

VOLCANO₃ - A Miniaturized Chemiluminescence Ozone Monitor for Drone-Based Measurements in Volcanic Plumes

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15 **Abstract.** High levels of bromine monoxide (BrO) observed in volcanic plumes indicate significant catalytic destruction of tropospheric ozone (O₃) at local to regional scales. The underlying chemical mechanisms are still incompletely understood and the quantification of O₃ concentrations and their distribution in volcanic plumes remain a major challenge. Common atmospheric O₃ measurement techniques (UV absorption spectroscopy and electrochemical sensors) suffer from strong interferences, especially from sulphur dioxide (SO₂), which is low in the atmospheric background but a main constituent of volcanic plumes (ppmv levels). This problem can be circumvented by using chemiluminescence (CL) O₃ monitors, which have negligible interference with SO₂ and other trace gases commonly found in volcanic plumes. However, volcanic plume measurements with modern CL O₃ monitors are impractical because they are heavy and bulky. Here we report on the development and application of a lightweight version of a CL O₃ instrument (1.5 kg, shoebox size) that can be mounted to a commercially available drone. Besides measurements of vertical O₃ profiles over several hundred metres, we present drone-based CL O₃ measurements in the volcanic plume of Mount Etna in Italy. Within 3 km of the emitting craters we measured an anti-correlation between SO₂ and O₃ concentrations, corresponding to ozone reductions by up to 60 % in the volcanic plume with respect to the surrounding atmosphere.

1 Introduction

30 Earth's stratospheric ozone (O₃) layer absorbs the short-wavelength part of the solar ultraviolet radiation, enabling life as we know it. Besides its prominent role and abundance in the stratosphere, O₃ is an important oxidant in the troposphere. Globally, tropospheric O₃ is the most important precursor of hydroxyl radicals (OH), which drive chemical conversion and removal of many pollutants and greenhouse gases. At the same time, tropospheric O₃ itself acts as a greenhouse gas, contributing to global warming. While today surface O₃ concentrations are routinely monitored (e.g. Cooper et al., 2014), measurements of the vertical profile with high spatial and temporal resolution in the lower troposphere, in particular the planetary boundary layer are rare, yet highly desirable. Consequently, there remain major gaps in our understanding of tropospheric O₃ sources and sinks, interaction of transport processes with O₃ chemistry, and the detailed impact of O₃ on the atmospheric composition.

For a long time, O₃ has been measured with chemiluminescence (CL) techniques. In fact, nitric oxide (NO) and ethylene CL measurements of O₃ have been the standard method in the United States (US EPA, 2023) and are considered the most reliable O₃ measurement methods (e.g. Long et al., 2014, Long et al., 2021). However, unfortunately – and despite their advantages in certain applications (see Sect. 2.1, below) ethylene CL monitors are no longer produced commercially. Today and since several decades, most O₃ measurements use short-path UV-absorption spectroscopy (e.g. Dunlea et al., 2006, Williams et al., 2006), which is much easier to implement, cost-effective, and, in most environments, similarly accurate. Such devices measure O₃ concentrations by detecting the attenuation of radiation around 254 nm. Typically, absorption measurement paths are a few tens of cm and the light intensity without O₃ absorption needs to be monitored (e.g. once a minute) by passing the air through an O₃-remover (“O₃ scrubber”). But volcanic and biomass burning plumes are examples of tropospheric environments, where the applicability of UV absorption O₃ monitors is questionable, because of interfering UV absorption due to SO₂, volatile organic carbon species, fine aerosol, and mercury vapour. For example, looking at the absorption cross-sections of O₃ and SO₂, it can be recognised that the sensitivity to SO₂ is about a factor of 100 lower than that to O₃ by considering the wavelength which the O₃ maximum absorption, where many UV-monitors operate (≈254 nm). Under most atmospheric conditions these interferences (especially due to SO₂) are negligible (Kleindienst et al., 1993; Williams et al., 2006) since ambient SO₂ levels are typically comparable to or lower than O₃ levels. Therefore, UV monitors can reliably measure O₃ at most remote, urban, and industrial locations. However, when probing volcanic emissions, SO₂ mixing ratios may reach values up to several ten ppmv and therefore can exceed O₃ mixing ratios by factors of 1000 or more (usual background O₃ mixing ratios are several ten ppbv). Consequently, SO₂ typically dominates UV absorption in volcanic plumes and prohibits an accurate quantification of the O₃ UV absorption signal (Kleindienst et al., 1993; Leston et al., 2005; Williams et al., 2006). The correction of the data with simultaneously measured SO₂ (Kelly et al., 2013) or the application of selective SO₂ scrubbers (Surl et al., 2015; Vance et al., 2010), however, are difficult and – at best - introduce significant additional uncertainty.

A similar problem can occur in biomass burning plumes, typically containing high levels of hydrocarbons and particles, which can also produce false O₃ signals in the UV (Cavanagh and Verkouteren, 2001, Kleindienst et al. 1993).

Other O₃ measurement techniques, which have been used for several decades, for instance electrochemical sensors for balloon-borne profile measurements of the atmospheric background (Witte et al., 2017) show strong cross interferences to higher amounts of nitrogen dioxide and SO₂ (Schenkel et al., 1982), which also makes them unreliable in near-source plume environments.

The VOLCANO₃ instrument implements the ‘traditional’ CL O₃ measurement technique in a compact, robust, and light-weight setup. It can be mounted to a commercial drone and thereby provides accurate and interference-free O₃ measurements with meter-scale spatial resolution in all tropospheric environments. In this work we focus on the application of drone-based CL O₃ measurements in volcanic plumes with the goal of improving the understanding of volcanic plumes and their impact on the atmospheric composition.

1.1 Volcanic gases and ozone

Volcanoes are a key component of Earth’s element cycles and have an impact on their local environment, particularly the surrounding atmosphere. However, their impact can also be regional and global in scale when emissions increase during eruptions or when quiescently degassing for a long time (e.g. Marti and Ernst, 2005, von Glasow et al., 2009).

The primary volcanic gas emissions, i.e. water vapour, carbon dioxide (CO₂), sulfur species (SO₂, H₂S), and hydrogen halides (HCl, HF, HBr, HI) mix and interact with the surrounding atmosphere, creating a unique atmospheric environment (e.g. Carroll and Holloway, 1994, Kuhn et al., 2022). For example, high amounts of secondary reactive halogen species, especially bromine monoxide (BrO), have been detected in volcanic plumes (e.g. Bobrowski et al., 2003, 2007; Kern et al., 2009; von Glasow, 2010; Gliß et al., 2015; General et al., 2015). This indicates heterogeneous photochemical reaction cycles (referred to as bromine explosion, see Platt and Janssen, 1995, Wennberg, 1999), involving volcanic halogen halides, aerosol particles, and atmospheric

oxidants (mainly O₃, see Bobrowski et al., 2007; Kern et al., 2009; Jourdain et al., 2016). The reactive cycles include the catalytic destruction of O₃, which has led to the widespread assumption that O₃ levels in volcanic plumes are depleted with respect to the atmospheric background.

The conclusion that the observed amounts of reactive halogens in volcanic plumes lead to depleted O₃ levels is, however, by no means trivial. Field studies (using CL as well as short-path UV absorption instruments) have shown varying degrees of O₃ depletion across different volcanoes, in some cases up to 90% O₃ loss compared to ambient levels were reported (e.g. at Mount St. Helens, USA, see Hobbs et al., 1982). In other cases, no O₃ depletion was found (e.g. at Kilauea, Hawaii, USA, see Roberts, 2018) which was explained by low concentrations of halogens and is supported by measurements of e.g. Kern et al, 2020. Elementary calculations assuming a constant influx of O₃, a basic turbulent mixing scheme and the BrO self-reaction as a rate determining step for the O₃ destruction in volcanic plumes, suggest that the influx should largely compensate O₃ destruction. Therefore, these simple calculations predict negligible (<1%) O₃ destruction in volcanic plumes (Rueth 2023). Conversely, model studies with more evolved multiphase atmospheric chemistry mechanisms predict significant destruction of O₃ in volcanic plumes (e.g. Surl et al. 2021, Nies et al. 2025, see Supplementary Material Fig. S1) in accordance with some observations (e.g. Surl et al. 2015).

Measuring O₃ levels in volcanic plumes is challenging and often relies on substantial logistical efforts such as aircrafts or requires specific meteorological or topographical conditions to access the plume with ground-based instruments. The aim of this study is to provide a technique for reliable O₃ measurements in volcanic plumes. Building upon previous studies, this work focuses on employing gas-phase chemiluminescence (CL)-based O₃ monitors on a drone for volcanic plume measurements (Hobbs et al., 1982; Vance et al., 2010; Carn et al., 2011).

After a short summary of the CL O₃ monitor principle (Sect. 2), we introduce our small and lightweight CL O₃ monitor, VOLCANO₃ (Sect. 3). In Sect. 4, we present field measurements including planetary boundary profiles and O₃ measurements in the volcanic plume of Mt. Etna, Italy, which show significant O₃ depletion compared to the ambient atmosphere. We further discuss future technical developments (Sect. 5) and applications (Sect. 6) of VOLCANO₃.

2 The principle of CL O₃-Monitors

The operation principle of CL O₃-Monitors relies on the generation of chemiluminescent species through reactions involving O₃. By measuring the emitted photon flux, it is possible to infer the concentration of O₃. There are various CL reactions involving O₃ that are employed in CL O₃ monitors. In this study we used C₂H₄ which we further motivate below in section 2.1. Nederbragt et al. (1965) were the first to make use of this chemiluminescent reaction to determine O₃ near an accelerator and Warren and Babcock (1970) then described the construction and calibration of such a monitor.

The reaction of O₃ with C₂H₄ produces various products (detailed in e.g. Kleindienst et al., 1993, R uth, 2023), including electronically excited species that emit photons upon de-excitation. The number of emitted photons, directly measured by a photomultiplier tube (PMT), is proportional to the O₃ concentration and thus to the O₃ volume mixing ratio X_{V,O_3} . In addition to being proportional to the O₃ mixing ratio, the number of photons emitted per second is influenced by several other parameters, including the quantum yield of the reaction, the ambient temperature and pressure, as well as the concentration of C₂H₄, regulated through the C₂H₄ flow ($f_{C_2H_4}$) in ml/s (or 10⁻⁶ m³/s). In order to determine the O₃ volume mixing ratio (X_{V,O_3}), the measured photomultiplier tube (PMT) signal has to be converted according to the theoretical description and calibrated experimentally using an O₃ generator,

$$X_{V,O_3} = \frac{Y_{kb}}{Q} c_{con}(f_{C_2H_4}, p, T) \quad (1)$$

$$c_{con}\left(f_{C_2H_4}, p, T = \frac{T}{p(f_{tot} - f_{C_2H_4}) \left[1 - \exp\left(-k_1(T) \frac{f_{C_2H_4} p V_{cell}}{f_{tot} k_b T}\right) \right]}\right) \quad (2)$$

where γ denotes the number of photons generated per second, Q the detector quantum efficiency of the PMT, $k_I(T)$ denotes the reaction rate constant of O_3 with C_2H_4 in $cm^3/molec$, p is the ambient pressure in Pa, T the ambient temperature in K, V_{cell} the measurement cell volume (in this study 20 ml), f_{tot} the flow rate of the pump in ml/s (or $10^{-6} m^3/s$) with V_{cell}/f_{tot} the residence time in the cell, k_b the Boltzmann constant and $c_{con}(f_{C_2H_4}, p, T)$ is the "conversion factor" given in $10^{-6} Ks/J$ (or $Ks/Pa \cdot ml = Ks/(Pa \cdot 10^{-6} m^3)$).

Note that the quantum efficiency Q also includes the probability of a CL photon to actually reach the photocathode of the PM. We estimate this probability at about 15% for a reflectivity of the reaction chamber walls of $R_W = 0.7$. For an in-depth description of the theoretical basis see (Rüth, 2023; Bräutigam, 2022).

2.1 Selection of chemiluminescence technique

The ethylene chemiluminescence (ET-CL) reaction was selected for ozone detection in UAV applications because it provides high photon yield in the visible range ($\lambda \approx 440$ nm) and operates stably without active drying. In contrast, the nitric oxide chemiluminescence (NO-CL) reaction produces electronically excited NO_2^* , emitting primarily in the red–near-infrared region with a broad maximum around 1200 nm (Clough, 1967). These long-wavelength photons have lower energy and are detected with markedly reduced quantum efficiency by standard photomultipliers, requiring active cooling to suppress high dark current noise. Such detector assemblies substantially increase mass and power consumption, which is a critical aspect of UAV applications.

The effect of water vapour also differs fundamentally between the two chemiluminescence systems. For NO-CL, water acts as an efficient collisional quencher of NO_2^* emission, strongly reducing signal intensity and linearity (Matthews et al., 1977). For ET-CL, in contrast, water vapour slightly enhances the chemiluminescence signal through secondary excitation of formaldehyde and OH^* products, leading to a small positive bias rather than suppression (Kleindienst et al., 1993). Consequently, the humidity response of ET-CL can be accounted for in the measurement uncertainty, whereas NO-CL requires complete gas drying and thermal stabilisation to achieve reproducible sensitivity. Measurements with the VOLCANO₃ instrument are typically performed in volcanic plumes that have cooled to ambient atmospheric temperature, so that atmospheric water vapour concentration cannot exceed local saturation level. Typical H_2O concentrations range from 1000 ppmv – 20000 ppmv and are lower at high altitudes and cold atmospheric temperatures. This adds an uncertainty of O_3 measurements of not more than about ± 4 ppbv even when assuming a relatively high 10,000 ppmv water vapour content as "standard". Note that even though water vapour is the most abundant component of volcanic emissions, its mixing ratio cannot exceed saturation level.

Considering these differences in spectral emission and related detector requirements as well as the response to humidity, the ET-CL configuration provides the favourable analytical performance, stability under ambient conditions, and compatibility with compact, low-power UAV payload operation.

Although ethylene is flammable, only a very small amount of gas was used, contained in a sealed aluminum minicylinder. The instrument was usually operated in the open atmosphere. The gas volume was far below limits typically relevant for hazardous classification. Consequently, the ET-CL setup did not represent a relevant safety risk during field deployment.

3 A compact CL ozone monitor

3.1 Configuration

In Fig. 1A a simplified schematic drawing of the monitor VOLCANO₃ is shown (Bräutigam, 2022) and Fig. 1 B displays a photograph of the instrument. It has the dimensions 38 x 20 x 11 cm and weighs around 1.5 kg. VOLCANO₃ consist of the following components: 1) a reaction chamber, 2) a pump, 3) an ethylene minican 4) a photomultiplier module, 5) a circuit board and a Raspberry Pi computer and 6) a lithium-polymer battery. Ambient air enters the instrument through an aerosol filter (A:

Schematic drawing of the CL O₃ monitor setup (without the electronics). Figure taken from Bräutigam (2022). B (part A)) and is then directed through a black Teflon hose (two windings to suppress ambient light entering the measurement cell through the hose) into the measurement cell (aluminium with a volume $V = 20$ mL, Figure 1b, (part B)), where it is mixed with C₂H₄. The measurement cell is fixed to the PMT (Figure 1b, (part C)) enabling the measurement of the photons emitted by the air + ethylene mixture. The PMT (HI0493-001 from Hamamatsu Photonics GmbH) is the central part of the instrument and main reason for the weight and size reduction due to its internal high voltage supply and preamplifier electronics. In contrast to the most commonly applied CL-monitors, the PMT is operated without temperature stabilisation (rather the temperature effect is compensated during signal evaluation, see Sect. 3.2), thus significantly reducing the power consumption of the monitor to only 3 W. Ethylene is supplied from a minican ($V_{\text{bottle}} = 1$ L, maximum overpressure $p_{\text{bottle}} = 12 \cdot 10^5$ Pa bar (Figure 1b, (part E)). The C₂H₄ flow into the measurement cell is regulated via a pressure regulator and a capillary. The gas mixture leaves the instrument through the airflow generated by the pump (model G 6/01-K-LC). Ambient temperature and pressure, as well as the temperature at the PMT are also monitored in order to convert the measured signal to the O₃ mixing ratio. All relevant data (date, time, temperature, ambient pressure, output voltage of the PMT, pressure of the minican) is recorded by the microprocessor (Figure 1b, (part N)) on a USB drive (Figure 1b, (part P)) and also shown on a small display to ensure operational functionality of the device during the measurement.

3.2 CL-Monitor Characterization

3.2.1. PMT Dark current

Since the PMT used is not temperature stabilised, the dark current (measurable signal if no photons hit the photosensitive area) is a temperature-dependent variable for which a correction is required. The temperature dependence of the dark current compiled from several measurements is shown in Figure 2.

In order to subtract the dark current as a function of temperature from the data the Richardson function (see Eqn. 3) is fitted:

$$S_0(T) = a_1 \cdot T^2 \cdot \exp\left(\frac{b_1}{T}\right) + c_1 \quad (3)$$

The parameters a_1 , b_1 , c_1 are determined at regular intervals. Typical values are $a_1 = (9.9 \pm 0.3) \cdot 10^{12}$ mV/K², $b_1 = (-1.2 \pm 0.001) \cdot 10^4$ K, and $c_1 = (0.55 \pm 0.003)$ mV.

The uncertainty arising from the dark current correction is determined by the mean deviation of the measurement to the fit $S_{0,\text{mean}}$, leading to a propagation into the uncertainty of the O₃ determination of only 1 ppbv. Temperature is the dominant driver of the dark current signal, minor signal deviations are additionally corrected by an easy pragmatical solution, subtracting the $S_{0,\text{mean}}$ from each measurement. Then the dark current corrected signal S_{corr} is calculated as:

$$S_{\text{corr}} = S - S_0(T) - S_{0,\text{mean}} \quad (4)$$

Where S is the measured signal, $S_0(T)$ the temperature dependent dark current as derived by Eq. 3, and $S_{0,\text{mean}}$ the mean fit residuals from the dark current temperature correction given in mV, because the output signal of the PMT is measured in mV.

3.2.2 Calibration

The CL method is not a direct measurement technique, therefore, in order to calculate the O₃ mixing ratio from the measured PMT-signal an experimental calibration is needed, based on the theoretical considerations in Sect. 2. (Eq. 1 and 2):

$$\gamma = a_{\text{cal}} \cdot S_{\text{corr}} \quad (5)$$

where a_{cal} represents the calibration constant.

Combining the above equations (2 and 5) we obtain an equation for the O₃ mixing ratio X_{V,O_3} :

$$X_{V,O_3} = a_{\text{cal}} \cdot c_{\text{con}}(f_{C_2H_4}, p, T) \cdot S_{\text{corr}} \quad (6)$$

where c_{con} is the theoretically determined conversion factor (see Sect. 2.). Here, no calibration offset is necessary, since any potential offset in the data is taken care of by applying the dark current correction.

The CL O₃ monitor is calibrated using an O₃ generator, in our study primarily the Ozone Calibration Source Model 306 by 2B Technologies. It is a portable O₃ generator and can provide O₃ in the range of 0 to 1000 ppbv. Additionally, the O₃ generator ANYSCO type SYCOS KT-O₃/SO₂, which can provide 0 and 150 ppbv of O₃, was used. In both instruments, ambient air is used as the feed gas and ozone is generated photolytically by UV irradiation of oxygen-containing air. Before entering the photolysis chamber, the incoming air is cleaned, and in the case of the O₃ generator ANYSCO type SYCOS KT-O₃/SO₂ it is additionally dried by passage through a silica-gel drying tube.

For calibration the CL O₃ monitor is directly connected to the O₃ generator using a hose. We set the ozone mixing ratio at discrete steps (e.g. 0, ~10, ~50, ~100, ~150 ppbv) and held each step for several minutes to ensure sufficient statistical averaging. For each of these steps of constant O₃, the mean and the standard deviation of the converted signal are calculated and plotted against the sampled O₃ mixing ratios as shown in Figure 3. A linear fit is performed; its fit parameters are then used for the calculation of the O₃ concentrations. Although VOLCANO₃ demonstrated stable behaviour, it is advisable to perform a regular calibration check before each measurement campaign.

3.2.3 Detection limit and measurement uncertainty

Both, the detection limit and the measurement uncertainty are crucial characteristics of any instrument. Following the definition of Gold (2019) the detection limit (also referred to as limit of detection (LoD)) is the ‘minimum single result which, with a stated probability, can be distinguished from a suitable blank value’ (Gold, 2019, p. 399). To determine the uncertainty of the mixing ratio of O₃, error propagation is applied to Eq. 6 for X_{V,O_3} :

$$\Delta X_{V,O_3} = \left[\left(\frac{\partial X_{V,O_3}}{\partial S} \Delta S \right)^2 + \left(\frac{\partial X_{V,O_3}}{\partial S_0} \Delta S_0 \right)^2 + \left(\frac{\partial X_{V,O_3}}{\partial c_{con}} \Delta c_{con} \right)^2 + \left(\frac{\partial X_{V,O_3}}{\partial a_{cal}} \Delta a_{cal} \right)^2 + (\Delta_{H_2O})^2 \right]^{\frac{1}{2}} \quad (7)$$

Signal uncertainty, ΔS , is deduced from the standard deviation σ_S of constant O₃ periods, represented by:

$$\Delta S = \sigma_S = a \cdot \ln(b \cdot S + c) \quad (8)$$

ΔS_0 is the uncertainty from temperature-corrected dark current, approximated as the mean deviation from the measurement to the fit, around 0.5 mV. Calibration uncertainty is quantified by $\Delta a_{cal} = 24$ (hPa·ml/(mV·K))

The uncertainty from signal conversion is estimated from the mean deviation of experimental data to theoretical description, about 4%.

The fifth term in Eq. (7) accounts for the well-established positive interference of water vapour on ethylene chemiluminescence. This interference is approximately 3–4 ppbv O₃ per 10,000 ppmv H₂O. As volcanic plume measurements are conducted at ambient atmospheric conditions, water vapour cannot exceed local saturation and typically ranges between 1,000 and 20,000 ppmv.

To include this effect in the instrument uncertainty, we assume a typical value of 10,000 ppmv and thus treat the humidity bias as a constant additive systematic uncertainty, Δ_{H_2O} , given by:

$$\Delta_{H_2O} \approx 4 \text{ ppbv}$$

For a representative signal of $S = 20$ mV, corresponding to O₃ mixing ratio of 40 ppbv, the uncertainty analysis is:

$$\Delta X_{V,O_3} = 0.1 \cdot X_{V,O_3} \quad (10)$$

The detection limit, assuming $\sigma_{S,0} \approx 0.4$ mV, is approximated as:

$$X_{V,O_3,0} = 4.16 \text{ ppbv} \quad (11)$$

3.2.4 Response time

In order to determine the response time of the monitor experimentally, step changes in the O₃ mixing ratios are of interest. To circumvent the ozone generator’s response time, one can produce a consistent O₃ mixing ratio using the generator, but without

attaching the hose to the monitor. Once a stable mixing ratio is achieved, the hose can be swiftly connected and disconnected to and from the monitor, respectively. The step change recorded in this manner should reflect the response time of the O₃ monitor.

240 To these increasing and decreasing step changes exponential increases and decreases are fitted, respectively. For the calculation of the response time increases and decreases are treated equally and only fits with a R^2 larger than 0.9 are used. The experimental response time changes of the monitor, determined from 25 step changes fulfilling the above requirements, results to $\tau_{\text{exp},1/e} = 2.00 \pm 0.04$ s.

4 Field Measurements

245 4.1 Vertical atmospheric profile

Vertical atmospheric O₃ profiles in the planetary boundary layer (PBL) with a high spatial resolution are still surprisingly rare. Some O₃ profiles were taken in the frame of this study using an UAV (Matrice300 RTK) as a vehicle to transport the VOLCANO₃ monitor to different heights of the PBL. One example of our measurement results is given in Fig. 4. The measurement was carried out at the southern flank of Etna, at Piano Vettore, Italy on 16th of June 2023 and shows the increase
250 of O₃ mixing ratios from about 50 ppbv to nearly 100 ppbv for an elevation change from about 1700 m.a.s.l. to about 2500 m.a.s.l.

4.2 Measurements in the volcanic plume of Mt. Etna

As mentioned in the introduction, the CL technique has particularly unique advantages when applied to O₃ determination in volcanic plumes. Therefore, we performed a first measurement campaign at the volcano Mt Etna to investigate the O₃
255 distribution of its plume.

4.2.1 Measurement Site and conditions

Mt. Etna, located at the eastern coast of Sicily, is one of the most active volcanoes globally and is characterized by significant continuous gas emissions. Geological evidence suggests it has been active for approximately 0.6 million years. The summit area of Etna, reaching to about 3350m above sea level during the campaign had four active summit craters: Bocca Nuova (BN),
260 Voragine (VOR), Southeast crater (SEC), and Northeast crater (NEC).

Our field measurement campaign was conducted from June 5th to 18th, 2023. During this period, Etna exhibited continuous outgassing primarily at Bocca Nuova and Southeast crater, with moderate SO₂ fluxes and consistently high CO₂ levels (INGV National Institute of Geophysics and Volcanology, 2023a, b).

4.2.2 Instrumentation and data evaluation

265 The measurement campaign employed the following instruments:

- CL O₃ monitor “VOLCANO₃”
- SO₂/CO₂ sensor “little-RAVEN” (Karch et al., 2022) with 868 MHz radio link (RFDesign, approx. 3 km range), GPS module for time and position (MTK3339 Adafruit), Alphasense electrochemical SO₂ sensor (calib. range: 0-16 ppmv), CO₂ sensor (K30 FR Senseair, not used in this work), temperature, humidity & pressure sensor (BME280).
270 Total weight: approx. 300 g
- Drone “Matrice 300 RTK”, DJI, <https://enterprise.dji.com/de/matrice-300/specs>

The little-RAVEN sensor system, designed around an ESP microcontroller (ESP32 from Espressif), manages various sensors to determine SO₂, CO₂, temperature, humidity, pressure, and GPS location. It logs data onto internal memory and also transmits it to a ground station, allowing real-time localisation of the plume and confirming plume gas measurements. For this work,

275 Little-RAVEN was used as a stand-alone system which could be attached to the measurement drone to provide the
aforementioned information simultaneously with the ozone measurements. The Matrice 300 RTK drone, with a maximum
payload of 2.7 kg, was utilized for carrying the VOLCANO₃ CL O₃ monitor and the little-RAVEN system. The total equipment
weight for the measurements was approximately 15 kg, including the instruments, the drone, and the ground station consisting
of a notebook, a tripod and an antenna.

280 4.2.3 Volcanic plume measurements and results

During the campaign, a total of four drone flights through the volcanic plume were conducted. The flight paths of all four
plume measurements are shown in Figure 5 and the main findings of the four plume measurements are summarised in Tab. 1.
Flight 1) On June 13th, the VOLCANO₃ mounted on the drone sampled the BN plume (black flight path in Figure 5). The
drone started at the Barbagallo craters (the upper pyroclastic cone from the 2002 eruption), south westerly of the SEC. Fig. S2
285 displays the data obtained from this flight. The meteorological conditions were sunny with some clouds and with wind blowing
mainly from North with low wind speeds. Due to the low wind speeds the plume rose before drifting to the south.

On June 18th, three more flights were carried out:

Flight 1) The first flight on that day started south-southeast of the Barbagallo craters, north-northwest of the Cisternazza (a
subsidence crater from 1792) and the pyroclastic cone from the 2001 eruption and navigated through the SEC plume, the
290 measurement is depicted in Fig. 6 and the flight path shown in Figure 5 in blue color. O₃ and SO₂ exhibit a strong anticorrelation
with a Pearson correlation coefficient of -0.64 and an R² value of 0.41 (see Tab. 1).

Flight 2) The subsequent flight first sampled the SEC plume before measuring inside the BN plume. This measurement is
shown in Fig. S3, the flight path can be seen in red on Fig. 5.

Flight 3) The final flight on the 18th of June 2023 focused solely on sampling the BN plume, which can be seen in Fig. S4. The
295 flight path is indicated in orange on Fig. 5.

These two latter flights started from the northern rim of the Barbagallo crater. The weather was sunny, and only little wind,
mainly from the north, this meant that the plume ascended and was not pushed down. Although conditions for gas sampling
aboard drones were ideal, ground-based measurements within the exhaust plume would have been impossible on that day,
illustrating well the strength of the newly developed instrument.

300 In all of these measurements an anti-correlation of O₃ and SO₂ levels is visible. The ambient O₃ fluctuations are in the range of
5 ppbv, whereas within the plume variations of up to 60 ppbv are observed. Unfortunately, the SO₂ sensor is only able to
measure SO₂ mixing ratios of up to 16 ppmv, leading to higher values being cut off. This obscures part of the correlation. In
particular, during the third measurement on the 18th of June 2023, the SO₂ sensor frequently reaches its maximum value within
the plume, at 16 ppmv SO₂. Nevertheless, a clear anti-correlation of SO₂ and O₃ can be observed for all measurement flights,
305 characterized by a R² of 0.04 – 0.41.

5 Future Developments

Based on the first miniaturised VOLCANO₃ CL O₃ monitor prototype described here a number of improvements of our CL
O₃ monitor are clearly possible. Our field measurements suggest that the instrument would profit from further size and weight
reduction to enhance aerodynamic properties and flight stability of the carrying drone.

310 The largest components of the instrument are the PMT, the Teflon aerosol filter and the C₂H₄ minican. Using a smaller aerosol
filter would reduce weight and size significantly. The minican itself is voluminous but quite lightweight, although the minican
adapter is particularly heavy. A custom-made adapter might solve this problem.

Exploring other gases (or liquids with high vapour pressure) than C₂H₄ to induce chemiluminescence might lead to a better
photon yield of the luminescence and may also present an opportunity to further reduce the size of the instrument. For example,

315 trimethylethylene and tetramethylethylene, (C_6H_{12}) might be used. Both species are liquid at room temperature and exhibit a
by a factor of 50 higher quantum efficiency and thus emission intensity compared to that of the $C_2H_4-O_3$ reaction (Pitts Jr et
al., 1971). Storing the reactant in the liquid phase would significantly reduce the volume of the monitor without reducing the
amount of reactant available. With the vapour pressure of 185 mbar (ChemSpider, 2023), C_6H_{12} is available with a mixing
ratio of $185/1000 \approx 18\%$. If the C_6H_{12} flow is set to around 10% of the total flow, C_6H_{12} could also be supplied to the
320 measurement cell with a mixing ratio of $\approx 2\%$ (which is the mean mixing ratio of C_2H_4 in the current setup).

As mentioned above, the PMT is a rather bulky part of the instrument, in principle it could be replaced by one or several
avalanche photodiodes (also known as silicon photomultipliers), which would save space and weight.

Furthermore, the fraction of CL-photons reaching the detector could be enhanced by lining the interior surface of the present
aluminium fluorescence cell ($R_w \approx 0.7$) with material of higher reflectivity, for instance with “spectralon-type” material (e.g.
325 ODM98 from Gigahertz Optik GmbH), ($R_w > 0.96-0.99$), which would enhance the signal by at least a factor of 2, for details
see Bräutigam (2022).

Although the current control unit, based on a Raspberry Pi microcomputer is not very large it could still be replaced by a
smaller microcomputer unit like an ESP microcontroller for instance. To summarize, there is still room for improvement, but
the current instrument is already working excellently.

330 **6 Discussion and Conclusion**

Our newly developed lightweight VOLCANO₃ CL O₃ instrument marks a significant advancement in O₃ monitoring
technology. Weighing only 1.5 kg and with substantially smaller dimensions compared to commercially available CL-O₃
monitors (minimum weight of about 15 kg, which is an order of magnitude heavier and a volume more than five times larger
as the monitor presented here)

335 (see e.g. https://www.teledyne-api.com/en-us/Products/_Documents/Manual/T265%20Manual%20Addendum_07337.pdf).

Additionally, a notable reduction in power consumption was achieved, with VOLCANO₃ consuming now about 3 W, which
is also several times smaller than current commercially available instruments.

Calibration measurements in Heidelberg, as well as measurements in the field, showed the monitor’s capability to reliably
measure O₃. While signal corrections are imperative for deducing the O₃ mixing ratio, a detection limit of 4.16 ppbv and a
340 measurement accuracy of around 7% for 40 ppbv O₃ (~ 2.8 ppbv) were accomplished, with a response time of $\tau_{exp,1/e} = 2.00 \pm$
0.04 s.

With this miniaturised monitor we successfully detected significant O₃ depletions within the volcanic plume ranging up to
60%. While our finding is consistent with measurements by Surl et al., 2015 (15% - 45%) and Vance et al., 2010 (15% - 40
%), which also indicate a strong anti-correlation between SO₂ and O₃, the observed O₃ depletion within the volcanic plume
345 presents still an intriguing scientific puzzle. For instance, to fully answer the question on O₃ distributions in volcanic plumes
and if O₃ might be a limiting factor on the bromine transformation in volcanic plumes, more comprehensive measurement
campaigns are essential and care should be taken to complement the O₃ measurements by applying an SO₂ sensor which covers
the entire range of SO₂ mixing ratios in the plume under investigation. It would be worthwhile to study various volcanoes to
get a more complete picture. The examination of further high halogen emitting volcanos (additionally to Etna), such as
350 Soufriere Hills and Ambrym as well as rather low halogen emitting volcanos, including the Hawaiian volcanoes such as
Kilauea and Mauna Loa, or Copahue in Argentina/Chile would be worth to undertake. Complete plume transects should be
conducted in various distances from the volcanic emission source to confirm or disprove our current understanding of the
volcanic plume chemistry. Today, our knowledge is mainly based on model studies. Such model studies show for instance a
complete ozone depletion in the centre of halogen rich volcanic plume after a relatively short distance from the emission point
355 (about 10 min downwind, Roberts et al., 2014) but a solid experimental proof of these model predictions is still missing.

This CL-O₃ monitor applied on an UAV can also be used in other environments to undertake for instance high spatial resolved O₃ measurements in biomass burning plumes, over polar ice fields, salt lakes, in the rain forest or cities, to provide high resolution maps of the O₃ distribution during day and night (e.g., Guimeres et al., 2019). However, the challenges encountered and the intriguing findings regarding O₃ depletion in volcanic plumes highlight the potential.

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Code/Data availability

370 All relevant data for the conclusions drawn in this manuscript will be made available on Zenodo upon acceptance of the manuscript.

Author contribution

MR and EB are the main developers of the VOLCANO₃ instrument with support and contributions from UP, RK, JK and NB.
375 MR, NB, AN, TH, NK und BG conducted the field campaign at Mt. Etna and provided complementary data. MR evaluated and analyzed the ozone data. All authors contributed to writing of the manuscript.

Competing interests

One of the authors (UP) is member of the editorial board of AMT. The authors declare to have no other competing interests.

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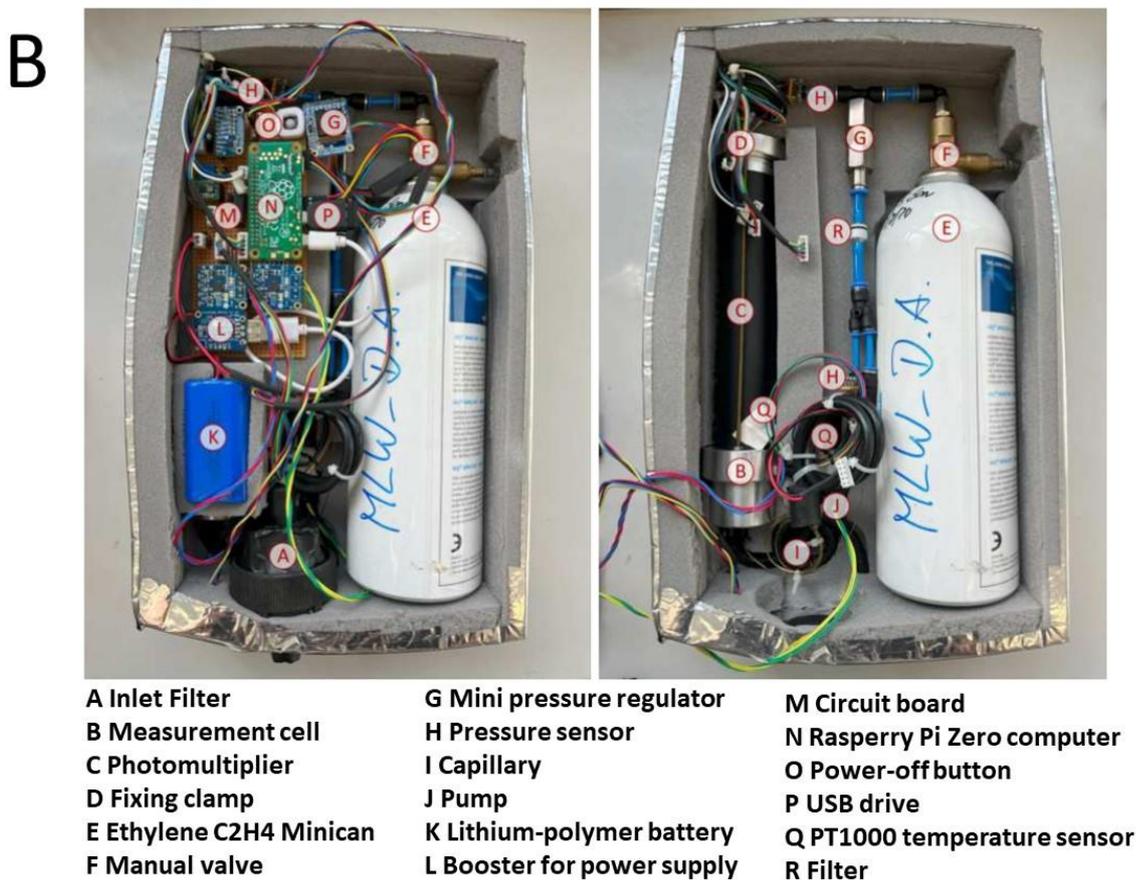
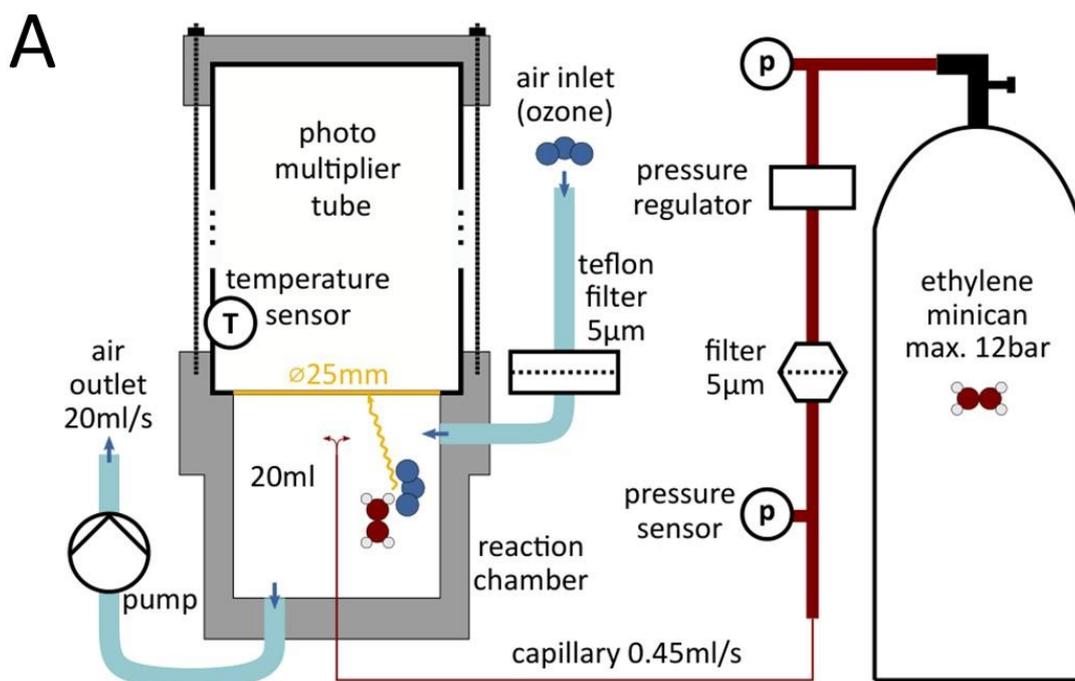


Figure 1 A: Schematic drawing of the CL O₃ monitor setup (without the electronics). Figure taken from Bräutigam (2022). **B:** Fully assembled monitor with the cover open in the photo on the left. In the photo on the right, the electronics have been removed to provide a better view of all parts of the monitor.

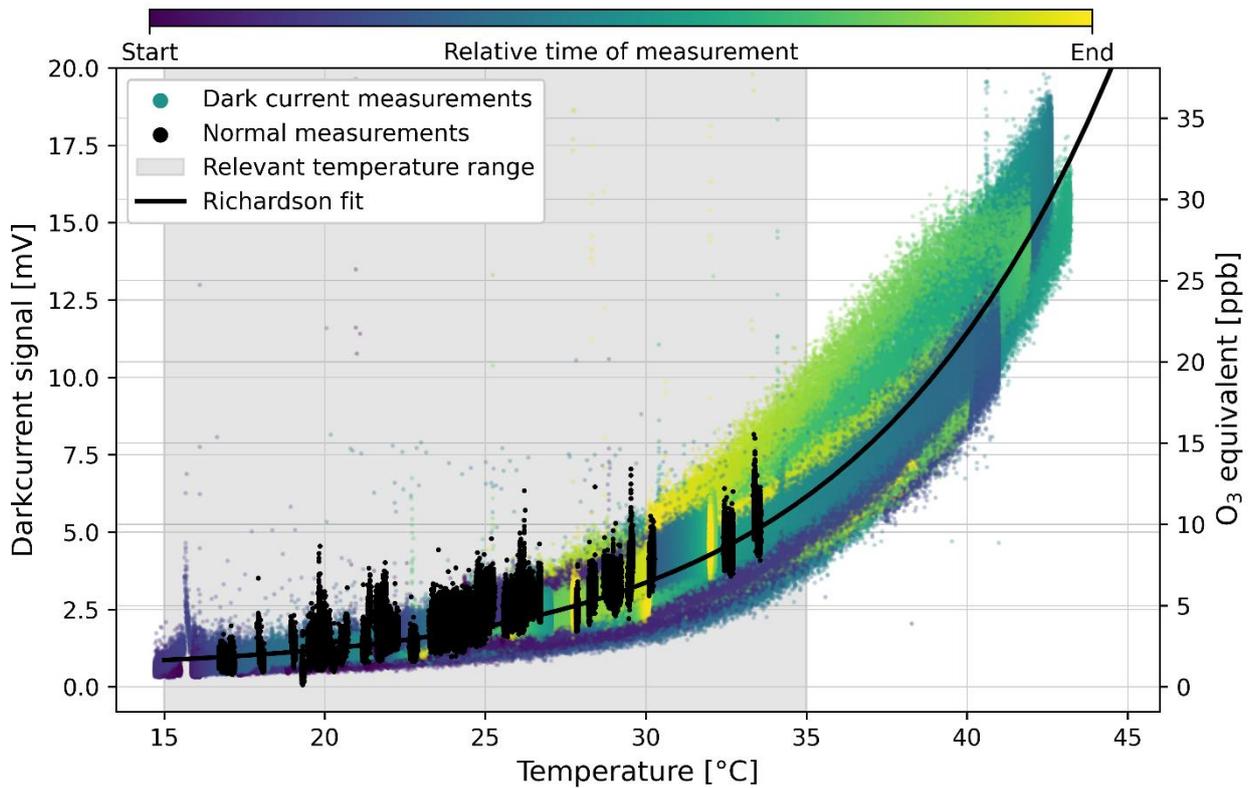


Figure 2: Temperature dependence of the dark current. The colour shade of the datapoints indicate time from blue (start of the measurement) towards yellow (end of the measurement). The black points represent ‘normal’ O₃ measurements during which also dark current periods were recorded. The data is fitted with a Richardson fit (see Eqn. 3, parameters: $a_1 = (9.9 \pm 0.3) \cdot 10^{12} \text{ mV/K}^2$, $b_1 = (-1.2 \pm 0.001) \cdot 10^4 \text{ K}$, $c_1 = (0.55 \pm 0.003) \text{ mV}$) which is then used to correct the temperature dependence of the dark current.

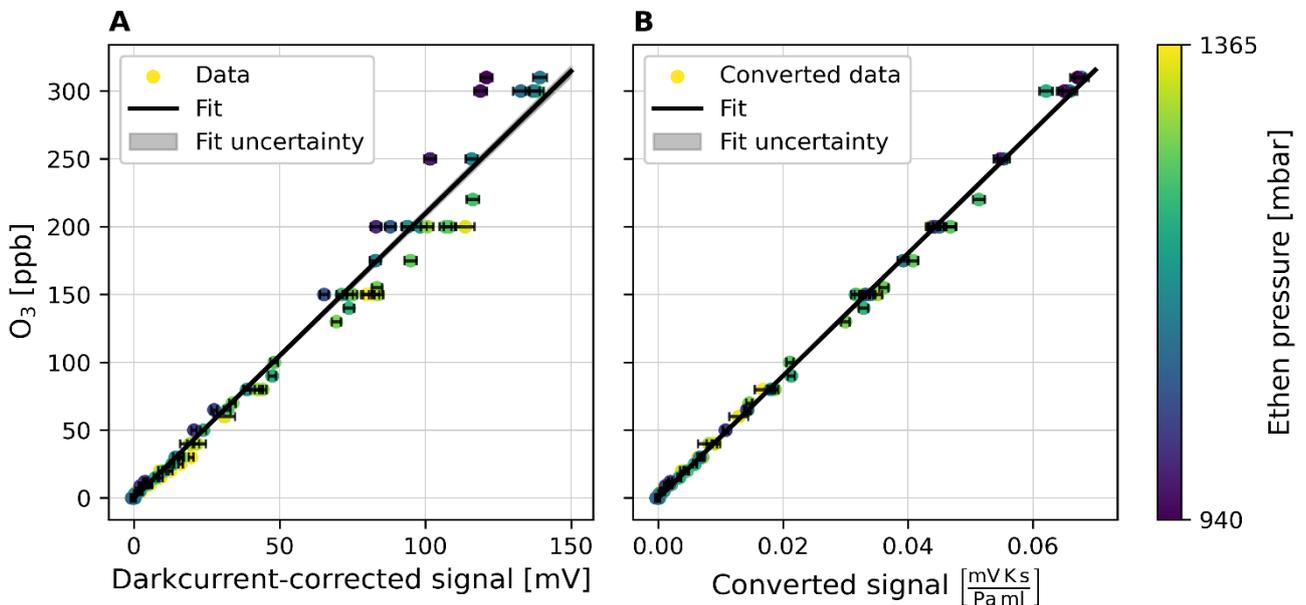
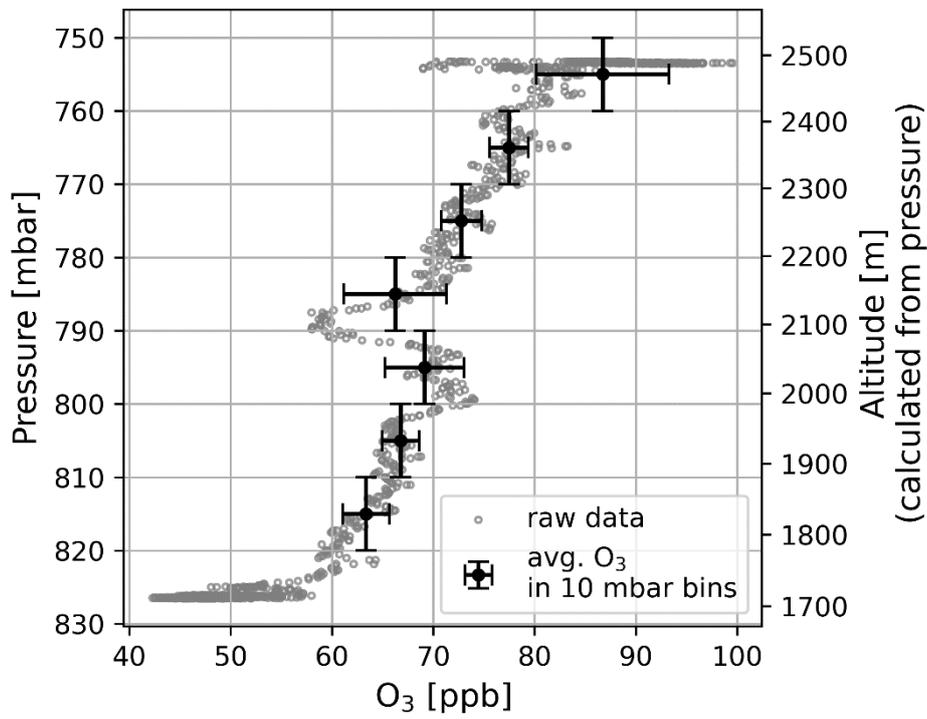
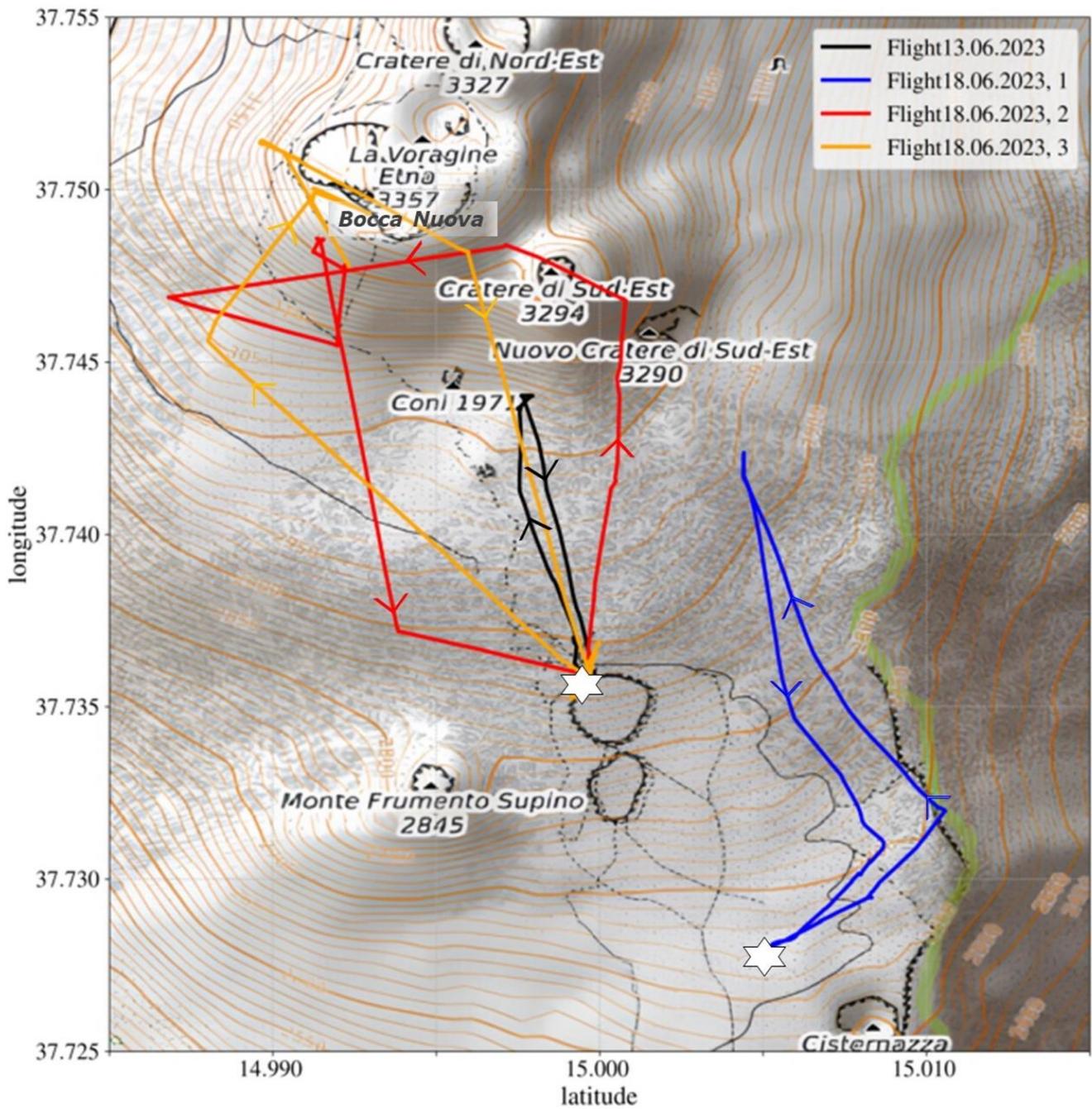


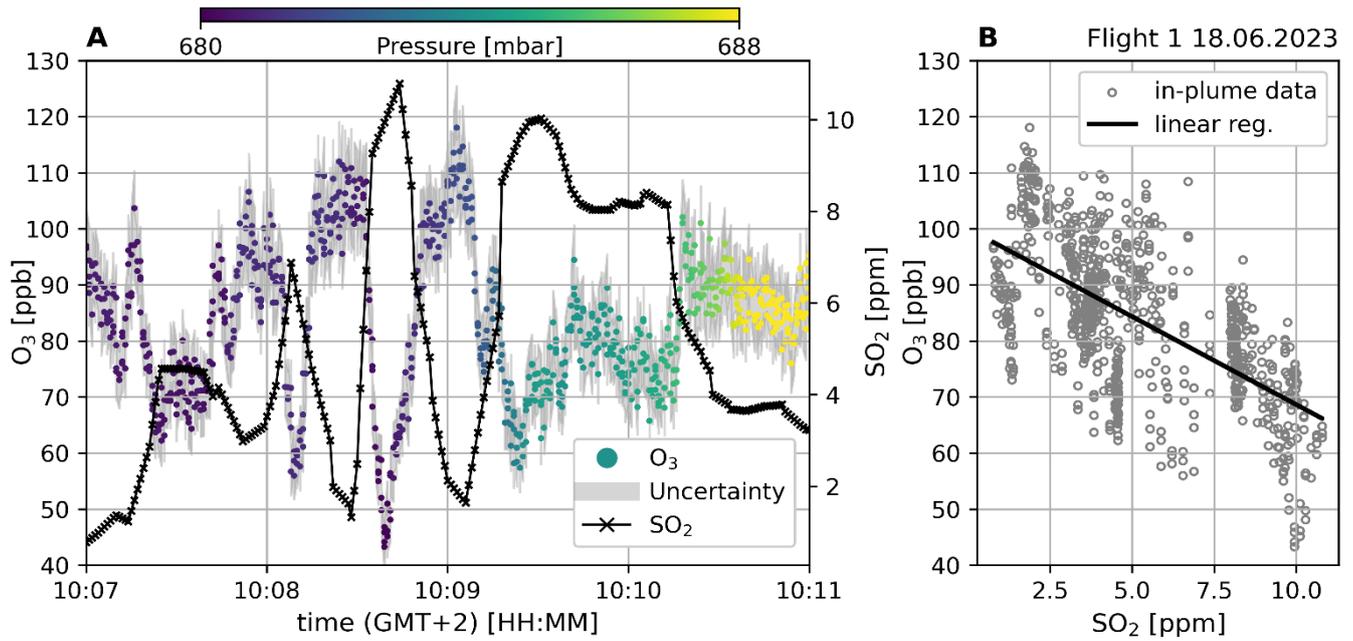
Figure 3: Calibration plot of the CL O₃ monitor. To obtain the data points, the signal is averaged over periods of constant O₃ mixing ratios and the uncertainty is given by the standard deviation. The color of the data points indicates the respective C₂H₄ pressure pC₂H₄. On the left side, the calibration plot with the dark current corrected signal is shown. Strong deviations from the calibration fit can be seen in the data points. After the conversion, on the right side, the calibration curve fits the data points significantly better. For both cases the calibration fit with the fit parameters as well as root mean squared error (RSME) is shown.



525 **Figure 4: Vertical O₃ profile for a measurement performed at Piano Vettore on 16.06.2023 during the Mt. Etna field campaign. The data points (and uncertainties) are determined by averaging the O₃ mixing ratio for all data points within a pressure bin of 10 mbar. The O₃ mixing ratio increases with height (increases with decreasing ambient pressure).**



530 **Figure 5:** Map of the summit area of Etna with the flight paths for the four plume measurements. The measurement on 13.06., as well as 18.06, 3, sampled the BN plume. The first flight on 18.06, 1, manoeuvred through the SEC plume and the second flight on 18.06, 2, first sampled the SEC plume before measuring the BN plume. The white stars indicate the launch and return points of the drone. Map data: © OpenStreetMap contributors, SRTM | Map display: © Open- TopoMap (CC-BY-SA).



540 **Figure 6: Plume measurement from 18.06.2023 measuring the SEC plume (see Figure 5 blue line). Panel A shows the time series from the CL instrument and a co-deployed (multi-gas) SO₂ sensor. The colour-coding indicates ambient pressure during sampling and the grey shaded area marks the uncertainty of the CL O₃ measurement. Panel B shows a correlation plot between O₃ and SO₂ for in-plume data points. In-plume datapoints are defined according to the SO₂ mixing ratio for values larger than 1.5 ppm and smaller than the saturation value of 16 ppmv. The correlation has a Pearson correlation coefficient of -0.64 and an R² of 0.41.**

Date and Flight	13.06.2023	18.06.2023, Flight 1	18.06.2023, Flight 2	18.06.2023, Flight 3
Average distance to crater [km]	1.5 ± 0.5	2.5 ± 0.5	1.1 ± 0.5	1.0 ± 0.5
Mean ambient O ₃ at ground [ppbv]	80 ± 4	66 ± 4	62 ± 4	64 ± 3
Mean ambient O ₃ at plume height [ppbv]	97 ± 6	101 ± 4	111 ± 4	109 ± 3
Mean O ₃ in the plume [ppbv]	93 ± 5	91 ± 13	97 ± 14	91 ± 13
Mean O ₃ depletion [%]	4	10	13	16
Minimum O ₃ in plume [ppbv]	82	52	60	45
Maximum O ₃ depletion [%]	15	49	46	59
Mean SO ₂ [ppmv]	7.4 ± 5.2	3.4 ± 2.5	5.2 ± 3.9	8.5 ± 5.4
Maximum SO ₂ [ppmv]	16*	9.7	16*	16*
Slope [ppbv/ppmv]	-0.87 ± 0.07	-3.6 ± 0.1	-1.9 ± 0.1	-0.51 ± 0.07
Correlation coefficient	-0.49	-0.64	-0.54	-0.20
R ²	0.24	0.41	0.29	0.04

545

Table 1: Summary of the plume measurement results. Mean ambient O₃ levels at the ground and at the height of the plume are determined by averaging over the respective periods determined by the ambient pressure (and excluding data points with SO₂ mixing ratios larger than 0.5 ppbv). The maximum SO₂ value marked with a * indicates that the sensor was in saturation and the real value is likely higher than reported. The mean O₃ inside the plume is obtained by averaging over periods for which SO₂ > 0.5 ppmv and below 16 ppmv. The mean O₃ depletion is determined by comparing the mean O₃ inside the plume with the mean ambient O₃ at height of the plume. The maximum O₃ depletion is calculated using the minimum O₃ value inside the plume and comparing it to the ambient O₃ at height of the plume.

550