scaling with the 2005 Pitot-tube data, for 2010 we find steam fluxes of  $5.3 \pm 1.1$  kg/s for CS and  $6.2 \pm 2.2$  kg/s for TAS, and heat fluxes of  $13.0 \pm 2.6$  MW for CS and  $15.2 \pm 5.4$  MW for TAS (Table 2).

We recalculate the thermal spring fluxes for 2010 based on an interpolation of the thermal springs time series the date of the aerial survey (see Fig. 5). These new values for the thermal springs fluxes are, in general, not appreciably different to the present values, except for TA which we find to be about half the reported value for 2010. Additionally, this process does allow us to calculate fluxes for PR, absent in the 2010 survey. Consequently, we find that the total heat flux in 2010 was likely to have been around 30 MW of which the fumarolic contribution was approximately 28 MW, or 95% of this total (see Table 2). This compares to the estimate of 98% by Gaudin et al (2016).

The starkest change in the past decade is an increase in ground heat flux by greater than an order of magnitude (see Table 2), reflecting the appearance of the strong and widespread thermal anomaly in the ZFNN area in particular. Indeed, the ground heat flux has increased from  $1.2 \pm 0.3$  MW of a total budget of  $29.8 \pm 8.3$  MW (4%) to  $7.6 \pm 1.1$  MW of  $36.5 \pm 7.9$  MW (21%). Nevertheless, in terms of heat flux density across the entire edifice, the present-day mean value of  $403 \pm 24$  W/m<sup>2</sup> is slightly higher than the  $337.1 \pm 196.6$  W/m<sup>2</sup> estimated in 2010. Using the reworked values from Gaudin et al (2016), fumarole heat flux has decreased at CS and TAS, decreasing from a total of  $28.2 \pm 8.0$  MW to  $17.8 \pm 4.5$  MW (see Table 2 and Fig. 6). However, if

578 we also include G56, our reanalysis shows that the overall change in output is insignificant and,  
579 along with the appearance of NAPN and NPE fumaroles (which contribute negligibly to the total  
580 budget), highlights a spreading of fumarole output over the dome. Heat transport at the thermal  
581 spring sites has increased from 0.4 to 0.6 MW. Taken together, these findings are suggestive of the  
582 edifice becoming more fractured and porous over time, allowing some of the hydrothermal fluids  
583 to escape via different pathways. The increase in porosity and fracturing may be a result of rock  
584 dissolution and weakening (e.g. Pola et al 2012; Wyering et al 2014; Heap et al 2015; Mordensky  
585 et al 2019) as a result of hydrothermal alteration, mainly by acid-sulphate fluids (Salaün et al 2011).  
586 An increase in porosity and fractures is coherent with the increased opening rate of fractures as  
587 well as the GNSS radial horizontal displacements of 3–10 mm/year (Moretti et al 2020a) and is  
588 corroborated by the appearance of mineralised deposits at the base of the dome in the upper  
589 northern branch of the Matylis river and in the Breislack area (Fig. 1b).

590 Given recent unrest events, these results are coherent with recent petrological analysis of the  
591 volcano's last major magmatic eruption which place a shallow magma reservoir at 5–8 km depth  
592 (Pichavant et al 2018). Furthermore, the interpretation of the geochemistry of emitted fluids (Ville-  
593 mant et al 2014; Moretti et al 2020a,b) infers a contribution of deep magmatic gases that trigger  
594 periodic transitory heating and pressurisation of the deep hydrothermal system near the critical  
595 point of water at 2–3 km below the summit. Despite a lack of further evidence for the ascent of  
596 magma to very shallow depths, it is important to keep in mind from a hazard perspective, that  
597 larger, more frequent, or more intense transitory pulses of hot magmatic gases could exceed the  
598 buffering capacity of the hydrothermal system bringing the overall system to critical conditions  
599 compatible with phreatic/hydrothermal eruptive unrest. Consequently we must expect that the  
600 extent and magnitude of thermal anomalies and diffuse degassing on La Soufrière will continue to  
601 increase over time and, indeed, this is evident in aerial images (cf. Figs. A.1 and A.3, supplementary  
602 material).

### 603 Comparison with other hydrothermal systems

604 La Soufrière's total heat budget is on par with other those of other volcanoes hosting major  
605 hydrothermal systems. For example, the heat output at Vulcano (Italy) was estimated at 10–  
606 12 MW from combining ground-based radiometer and ASTER measurements (Mannini et al 2019).  
607 At Whakaari, heat output estimated using crater floor soil CO<sub>2</sub> degassing as a proxy was found to  
608 be  $20.0 \pm 2.5$  MW (Bloomberg et al 2014). Our value is somewhat smaller than Nisyros (Greece),  
609 Ischia or Campi Flegrei (both Italy) which release in the range of 40–100 MW (Chiodini et al 2005).  
610 However, as noted by Harvey et al. (2015), whilst total heat budget is helpful for following the  
611 temporal evolution of an individual volcano, a more useful metric for comparing between systems is

the heat flux density as, in many of these cases, the major component of heat flux is through diffuse soil heating so larger systems naturally tend to emit more heat. The mean ground heat flux density (combining radiative and convective fluxes) for the entire La Soufrière complex is  $406 \pm 24 \text{ W/m}^2$  (Table 1) which, if we consider the total heat budget over the total heated area of  $26\,280 \text{ m}^2$ , the mean flux density of the currently active part of the La Soufrière complex climbs to  $1366 \pm 82 \text{ W/m}^2$  (Tables 1 and 2). Based on the data compiled in Harvey et al (2015) from heat flux density calculated from  $\text{CO}_2$  flux, this ranks La Soufrière amongst the world's most powerful heat producing volcanoes, well above Whakaari ( $205 \text{ W/m}^2$ ), Vulcano ( $140 \text{ W/m}^2$ , Mannini et al 2019), Campi Flegrei ( $118 \text{ W/m}^2$ ) and Nisyros ( $19 \text{ W/m}^2$ ), and roughly on par with Ischia ( $764 \text{ W/m}^2$ ). Though individual volcano-tectonic and geological settings vary from La Soufrière, Vulcano, Whakaari and Ischia are also volcanoes with a prominent dome and extensive hydrothermal systems. The larger heat flux densities seen at such volcanoes may indicate optimal steam transport to the surface along high-permeable pathways associated with dome emplacement. Ischia, in particular, has a fumarolic  $\text{H}_2\text{O}/\text{CO}_2$  ratio similar to that at la Soufrière, ( $\text{H}_2\text{O}/\text{CO}_2=147$  in 2001, Chiodini et al 2005). This indicates the dominance of heat and mass transport by water vapour generated through the interaction of hot magmatic fluids and the water table. Taken together, especially with respect to the recent evolution at the summit, these findings indicate the importance of ground heating and thermal anomalies as a precursor for unrest of volcanic sites such as La Soufrière which may be far more relevant than at caldera-type sites (e.g. Campi Flegrei or Nisyros) where  $\text{CO}_2$  degassing is far more pervasive and heat loss through the ground is dominant.

## Conclusions

La Soufrière is an andesitic stratovolcano in the lesser Antilles arc with an extensive hydrothermal system that has undergone six phreatic/hydrothermal eruptions since 1635 C.E. Here, we have concentrated on using thermal measurements to highlight the changes to the system over the past two decades which cover most of the current unrest since its onset in 1992. Direct measurements were made of the temperature and mass flux at the key fumarolic emission sites and at numerous thermal springs linked to the hydrothermal activity of La Soufrière. The ground temperature at sites showing extensive thermal anomalies was determined from airborne thermal imagery. From these and ancillary measurements for ambient conditions, we have deduced heat and mass fluxes as well as heat flux densities. We have compared and discussed our measurements in light of historic data available in the literature. Based on a reinterpretation of previously published data, we deduce that fumarolic output has proportionally decreased from 95% of the total heat budget in 2010 to 78% currently, whereas ground heating has increased from 4% in 2010 to 21% currently. The present-day convective and radiative heat fluxes in the summit area of 4.94 and 0.74 MW,

646 respectively, have increased by an order of magnitude in the past decade which is largely due to  
647 an increase in heated area and also, to a lesser degree, because of increased temperatures. The  
648 total heat flux density presently estimated at  $403 \pm 24 \text{ W/m}^2$  is greater than the 2010 estimate of  
649  $326 \pm 69 \text{ W/m}^2$ , and thus suggests an increase in thermal intensity at the summit, though these  
650 values are within the bounds of measurement uncertainty. These changes are explained partly by a  
651 spreading in fumarolic sites over the dome during the past decade but also, and more importantly,  
652 by ground thermal anomalies on the summit that have propagated significantly in recent years.  
653 Fractures on the dome along with steady horizontal radial displacements of 3–10 mm/year have  
654 been observed over the same period. The thermal spring activity has changed little in 20 years  
655 although several of the thermal springs closest to the dome have shown a steady linear increase of  
656 their temperature and heat flux rate since 2000.

657 We find that, in terms of heat flux density (heat loss per unit area), La Soufrière is amongst  
658 the most intense emitters of heat for volcanoes worldwide, and that its ranking has dramatically  
659 increased in recent years. With recent unrest events in mind, plus petrological evidence and geo-  
660 chemical analysis of magmatic fluids, we must consider that conditions with the potential to lead  
661 to phreatic/hydrothermal events currently exist at La Soufrière. Hence, La Soufrière remains the  
662 subject of continued and enhanced surveillance and research strategies to understand better the  
663 origin of unrest and track its dynamic evolution.

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## 677 **Author Contributions**

678 DEJ collected and analysed the field data, prepared the figures and wrote the manuscript. SM, RM,  
679 DG, JCK and VR also collected and analysed field data. All authors contributed in the writing  
680 and discussion of the manuscript, and consented to its submission.

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**Fig. 1:** a) Topography of La Soufrière region showing the sites (thermal springs and fumaroles) reported here. b) Zoom of the summit area. Site codes are: Cratère Sud Centre (CSC), Cratère Sud Nord (CSN), Cratère Sud Sud (CSS), Gouffre 56 (G56), Tarissan (TAS), Napoléon Est (NPE), Napoléon Nord (NAPN); Carbet-Echelle (CE), Galion (GA), Tarade (TA), Pas du Roy (PR), Bains Jaunes (BJ) and Ravine Marchand (RM3). The base DEM, hill shading and contours were calculated from an aerial LiDAR survey at 5 m resolution. Road, footpath and waterway information were obtained via the French Government’s open data platform (<https://www.data.gouv.fr/>, accessed 2020-04-16). Geological information is as presented in Lesparre et al (2012), Brothelande et al (2014), Moretti et al (2020a) and references therein.

**Fig. 2:** a) Georeferenced thermal images of La Soufrière and surroundings taken from the helicopter on 22 November 2019 between 05:40 and 06:05 local time. b) Zoom showing the summit thermal anomalies. The base map is the 2017 IGN aerial orthophoto (BDOrtho). The thermal images are shown in greyscale, where white and black denote hot and cold, respectively.

**Fig. 3:** Conceptual model of the heat flux measured by a thermal camera,  $L_b$  viewing and displayed as the “brightness” temperature,  $T_b$ . The incoming heat fluxes (left) from solar radiation,  $L_{\text{sol}}$ , and from the atmosphere,  $L_{\text{atm}}$ , are reflected from the surface in proportion to the surface albedo,  $\alpha$ , and  $1 - \epsilon$ , respectively, where  $\epsilon$  is the emissivity of the surface. For a long-wave infrared sensor such as the thermal camera used here,  $\alpha \approx 1 - \epsilon$ . The emitted radiation of the surface,  $L$  (which depends on surface temperature,  $T$ , though the Planck function,  $P$ ) is added to these reflected fluxes which arrive at the camera having been transmitted through a mixture of atmospheric and volcanogenic gases at temperature,  $T_g$ , and having transmissivity,  $\tau$ .

**Fig. 4:** Time series of the summit fumarole fluxes since the last quarter of 2017 as estimated from Pitot tube and MultiGAS data. Steam fluxes are shown in a) and heat fluxes in b). Vertical grey bars indicate the record of VT earthquakes with magnitude  $> M2.0$  (including felt VTs), which are limited to a sequence in early 2018. To aid interpretation of the Pitot tube data, we used a linear smoothing function to produce the dashed curves. The MultiGAS data shown (filled symbols) incorporate the 35% increase in steam flux due to condensation of vapour within the plume. The reworked MultiGAS data correspond to the  $\text{CO}_2$  flux multiplied by the  $\text{H}_2\text{O}/\text{CO}_2$  ratio determined from Giggenbach bottle analyses.

**Fig. 5:** Mass flow rate, water temperature and heat flux for the thermal springs monitored by the OVSG for the period from 2000–2020. See text for site codes. The colour code for each site is identical between plots and dashed lines show linear trends for the GA, TA and PR sites.

**Fig. 6:** Long term temporal evolution of the fumarole fluxes from 2005 to present. In addition to the present dataset, we also plot Pitot-tube data from March–May 2018 (Moretti et al 2020a) and 2005, along with thermal camera data from 2010 (both Gaudin et al 2016, MultiGAS data from 2006, 2012 (Allard et al 2014), and reworked data from 2016–2017 (Tamburello et al, 2019). The arrow labelled “True fluxes?” indicates our reworking of the 2010 thermal camera data. Grey shading indicates the period covered by the present dataset, as shown in Fig. 4.

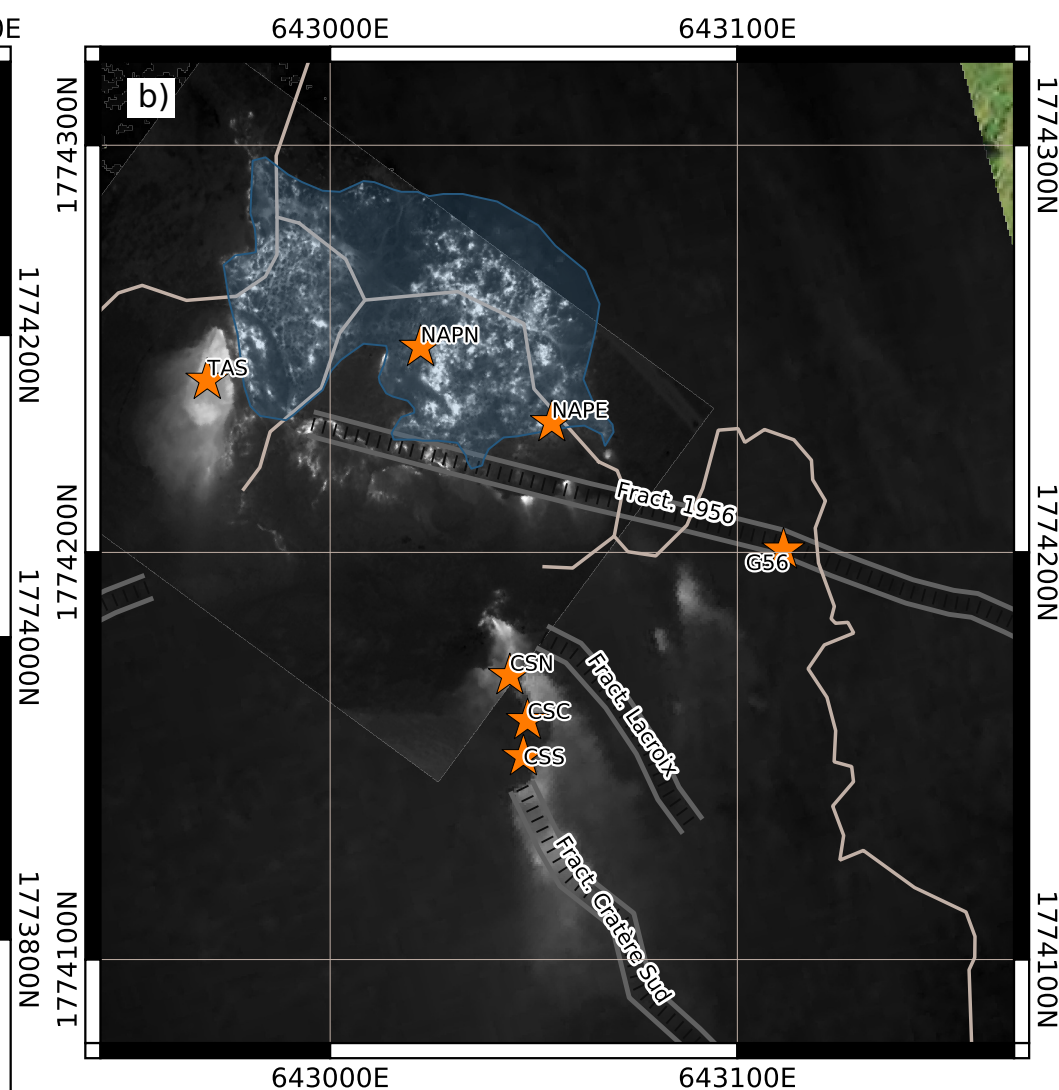
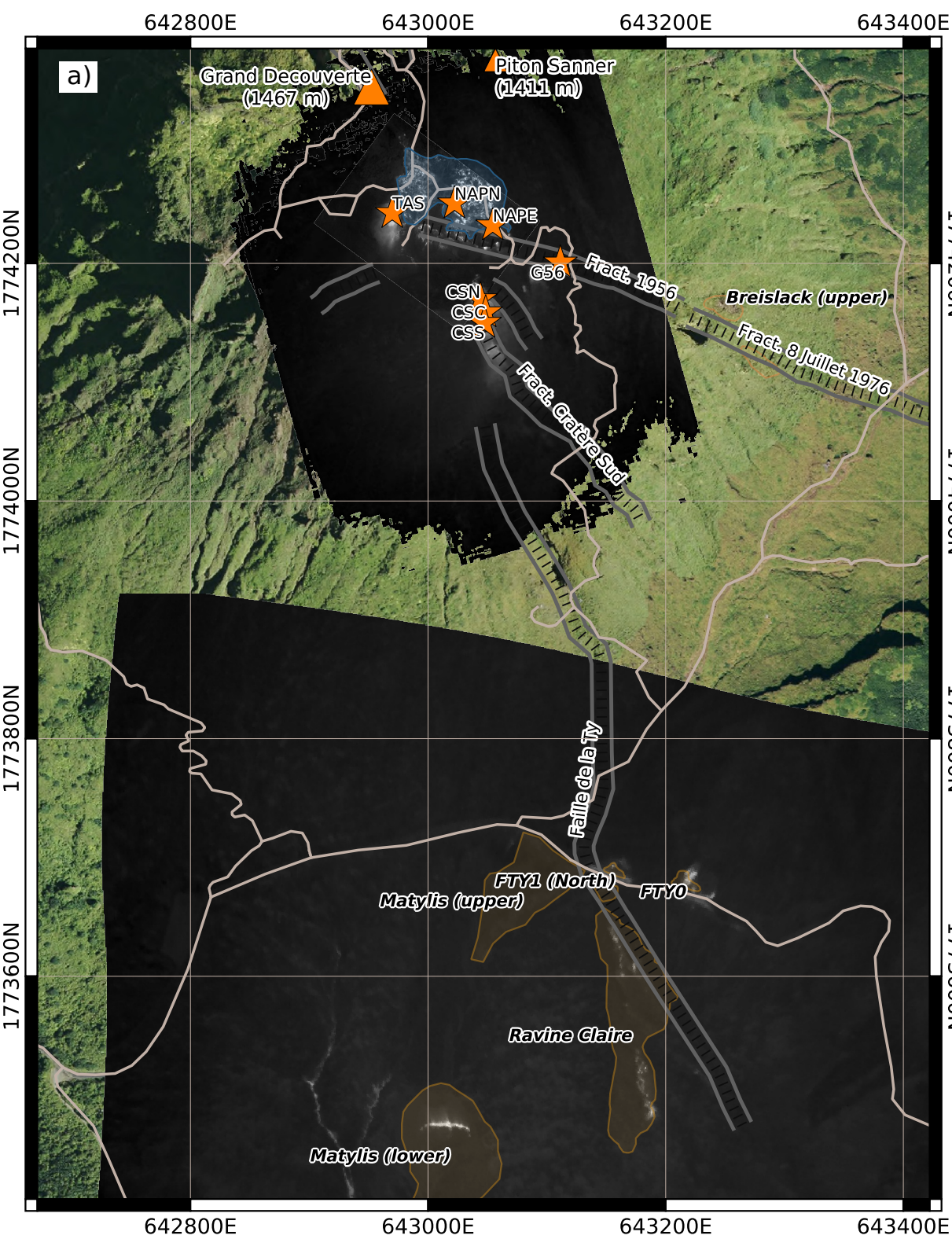
**Table 1:** Comparison of radiative and convective fluxes, and heat flux density for the ground-heated sites around La Soufrière from 2019 aerial imagery. In the right-hand column, mean values are given for the flux densities, and total values for all other quantities.

	Summit	Matylys (lower)	Ravine Claire	FTY0	FTY1	Total/Mean*
Rad. flux density, $q_{\text{rad}}/[\text{W}/\text{m}^2]$	52	76	50	71	62	54.59
Conv. flux density, $q_{\text{conv}}/[\text{W}/\text{m}^2]$	350	455	310	432	383	351.16
Total heat flux density/ $[\text{W}/\text{m}^2]$	403	531	360	504	446	405.75
Radiative flux, $Q_{\text{rad}}/[\text{MW}]$	0.74	0.12	0.40	0.10	0.07	1.43
Convective flux, $Q_{\text{conv}}/[\text{MW}]$	4.94	0.74	2.49	0.63	0.42	9.23
Total ( $Q_{\text{rad}} + Q_{\text{conv}})/[\text{MW}]$	5.67	0.87	2.89	0.74	0.49	10.66
Heated area/ $[\text{m}^2]$	14070	1630	8010	1460	1100	26270

**Table 2:** Comparison of the mass and heat flux estimates in 2010 (Gaudin et al, 2016) and for 2020 (present study). Values in parentheses are reworked fluxes which, for 2010, are based on the likely evolution of fumarole fluxes (see Fig. 6) and the thermal springs data (Fig. 5) and for the present-day fumarole estimates are the MultiGAS traverse results scaled to the Pitot-tube measurements. Dashes indicate that no data was acquired at that site. \*Encompasses RC and Matylys.

Flux	Year	Fumaroles				Ground thermal anomaly				Hot springs				Total
		CS	G56	TAS	Total	Summit	FTY	Flanks*	Total	GA	PR	TA	Total	
Mass [kg/s]	2010	$19.5 \pm 4.0$ ( $5.3 \pm 1.1$ )	-	$22.8 \pm 8.1$ ( $6.2 \pm 2.2$ )	$42.3 \pm 12.1$ ( $11.5 \pm 3.3$ )	-	-	-	-	2.5 (2.51)	- (0.44)	1.5 (1.61)	4.5 (4.56)	$46.8 \pm 12.1$ ( $16.06 \pm 3.3$ )
	2020	$4.9 \pm 0.5$	( $2.1 \pm 0.8$ )	( $2.8 \pm 1.1$ )	( $9.8 \pm 2.5$ )	-	-	-	-	3.35	0.44	1.61	5.40	( $15.2 \pm 2.5$ )
Heat [MW]	2010	$48.0 \pm 9.8$ ( $13.0 \pm 2.6$ )	-	$56.1 \pm 19.9$ ( $15.2 \pm 5.4$ )	$104.1 \pm 20.7$ ( $28.2 \pm 8.0$ )	$0.2 \pm 0.1$	$1.0 \pm 0.2$	-	$1.2 \pm 0.3$	0.3 (0.25)	- (0.02)	0.2 (0.12)	0.5 (0.39)	$105.8 \pm 21.0$ ( $29.8 \pm 8.3$ )
	2020	$15.3 \pm 1.5$	( $5.5 \pm 2.2$ )	( $7.5 \pm 3.0$ )	( $28.3 \pm 6.8$ )	$5.7 \pm 0.9$	$0.6 \pm 0.1$	$1.3 \pm 0.2$	$7.6 \pm 1.1$	0.36	0.01	0.19	0.56	( $36.5 \pm 7.9$ )





Surface temperature/[°C]

