

# Ultraviolet camera measurements of passive and explosive sulphur dioxide emissions at Yasur volcano, Vanuatu

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## Abstract

Here, we present ultraviolet (UV) camera measurements of sulphur dioxide (SO<sub>2</sub>) flux from Yasur volcano, Vanuatu, for the period 6<sup>th</sup> – 9<sup>th</sup> July 2018. Yasur commonly exhibits frequent strombolian explosions alongside persistent gas release. We use the ‘PiCam’ Raspberry Pi UV Cameras (Wilkes et al., 2017, 2016) but adapted for portability and powering through solar panels. Our determined median SO<sub>2</sub> fluxes range 4.0-5.1 kg s<sup>-1</sup>, with maximum determined errors of -10.6% to +14.7%, including errors from gas cell calibration drift, plume direction and distance, and plume velocity. This work highlights the use of particle image velocimetry for the use of plume velocity determination, which was deemed preferable to typically used methods of cross-correlation and optical flow because of ability to operate across a range of plume conditions, and hence produced more reliable fluxes. We calculate SO<sub>2</sub> masses for strombolian explosions of 8 to 81 kg (mean of 32 kg) and through use of a simple statistical measure using the moving minimum we estimate that passive degassing is the dominant mode of degassing at Yasur averaging 69% of gas released. Our work serves to further highlight the utility of UV cameras measurements and the benefit of the multiple camera approach. Our work also adds to our inventory of gas-based data for strombolian explosions globally.

## Highlights

- Long time series data collected using a portable solar chargeable UV cameras.
- Particle image velocimetry (PIV) used for plume velocity measurements.
- SO<sub>2</sub> fluxes range median values of 4.0-5.1 kg s<sup>-1</sup>.
- SO<sub>2</sub> masses of strombolian explosions range 8 to 81 kg (mean 32 kg).

## 1. Introduction

Strombolian volcanism is one of the more common forms of basaltic explosive activity globally, associated with the rapid ejection of hot pyroclasts from a vent in a single impulsive event (Blackburn et al., 1976; Taddeucci et al., 2015), with event frequencies ranging from seconds to minutes (Pering and McGonigle, 2018). Volcanoes with frequent strombolian activity include: the archetypal Stromboli, Italy (Patrick et al., 2007; Ripepe et al., 2002); Pacaya, Guatemala (Battaglia et al., 2018; Dalton et al., 2010); Erebus, Antarctica (Johnson and Aster, 2005; Sawyer et al., 2011; Sweeney et al., 2008); and the subject of this study Yasur, Vanuatu (Bani and Lardy, 2007; Kremers et al., 2013; Oppenheimer et al., 2006). Other volcanoes also known to produce strombolian activity include: Etna, Italy (Aiuppa et al., 2016; Branca and Del Carlo, 2005; Pering et al., 2015) Villarrica, Chile (Shinohara and Witter, 2005); Arenal, Costa Rica (Garcés et al., 1998; Szramek et al., 2006); Batu Tara, Indonesia (Gaudin et al., 2017a; Laiolo et al., 2018); and Shishaldin, USA (Vergnolle et al., 2004).

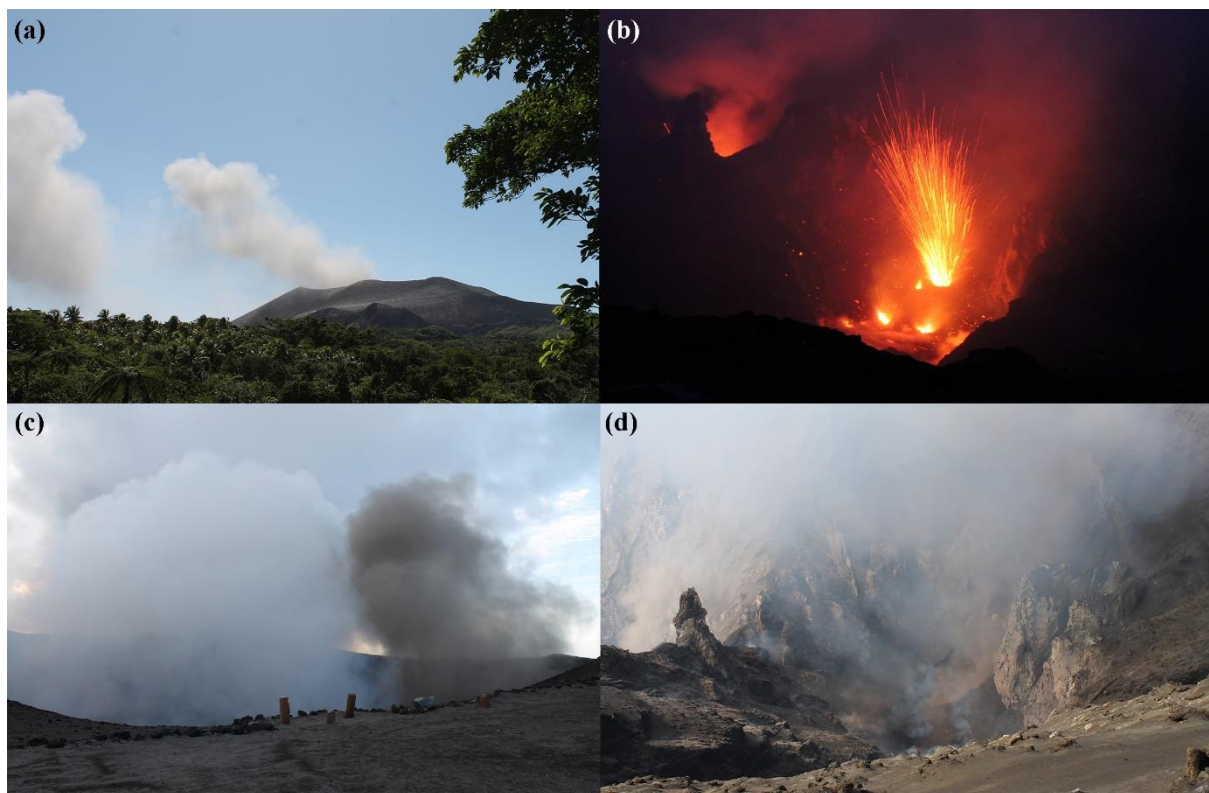
Classically, this style of behaviour has been related to the ascent from depth of elongated and over-pressured bubbles, which expand in length rapidly towards the surface, termed gas slugs (Taylor bubbles) (Del Bello et al., 2012; James et al., 2008; Seyfried and Freundt, 2000; Taddeucci et al., 2015). However, recent research has suggested that causal mechanisms may be far more diverse (Barth et al., 2019; Suckale et al., 2016), and that the presence of crystal-rich layers in the magmatic column is important to the formation of a strombolian explosion. To test these hypotheses, it is useful to investigate the spectrum of strombolian activity at sites, such as Yasur, where it is typical. In addition, recent studies have highlighted the importance of eruption frequency in determining the behaviour of ascending gas slugs (Gaudin et al., 2017a) and their interaction for the classification of styles of behaviour into rapidly bursting slugs which may experience interaction with one another during ascent, through to single bursting slugs which do not interact (Pering et al., 2017; Pering and McGonigle, 2018)

There are several ways to obtain information about individual strombolian explosions, which could include seismic (Chouet et al., 2003; Ripepe et al., 2002), infrasonic (Dalton et al., 2010; Delle Donne et al., 2016; Johnson and Ripepe, 2011; Marchetti et al., 2009), thermal (Patrick et al., 2007; Ripepe et al., 2002), and gas-derived (McGonigle et al., 2009; Pering et al., 2015; Pering et al., 2016; Tamburello et al., 2012) data. Here, we focus on gas emissions using the ultraviolet (UV) camera, a frequently used technique for the measurement of gas release from persistently outgassing volcanoes (McGonigle et al., 2017; Pering et al., 2019a). The UV camera is able to resolve high-time resolution fluctuations in the release of sulphur dioxide (SO<sub>2</sub>) emissions, and in tandem with a multi component gas analyser (Multi-GAS), which can measure gas ratios when placed inside a volcanic plume (Aiuppa et al., 2005; Shinohara et al., 2015), it is possible to estimate the total gas emissions (Pering et al., 2014). An important tool for understanding the mechanism for generation of strombolian explosions is the relative proportions of gas released during explosions and that released passively (Barth et al., 2019; Jaupart and Vergnolle, 1989, 1988; Parfitt, 2004; Suckale et al., 2016; Vergnolle and Jaupart, 1986). This 'active' to passive degassing ratio also provides information about conduit fluid dynamics (Gaudin et al., 2017a, 2017b; Pering et al., 2015; Pering et al., 2016). For example,

Tamburello et al., (2012) discovered that the most efficient mode of degassing at Stromboli was actually the passive degassing, which contributed to ~77% of gas release.

On determining gas mass from strombolian explosions (Mori and Burton, 2009; Pering et al., 2015; Tamburello et al., 2012), it is also possible, using gas ratios, to estimate total gas masses and volumes of events (Burton et al., 2007; Pering et al., 2016). These data can be applied to mathematical models, to glean further information about activity and generating mechanisms, including aspects such as slug length, explosive vigour, and categorising burst behaviour using fluid dynamics (Del Bello et al., 2012; James et al., 2009, 2008; Pering and McGonigle, 2018).

For this study we demonstrate the use of a very portable, solar-chargeable, version of the low-cost Raspberry Pi ultraviolet camera (Wilkes et al., 2017, 2016) combined with a new approach to plume velocity analysis for UV camera imagery to obtain sulphur dioxide fluxes. We illustrate the use of statistical methods to differentiate between passive and explosive gas release, and finally apply mathematical models to characterise the strombolian explosions at Yasur volcano.



**Figure 1:** Example photos of activity during the field campaign. In (a) image of the gas plume from the summit crater, large pulses are associated with explosions; (b) a night-time view with the south crater in the foreground and incandescence of the north crater in the background, several vents are visible in the south crater with one in the process of a strombolian explosion; (c) the occasional contrast between the separate gas plumes of the ash-rich strombolian explosions from the north crater and the ash-poor plume and explosions from the south crater; (d) a day-time view into the north crater, with dividing boundary on the image left-hand-side.

## **2. Yasur volcano and activity during 5<sup>th</sup>-11<sup>th</sup> July 2018**

Yasur is a basaltic volcano located on the southeast of Tanna Island, which is thought to have been predominantly persistently active for the last ~800-1400 years (Firth et al., 2014; Vergnolle and Métrich, 2016). The main volcanic edifice is a cone with a crater area of 350-450 m diameter, divided by a septum into northern and southern craters, each containing multiple active vents. During our period of measurement an ashy plume was present throughout the week, related to ash-rich strombolian explosions occurring from both craters (Figure 1). From the summit, multiple vents displaying incandescence were visible within the southern crater, each exhibiting different styles of explosive behaviour. Gas release from summit vents was constantly visible, occasionally including ‘puffing’ behaviour (Gaudin et al., 2017b, 2017a; Pering and McGonigle, 2018; Tamburello et al., 2012). The northern crater had at least two vents, but access to the northern crater rim was prevented by the frequency and location of ballistic impacts from strombolian explosions, which appeared to more ash-rich than those from the southern crater. From the southern crater we directly observed explosions from at least three vents, each exhibiting different characteristics, some with jet-like characteristics (i.e., rapid vertical movement in comparison to horizontal), highlight the potential for interaction with the conduit wall during the explosion process, i.e., the explosion (slug burst) happens within the conduit (Delle Donne and Ripepe, 2012). Another vent exhibited the parabolic transport of incandescent pyroclasts, as though an ascending bubble burst within an over-topped magma column (Del Bello et al., 2012) or within a flared geometry (Dibble et al., 2008). Interestingly these different styles of strombolian explosions also had clearly differentiable audible properties, with the latter associated with a deeper booming sound. During 8 – 9 July, explosions were frequently associated with visible shockwaves propagating through the condensed plume. Throughout the measurement period, the morphology of the crater was dynamic, with spatter and ash accumulating around vents leading to changes in the apparent size, shape, and position of vents.

A number of studies of Yasur have focused on the characteristics of strombolian activity and, in particular, its dynamism. Multi-vent basaltic volcanoes are known to exhibit vent-specific behaviours which can change through time, e.g., Salvatore et al. (2018) at Stromboli. Simons et al., (2020) discuss systematic changes in behaviour at individual vents within the southern crater at Yasur, with switching from bomb-rich (incandescent pyroclasts) through to ash-rich explosions. They also discuss conduit branching and a common source bubble (i.e., gas slug) for paired explosions from separate vents, with the potential for diverging eruption styles at the vents linked to cooling of the magma in the upper conduit. Spina et al., (2016), using infrasound, observed two distinct styles of degassing (puffing, which was observed as near-constant, and strombolian explosions), indicating that their behaviours were decoupled, while Meier et al., (2016) highlight the ash-rich and ash-poor (or bomb-rich) styles and their similarity to those of Stromboli (Gaudin et al., 2014b; Patrick et al., 2007; Ripepe et al., 2005; Ripepe and Marchetti, 2002; Taddeucci et al., 2012). Kremers et al. (2013) were able to calculate the lengths of gas slugs generating the strombolian explosions, ranging from 59 to 244 m, with a mean and median of 112 m and 103 m, respectively.

Regarding SO<sub>2</sub> fluxes, Bani and Lardy (2007), measured values ranging 2.5 to 17.2 kg s<sup>-1</sup>, with a mean of 7.9 kg s<sup>-1</sup> from April 2004 to November 2005 based on differential optical absorption spectroscopy (DOAS) traverses. Bani et al. (2012) published additional data collected between August 2007 and December 2008, with fluxes of 1.3 to 11.1 kg s<sup>-1</sup>, with a mean and median of 7.2 kg s<sup>-1</sup> and 7.1 kg s<sup>-1</sup> respectively. Métrich et al., (2011) measured a mean of 8.0 ± 3.8 kg s<sup>-1</sup> across four days of traverses in October 2007. Finally, Carn et al., (2017) show a satellite derived range of 6.8 to 23.3 kg s<sup>-1</sup> measured between 2000-2015, with a mean and median of 16.3 kg s<sup>-1</sup> and 19.2 kg s<sup>-1</sup>, respectively. With these values we emphasise the need to treat comparisons in gas flux between different periods of observations carefully, and that they may, in discrete campaigns such as presented in this study, not represent broader changes through time (or even match concurrent satellite derived estimates). Indeed, there are known issues when directly comparing satellite and ground-derived estimates of SO<sub>2</sub> outputs (Campion et al., 2012; McCormick et al., 2014).

### 3. UV Camera methods

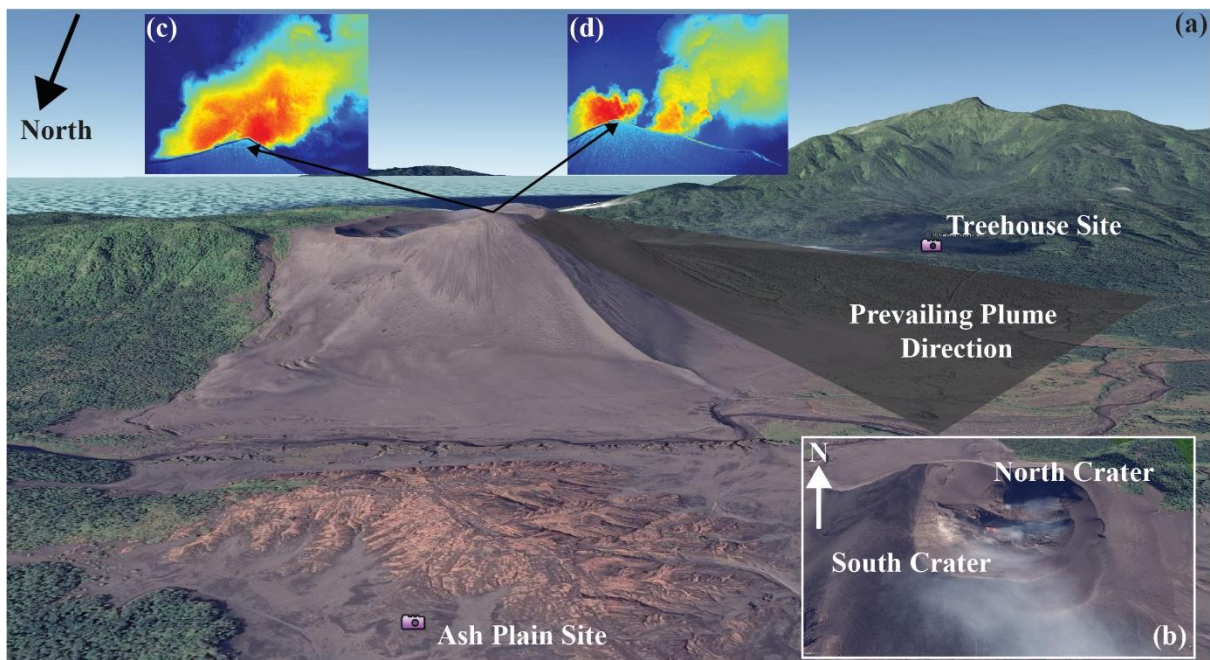
Low-cost Raspberry Pi ultraviolet (UV) camera systems (the ‘PiCam’) were used, following (Wilkes et al., 2017, 2016); however, the PiCam had been modified to include ‘PiJuice’ hardware and software, attached to the Raspberry Pi board of each camera (<https://github.com/PiSupply/PiJuice>), i.e., there are two Raspberry Pi boards and PiJuice units per camera system. The PiJuice allows the continuous supply of power via mobile phone batteries, which can be recharged using solar panels. In the field, we used both 1600 mAh and 2300 mAh batteries and found that with continuous solar charging (using 40 W solar panels for each Pi board) these were effective in running the systems a day in the field. This camera setup also omitted the incorporated GPS module which automatically syncs time on Raspberry Pi start-up, instead opting to GPS sync time manually via the command line, which the PiJuice helps to regulate via a real-time-clock. Each camera system was equipped with two Edmund Optics Inc. filters (full width at half maximum of 10 nm), centred around 310 and 330 nm, one for each lens, to account for where SO<sub>2</sub> does and does not absorb UV radiation respectively.

Two separate camera systems (Camera 1 and Camera 2) were set up viewing the plume, southwest of the summit crater from a treehouse (Treehouse site, ~1900 m to plume) at the Jungle Oasis, on 6<sup>th</sup> and 7<sup>th</sup> July, and from the ash plain (Ash Plain site, ~2300 m to plume) to the north-northwest on 8<sup>th</sup> and 9<sup>th</sup> July (see Figure 2). The UV cameras were also operated on the 11<sup>th</sup> July; however, inclement weather and a grounding plume prevented reliable data processing. All times and dates are UTC. During measurement days, plume direction varied from west to northwest, with dry and predominantly cloud-free weather (bar a brief period of rain on the 9<sup>th</sup> July). Of the five days on which measurements were attempted, we acquired four days of high quality data, with over 16 hours collected across these days. We found it was possible to operate the cameras for 6–7 hours per day. From both locations, we had the opportunity to run both camera systems simultaneously. Images were acquired at acquisition rates of 0.5 – 0.25 Hz, with collection of clear images to account for vignetting and dark images for sensor noise, for each individual sequence. We conducted frequent calibrations using gas cells with known SO<sub>2</sub> concentrations (0 ppm, 412 ppm, and 1613 ppm; with a manufacturer



quoted error of 10 %) between measurement sequences, every 1-1.5 hours at a minimum, with additional calibrations when light conditions changed more rapidly. The data were then processed as per common protocols (D'Aleo et al., 2016; Kantzas et al., 2010; Kern et al., 2014; McGonigle et al., 2017) involving aligning images; selecting a clear sky background region; and choosing a plume cross-section along which to determine an integrated column amount (ICA), before multiplying by a plume speed to calculate flux. For flux data, we determined data normality statistically with the Kolmogorov-Smirnov normality test. The data were all non-normally distributed, therefore the median has been used in further calculations. However, we quote both mean and median values.

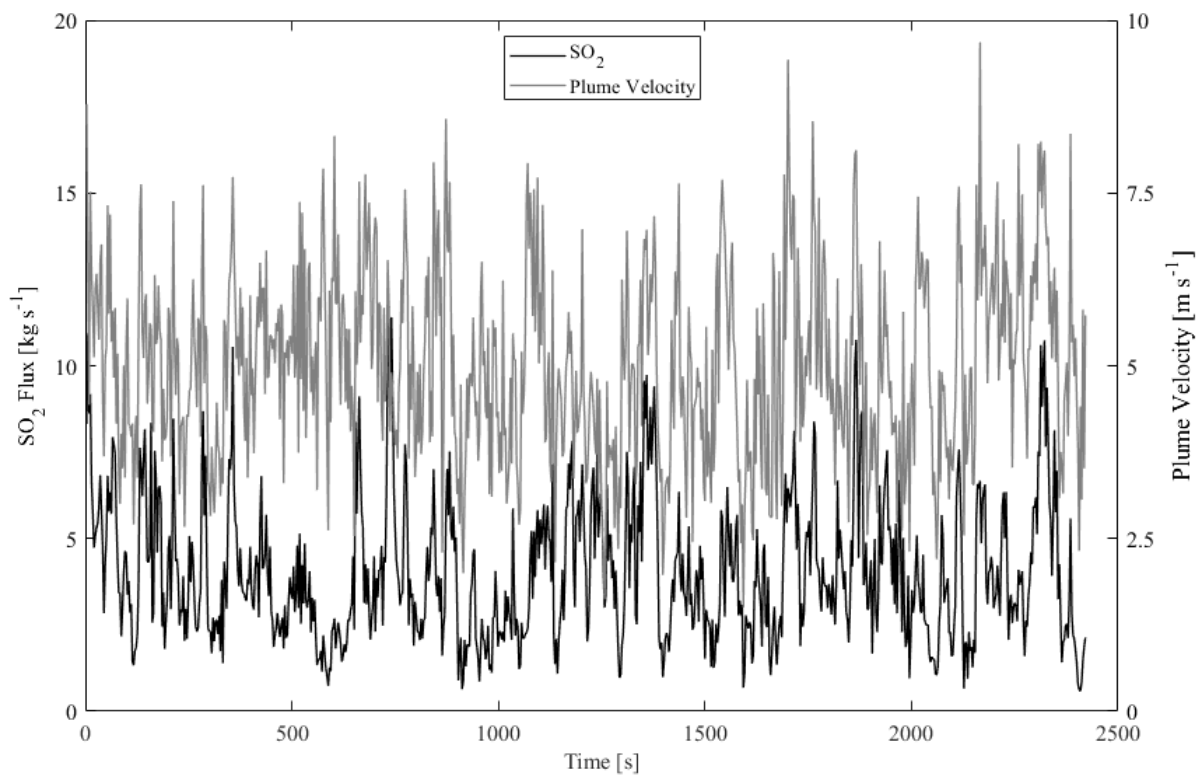
One of the goals of this study was to attempt to differentiate degassing from the vent areas. However, it was not possible to do this rigorously, or with enough confidence that we were resolving real differences in degassing, given that variable wind shear could lead to a separation of plumes or through crater derived eddying (Pering et al., 2019b; Tamburello et al., 2013), leading to incorrect identification. At times, the view from the ash plain site did allow us to identify gas pulses from separate sources, probably associated with explosions from the two craters (Figure 2c & d). Indeed, the plume predominantly appeared well-mixed on emergence from the summit crater. For example UV camera imagery see the supplementary video.



**Figure 2:** (a) An elevation-based perspective of the low summit of Yasur volcano, along with measurement positions and prevailing plume transport direction with inset (b) showing a close-up of the summit crater. In (c) a typical view of the plume with red colours representing higher concentrations of  $\text{SO}_2$  with mixing between both crater areas, and (d) shows an example of where it was possible to differentiate between emissions from both craters. Imagery is from Google Earth®.

### 3.1. Particle Image Velocimetry (PIV) for plume velocity determination.

One of the most important, and yet potentially predominantly overlooked errors with UV camera analysis is with plume velocity determination, for which three main methods are commonly used: cross-correlation (McGonigle et al., 2005; Williams-Jones et al., 2006), optical-flow (Gliß et al., 2017; Kern et al., 2015; Peters et al., 2015; Peters and Oppenheimer, 2018), and manual tracking (Ilanko et al., 2019). Indeed, the exact method used will largely be determined by plume conditions, with no single method suited to all situations. Manual tracking can be good for stable plumes travelling at slow velocities, or for calculations at greater distances from the plume, which can be difficult to resolve using cross-correlation or optical-flow, as the plume is more dilute and fewer pixels containing SO<sub>2</sub> are available. Cross-correlation could be preferred on broadly homogenous plumes that are well-mixed and present little turbulence (e.g., SO<sub>2</sub> travelling backwards relative to the plume); and optical-flow may be better with similar high velocity plumes, where there may be differential velocities through pulsed gas outputs from craters, e.g., through strombolian explosions or puffing (Liu et al., 2019; Peters et al., 2015).



**Figure 3:** Example plume velocity and SO<sub>2</sub> fluxes for a period on the 7<sup>th</sup> July 2018, clear acceleration in plume velocities are evident, these occur during strombolian explosions.

Here, we encountered difficulties using these traditional methods, and needed a more efficient mechanism than manually tracking pulses of gas in such a large dataset. Indeed, cross-correlation sometimes failed, likely a result of turbulent motion in the plume, neglecting the fact that given visible increase in gas velocity (presumably following strombolian explosions), this method is probably the least favourable. Meanwhile, a lack of structure in the plume

appeared to lead to the failure of optical flow algorithms (Gliß et al., 2017; Peters et al., 2015; Wilkes et al., 2017). We therefore introduced the use of Particle Image Velocimetry (PIV) for plume velocity determination. Previous use of PIV in a volcanic context has included tracking of lava lake velocity at Masaya (Pering et al. 2019) and similar to the pyroclast tracking velocimetry of Gaudin et al. (2014a, 2014b). Here, we used PIVlab, a user-friendly MATLAB toolbox and app (Thielicke, 2014; Thielicke and Stamhuis, 2014). PIV works by comparing image pairs in sequences and looking for differences between them through two methods; direct cross-correlation or through the correlation of Fourier transforms. Both of these are conducted on integration areas (here we used three), with decreasing size on each pass. The end result is a velocity grid for the whole plume image, similar to those produced during optical flow (Gliß et al., 2017; Peters et al., 2015). We found that using PIV we were able to pick up velocity differences in more homogenous plumes (i.e., uniform SO<sub>2</sub> distribution through most of the plume, except during strombolian explosions). PIV was used to extract velocity components along the integration line, which could then be multiplied by the corresponding column amounts to yield our SO<sub>2</sub> fluxes (see Figure 3). The PIV analyses show temporal and spatial variability in plume speed, and this heterogeneity is not captured by cross-correlation or manual tracking. We report error for PIV analysis as the distance represented by each pixel divided by the lowest image capture frequency; for the ash plain this equates to an error of  $2 \pm 0.3 \text{ ms}^{-1}$  or  $\sim \pm 15\%$ , and for the treehouse site an error of  $5 \pm 0.6 \text{ ms}^{-1}$  or  $\sim 9 \pm \%$ . These are based on typical plume speeds for each site.

### ***3.2. Estimation of a total UV Camera Measurement Error***

Error is an important aspect of all UV camera measurements. Here, we highlight the range of possible error sources, and perform additional analyses on our data related to calibration curve drift, plume orientation, and plume distance. The final determined values for error are our best possible estimate given the information and procedures conducted in-the-field, which, wherever possible, were designed to minimise error.

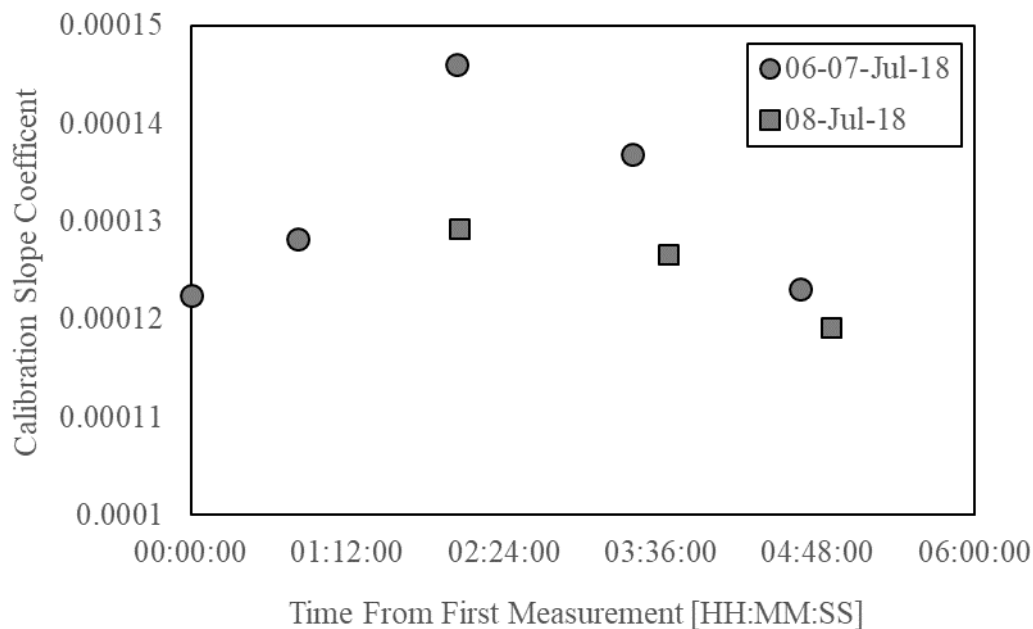
Campion et al., (2015) investigated the effects of light dilution at a range of volcanic targets and identified a broad range of  $\sim 10\text{-}60\%$  underestimation in SO<sub>2</sub> content over a range of distances (2.1 km to 6.5 km) and conditions from hazy through to very clear. Light dilution will have a larger effect during hazier conditions, which were not present during our successful measurement days. Ilanko et al., (2019), calculated that at  $\sim 10.3$  km distance (during clear conditions) SO<sub>2</sub> fluxes could be underestimated by 2.5 times, and at 4.25 km by 1.5 times (which would give  $\sim 1.18$  times [18%] at our maximum distance of 2300 m). It is important to note that light dilution estimates are specific to each measurement location and given our range of distances to the plume and clear measurement conditions we suggest that error relating to light dilution is  $< +20\%$ . We therefore take the value of  $+20\%$  as our maximum light dilution error (i.e., light dilution causes underestimation). We also note that the plume was not optically thick, except following ash-rich strombolian explosions. Unfortunately, exact errors due to ash are currently not quantifiable, but we can state that ash within the plume will likely lead to an underestimation of values (Kern et al., 2013; Tamburello et al., 2012). We attempted to



minimise this error by integrating away from the summit area. We also further note that the peaks from strombolian explosions are well defined within the resulting dataset (Figure 3).

**Table 1:** A summary of error sources, values, including short comments and total RMS error.

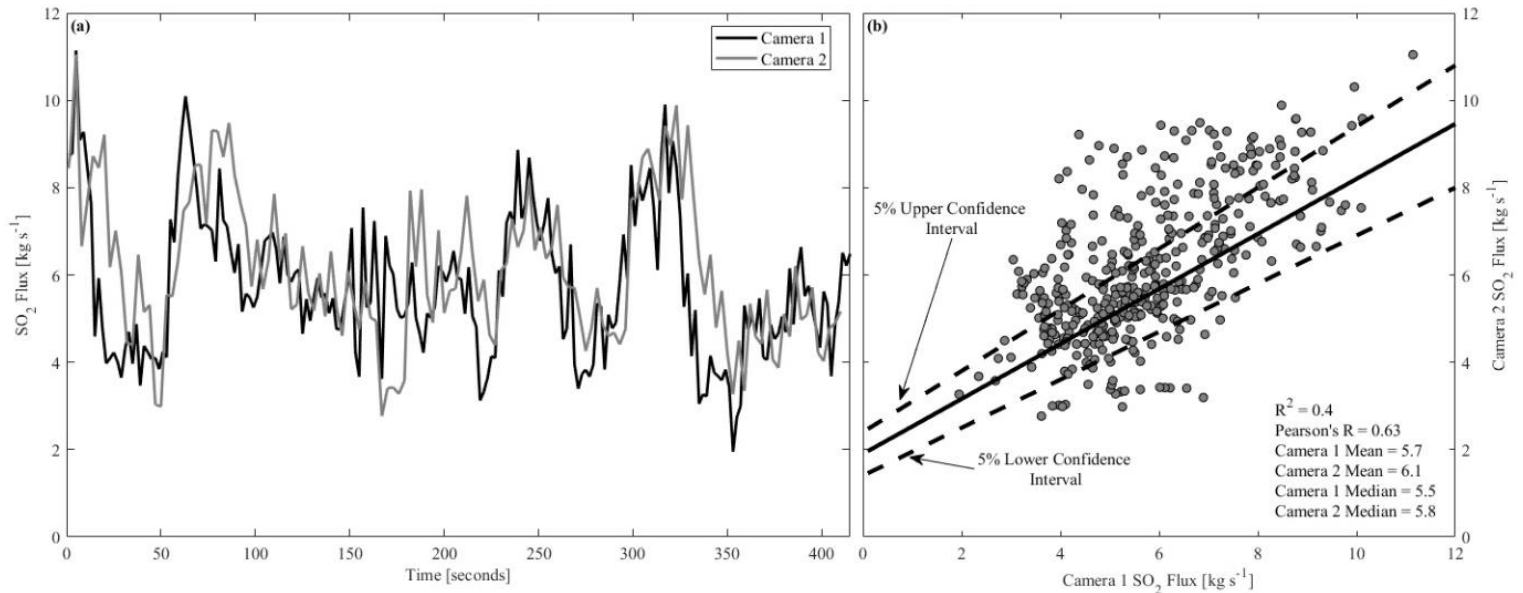
	Treehouse	Ash Plain	Comments
<b>Distance</b>	1900 m	2300 m	-
<b>Description</b>	<b>Error</b>	<b>Error</b>	-
Light Dilution	+20%	+20%	Underestimation only, low given plume proximity.
Gas Cell Concentration	±10%	±10%	Manufacturer quoted
Calibration drift	+15%	+15%	Changing calibration conditions (see text)
Plume Velocity	±9	±15%	Based on pixel size (see text)
Plume Direction	±5%	±5%	Based on coincident UV camera data
Plume Distance	±18%	±18%	Based on plume deviation of 200 m.
Ash content	-	-	Underestimation, not quantifiable
<b>RMS Error (Total)</b>	-9.4% / +13.9%	-10.6% / +14.7%	Note the higher error related to underestimation (positive error).



**Figure 4:** Example calibration slope coefficients from two days of data. Timings are from first calibration.

A gas cell calibration will alter throughout the day based on position of the sun and changing illumination as a result of background clouds, with gas-cell calibrations potentially overestimating SO<sub>2</sub> column densities by up to 60% (Lübcke et al., 2013). Figure 4 shows the change in slope coefficient throughout the day from time of first calibration (rather than using UTC), varying from  $1.22 \times 10^{-4}$  to  $1.46 \times 10^{-4}$ . When taking into account the range in slope of  $2.4 \times 10^{-5}$ , and the broad assumption (for purposes of illustration) that there is linear change between

the first point and highest point (around the time of maximum solar height) we arrive, over 122 minutes, at a value of  $1.97 \times 10^{-7}$  increase in slope coefficient per minute. This would equate to a potential error of  $\sim 0.16\%$  per minute, which expanded over an hour could become  $9.6\%$  - or, at our maximum calibration interval of  $\sim 95$  minutes, an error of  $15.2\%$ . It is possible therefore that any underlying trends below these thresholds are not differentiable from this error, i.e., an increasing or decreasing flux at a rate of  $< \sim 0.16\%$  per minute. We suggest therefore that errors from cell calibration (notwithstanding the  $\sim \pm 10\%$  manufacturer quoted cell content error) amount to a maximum of  $\sim +15\%$  for our measurement period.



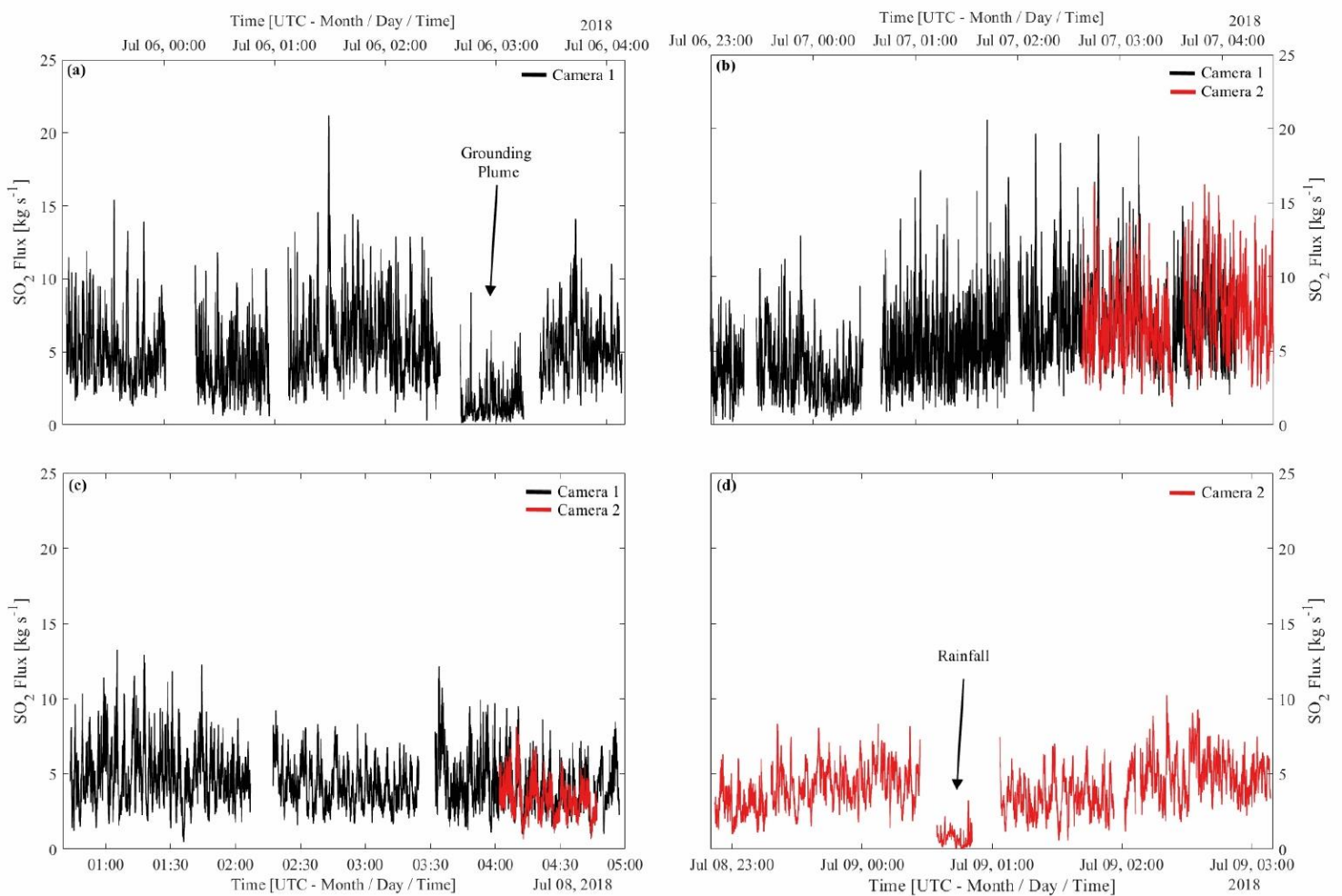
**Figure 5:** (a) Example period of overlapping data from two separately acquiring synchronous cameras at the same locations but with offset viewing angles. The data have been shifted by their maximum lag, determined using cross correlation. In (b) a linear regression model  $R^2 = 0.4$  showing the best fit between data and confidence intervals. Statistical parameters are similar, but there are differences in peaks and troughs between the two datasets.

We used fixed distances of 1900 m and 2300 m from the camera to the plume for retrievals of data from the treehouse and ash plain sites respectively. For the ash plain data, we calculated that a 100 m error in plume distance leads to a  $< 5\%$  difference in mass (with underestimation of distance corresponding to underestimation of mass), and a 200 m error in distance to  $< 9\%$  error in mass. Comparisons of the same test dataset with different velocities in PIV analyses showed variations from  $1 - 7\%$  with 100 m distance and  $5 - 11\%$  with 200 m distance. The combined effect of these distance errors on mass and velocity gives a  $7-10\%$  difference in fluxes for 100 m, and  $16-18\%$  for 200 m. We therefore take the maximum value here of  $\sim 18\%$  and apply this as a reasonable value for our entire dataset. As an example of how distance errors might affect our data, during the measurement period on 9<sup>th</sup> July the plume appeared to drift closer to the camera, which could have caused subsequent SO<sub>2</sub> fluxes to be underestimated.

Given changes in plume direction, the angle of the plume and where the integration line bisects the plume could also become important (Klein et al., 2017). To investigate this, we use overlaps

between data from two synchronously acquiring cameras (Figure 5), but with slightly different plume views and hence integration lines, simulating a potential difference from changing plume angle. Datasets were cross-correlated and shifted by lag indicated by maximum correlation to account for different transport times between integration lines. The calculated difference in flux retrieval, based on median values is  $\pm 5\%$ . We note here that, there will be differences in arrival times and magnitudes of peaks and troughs in datasets based on smoothing or turbulence during plume movement through the atmosphere.

As an aside, we also include data on our final fluxes (Figure 6) during periods when the plume grounded and during heavy rainfall. During these periods, median  $\text{SO}_2$  fluxes were underestimated by  $\sim 4.3$ - $4.4$  and  $5.6$ - $7.3$  times, respectively, based on median values of measurements either side.



**Figure 6:** In (a) through (d) all measurement data during the observation period. Also highlighted are periods of grounding plume in (a) and rainfall in (d).

## 4. Results and Discussion

### 4.1 SO<sub>2</sub> Fluxes and Manually Derived Explosion Masses

Time series gas fluxes are shown in Figure 6, with a summary of daily statistics in Table 1. The median flux across the four days of measurements was 4.5 kg s<sup>-1</sup> and the mean was 4.9 kg s<sup>-1</sup>, reflecting the peaks in SO<sub>2</sub> flux associated with frequent strombolian explosions. This corresponds to a daily median and mean of 389 and 423 t·d<sup>-1</sup>. Daily statistics are given in Table 1: median SO<sub>2</sub> fluxes (calculated using the camera providing the bulk of each day's data) were between 4.0 – 5.1 kg s<sup>-1</sup>. The means for each day were 4.1 to 5.5 kg s<sup>-1</sup>. The timeseries data are suggestive of gradual changes in background gas release trends over several hours, but it is not clear whether these are real or a product of error. A shift in activity is, however, suggested by the observation of large strombolian explosions with visible ballistics and shockwaves, particularly on 8<sup>th</sup> and 9<sup>th</sup> July, when lower fluxes were measured.

**Table 2.** A summary of measurement durations and SO<sub>2</sub> flux statistics for our measurement period.

Date (UTC)	05-06/07/18	06-07/07/18	08/07/18	08-09/07/18
Time series duration (hh:mm)	4:15	04:42	03:54	03:33
Total time (hh:mm)	05:01	05:31	04:14	04:17
Mean (kg/s)	5.2	5.5	4.5	4.1
Median (kg/s)	4.7	5.1	4.2	4.0

Explosion masses were calculated following the method of Tamburello et al., (2012), by integrating beneath the explosive pulse and summing the total SO<sub>2</sub> released. Although, a key difference here is that strombolian explosions were not visible within the imagery (i.e., vents were at depth within the crater). We use two methods to determine explosion occurrence within the UV camera imagery, primarily when gas pulses in the camera images are seen to originate and visibly accelerate from the edge of the summit crater (see Figure 1a), but also where explosion traces show the characteristic coda detailed in Pering et al., (2016). Note here that there is therefore an underestimation in the number of explosions, however, their calculation is useful for comparison to literature values, and Table 3 shows a summary. Overall, we extracted SO<sub>2</sub> masses for 135 explosions, across the five UTC dates. We show an increasing mean mass from 6<sup>th</sup> to 9<sup>th</sup> July, which corresponds to visual observations of more powerful explosions on 8<sup>th</sup> and 9<sup>th</sup> July.

The range of SO<sub>2</sub> explosion masses at Yasur are in the range of 8 – 69 kg (mean 32 kg) and similar to those of Tamburello et al., (2012) at Stromboli with a range of 2 – 55 kg (mean of 20 kg); but higher than those at Etna during mild strombolian activity determined by Pering et al., (2015) on Etna, with a range of 0.1 – 14 kg. Woitischek et al., (In Review) present gas ratio data (SO<sub>2</sub>, H<sub>2</sub>S, H<sub>2</sub>O and CO<sub>2</sub>) derived from a combined Fourier transform infrared spectroscopy (FTIR) and Multi-GAS study, from a study period following our observations, with a brief overlap. The authors separated passive from explosive molar ratios, determining active (strombolian) ratios of: CO<sub>2</sub>/SO<sub>2</sub> = 2.85 ± 0.17; H<sub>2</sub>O/SO<sub>2</sub> = 315 ± 71.8; SO<sub>2</sub>/HCl = 1.6 ± 0.22. From these we can estimate total slug masses which are highlighted in Table 3. Our

mean total gas mass based on mean active ratios indicate a mean of ~2960 kg and range of 910 – 5940 kg for strombolian explosions at Yasur. These compare with a range of 170 – 1674 kg for Pacaya (Dalton et al., 2010); whilst at Stromboli values range 44 to 238 kg according to Barnie et al. (2015) and 2 to 1425 kg as determined by Delle Donne et al. (2016).

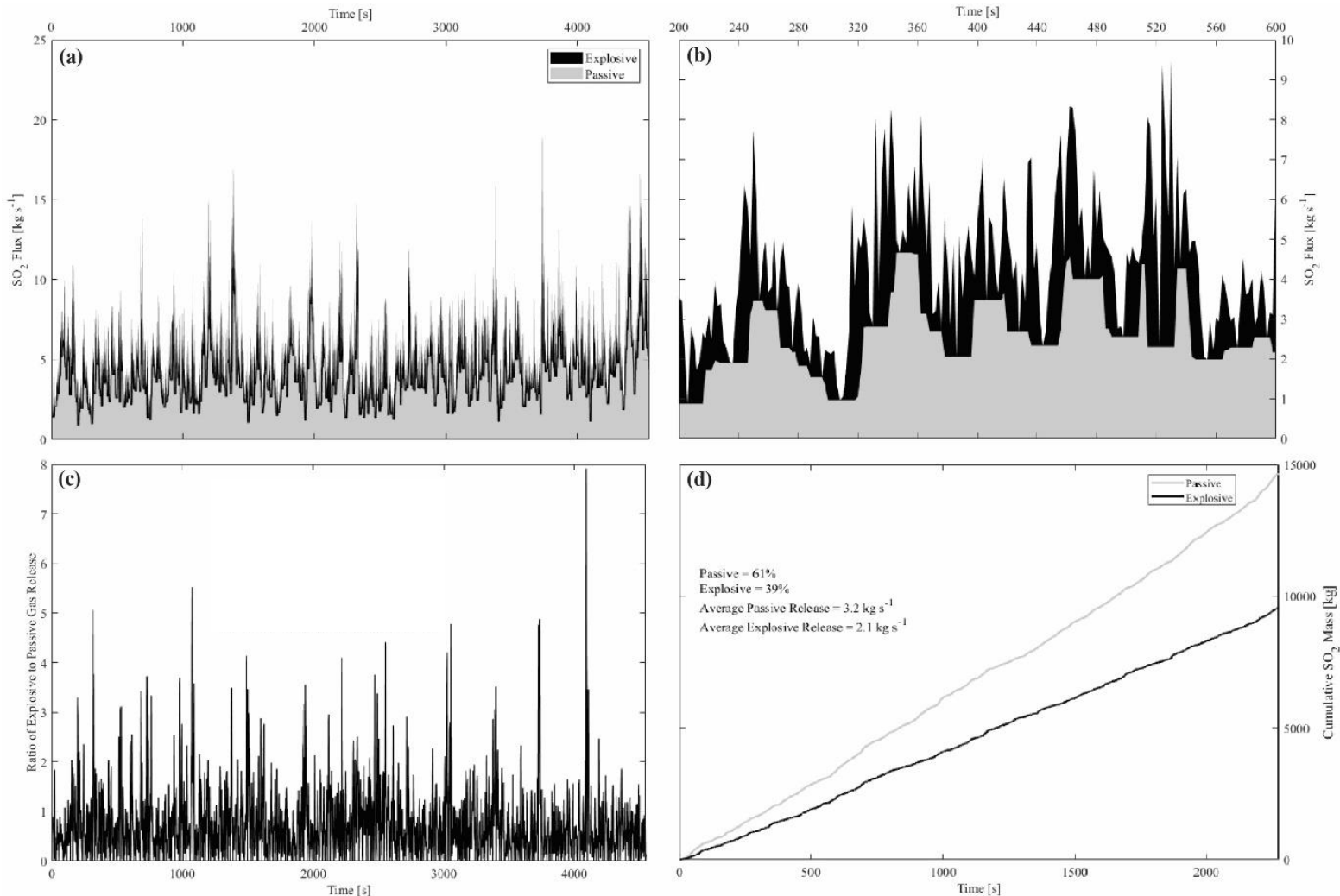
**Table 3:** A breakdown of data related to SO<sub>2</sub> explosion masses measured on a data by data basis. Also displayed are our passive to explosive percentages. Lower and Max ratios refer to the ranges indicated in determined active molar ratios by (Woitischek et al., In Review; Table 3 - CO<sub>2</sub>/SO<sub>2</sub> = 2.85 ± 0.17; H<sub>2</sub>O/SO<sub>2</sub> = 315 ± 71.8; SO<sub>2</sub>/HCl = 1.6 ± 0.22).

<b>Date</b>	<b>05/07/2018</b>	<b>06/07/2018</b>	<b>07/07/2018</b>	<b>08/07/2018</b>	<b>09/07/2019</b>	<b>Total</b>
Explosions Counted	8	43	39	36	9	<b>135</b>
<b>Data</b>	<b>05/07/2018</b>	<b>06/07/2018</b>	<b>07/07/2018</b>	<b>08/07/2018</b>	<b>09/07/2019</b>	<b>Mean</b>
SO <sub>2</sub> Min (kg)	10.2	8.9	8	12	10	<b>9.8</b>
SO <sub>2</sub> Mean (kg)	26.9	22.1	27	39	45	<b>32.0</b>
SO <sub>2</sub> Max (kg)	64.1	44.9	62	81	69	<b>64.2</b>
Passive %	66	64	70	78	68	<b>69</b>
Explosive %	34	36	30	22	32	<b>31</b>
<b>Lower Ratios</b>	<b>05/07/2018</b>	<b>06/07/2018</b>	<b>07/07/2018</b>	<b>08/07/2018</b>	<b>09/07/2019</b>	<b>Mean</b>
Total – Min (kg)	712	625	575	822	721	<b>691</b>
Total – Mean (kg)	1884	1550	1892	2713	3164	<b>2241</b>
Total – Max (kg)	3020	3148	4370	5678	4813	<b>4206</b>
<b>Mean Ratios</b>	<b>05/07/2018</b>	<b>06/07/2018</b>	<b>07/07/2018</b>	<b>08/07/2018</b>	<b>09/07/2019</b>	<b>Mean</b>
Total – Min (kg)	940	824	759	1084	952	<b>912</b>
Total – Mean (kg)	2486	2046	2497	3579	4175	<b>2957</b>
Total – Max (kg)	5929	4153	5767	7493	6350	<b>5938</b>
<b>Max Ratios</b>	<b>05/07/2018</b>	<b>06/07/2018</b>	<b>07/07/2018</b>	<b>08/07/2018</b>	<b>09/07/2019</b>	<b>Mean</b>
Total – Min (kg)	1167	1024	942	1347	1182	<b>1132</b>
Total – Mean (kg)	3088	2541	3102	4446	5186	<b>3673</b>
Total – Max (kg)	7365	5159	7163	9307	7888	<b>7376</b>

#### 4.2. Simple Statistical Separation of Passive and Explosive Degassing

Others have looked at the proportions of explosive to passive release during strombolian explosions on Stromboli (Tamburello et al., 2012) and Etna (Pering et al., 2015). Here, we attempt to expand on this by using a simple statistical measure involving the moving minimum, to estimate the passive release of gas through time, which when subtracted from total flux provides an approximate estimate of passive vs. explosive release. On an example period (Figure 7) we highlight the moving minimum, which is set to a window size of 20 s, which is generally the characteristic timeframe of large peaks and troughs associated with strombolian

explosions (Pering et al., 2016). In this instance a moving minimum over this window proved best, given the higher frequency of explosive events, however, with a greater timeframe between events, the moving median may be a better measure. We also prefer this estimation technique over using our estimated SO<sub>2</sub> masses, given that we were selective of strombolian explosions. Commonly this is achieved manually, but simple automated techniques such as this could be valuable in monitoring activity from strombolian explosion producing volcanic systems.



**Figure 7:** (a) Separation of passive and explosion gas release for a period on the 7<sup>th</sup> July; (b) a zoomed illustration of the simple statistical model, showing oscillation in background passive degassing overtopped by explosive contributions; (c) the ratio of passive to explosive degassing; and (d) a cumulative plot showing passive and explosive proportioning of gas release.

Day-by-day estimates are detailed in Table 3, with a mean of 69% passive to 31% explosive. These are in line with those measured at Stromboli, 77% passive to 23% explosive (termed active which also includes puffing); and Etna, 67% passive to 33% explosive (Pering et al., 2015). These datasets serve to illustrate the dominant role of passive degassing at volcanoes with strombolian eruptions. On the 8<sup>th</sup> July we calculated a higher passive contribution at 78%.



At the same time, we measure higher SO<sub>2</sub> masses, but lower overall SO<sub>2</sub> fluxes. This could indicate a magmatic system that is cooling or housing a thickening plug (Polacci et al., 2012; Simons et al., 2020), so becoming less permeable for passive gas release, but also requiring a higher gas mass to drive explosions that may be more powerful, matching our visual observations.

### 4.3. Models of gas slug behaviour

Using our determined values for total slug mass, we can estimate slug lengths using the static pressure model of (Del Bello et al., 2012). We use fixed values of 2600 kg m<sup>-3</sup> and 1000 Pa s<sup>-1</sup> for density and viscosity, respectively, with an atmospheric pressure of 101,325 Pa. The only parameter we vary in the model is that of conduit diameter, which we step from 3 m to 7 m. We also use only the mean explosive ratios and masses (and not the range obtained when including error) for simplicity. Our results are summarised in Table 4. We determine slug lengths ranging 188 – 609 m (median and mean of 347 m and 366 m respectively) for a conduit diameter of 3 m, however, this reduces to 76 – 260 m (median and mean of 146 and 154 m respectively) for a conduit diameter of 7 m. Kremers et al. (2013) calculated distinctly lower values of 59 – 244 m using seismo-acoustic data, and it would therefore seem that a larger conduit diameter may be more plausible at Yasur, perhaps prior to any conduit bifurcation or splitting at shallow depths (Simons et al., 2020). It should also be noted that the H<sub>2</sub>O/SO<sub>2</sub> ratio used in this work are high and variable (Woitischek et al., In Review). As water is the gas contributing most to the mass of the slug, it is likely that our determined masses are an overestimation. In addition, others have shown that UV camera-derived masses can be overestimated compared with infrasound-derived masses (Dalton et al., 2010).

**Table 4:** A summary of slug volumes and calculated slug lengths using the model of Del Bello et al. (2012), based on mean ratios for explosive (active) degassing.

Statistic	Slug Volume (m <sup>3</sup> )	D = 3 m	D = 4 m	D = 5 m	D = 6 m	D = 7 m
Min	4286	188	139	110	90	76
Median	14055	347	259	205	170	146
Mean	15556	366	272	217	180	154
Max	42337	609	455	364	303	260

## 5. Summary and Conclusions

In this work we highlighted the utility of using low-cost solar-powered Raspberry Pi UV cameras for prolonged field campaigns. We were able to continuously image the volcanic plume to yield both velocity, using a PIV method (Thielicke, 2014; Thielicke and Stamhuis, 2014), and SO<sub>2</sub> fluxes over periods of several hours per day, at temporal resolutions of up to 0.5 Hz with brief pauses for calibration. Our determined SO<sub>2</sub> fluxes with means 4.1-5.5 kg s<sup>-1</sup> (medians from 4.0-5.1 kg s<sup>-1</sup>) are within the ranges of those measured previously using ground based methods of 2.5 to 17.2 kg s<sup>-1</sup> (Bani et al., 2012; Bani and Lardy, 2007). We further highlight SO<sub>2</sub> masses of strombolian explosions of 8-81 kg, showing these to be of similar to events at Stromboli at 2 – 55 kg (Tamburello et al., 2012). By using a simple statistical measure

we estimate that passive degassing, at 69%, is the dominant mode of degassing at Yasur, compared to 31% explosive. Our brief observation period are suggestive that periods of lower gas output could lead to conduit sealing and more visibility violent explosions, however, a longer dataset would be needed to test this hypothesis substantively. By combining SO<sub>2</sub> explosion masses with gas ratios (Woitischek et al., In Review) we determined total explosion masses of mean ~910-5940 kg, which correspond to slug lengths, using the model of Del Bello et al. (2012) of ~76-260 m, if a larger conduit diameter of ~7 m is used. Smaller conduit diameters lead to far longer slug lengths ~188-600 m at 3 m diameter, larger than those estimated previously of ~59 – 244 m (Kremers et al., 2013).

This work and others (Ilanko et al., 2019) have further highlighted the utility of using a multiple UV camera approach to gas measurements at volcanoes, particularly when thinking about errors of data, and introduced the PIV method for plume flux determination, which we suggest could be robustly used across a variety of plume conditions. The data presented here represent an important addition to our arsenal of gas based information about strombolian style activity globally.

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