Volcanic plume characteristics determined using an infrared imaging camera

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Abstract

Measurements of volcanic emissions (ash and SO₂) from small-sized eruptions at three geographically dispersed volcanoes are presented from a novel, multichannel, uncooled imaging infrared camera. Infrared instruments and cameras have been used previously at volcanoes to study lava bodies and to assess plume dynamics using high temperature sources. Here we use spectrally resolved narrowband ($\sim 0.5\text{--}1\text{ }\mu\text{m}$ bandwidth) imagery to retrieve SO₂ and ash slant column densities (g m^{-2}) and emission rates or fluxes from infrared thermal imagery at close to ambient atmospheric temperatures. The relatively fast sampling (0.1–0.5 Hz) of the multispectral imagery and the fast sampling (~ 1 Hz) of single channel temperature data permit analysis of some aspects of plume dynamics. Estimations of SO₂ and ash mass fluxes, and total slant column densities of SO₂ and fine ash in individual small ex-

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plosions from Stromboli (Italy) and Karymsky (Russia), and total SO₂ slant column densities and fluxes from Láscar (Chile) volcanoes, are provided. We evaluate the temporal evolution of fine ash particle sizes in ash-rich explosions at Stromboli and Karymsky and use these observations to infer the presence of at least two distinct fine ash modes, with mean radii $<10\ \mu\text{m}$ and $>10\ \mu\text{m}$. The camera and techniques detailed here provide a tool to quickly and remotely estimate fluxes of fine ash and SO₂ gas and characterize eruption size.

Keywords: Volcanic emissions, Volcanic plumes, Thermal imaging broadband

1. Introduction

Ground-based thermal infrared (IR) remote sensing has become a primary tool for volcano monitoring in recent years. High temporal resolution broadband thermal imaging cameras deployed in both ground-based and airborne configuration have been used at active volcanoes to monitor changes in surface temperature (Carter et al., 2007; Wessels et al., 2013), estimate lava effusion rates (Harris et al., 2007), evaluate changes in lava lake surfaces (Oppenheimer and Yirgu, 2002), image plume degassing (Ripepe et al., 2002; Harris and Ripepe, 2007b; Patrick, 2007), characterize explosive eruptions (Ripepe et al., 2005a,b; Patrick et al., 2007; Marchetti et al., 2009; Lopez et al., 2013), constrain emission velocities and plume dynamics (Harris and Ripepe, 2007a; Patrick et al., 2007; Delle Donne and Ripepe, 2012), and quantify eruption and emission masses (Delle Donne and Ripepe, 2012; Harris et al., 2013). Additionally, multichannel thermal IR instruments, such as the

15 airborne NASA Thermal Infrared Multispectral Scanner (TIMS) (Realmuto
 16 et al., 1994, 1997), and the nicAIR camera (Prata and Bernardo, 2009; Lopez
 17 et al., 2013), discussed here, exploit the characteristic IR absorption features
 18 of SO₂ gas, ice, and silicate ash, to allow SO₂ and very fine silicate ash to be
 19 detected and quantified. Here we describe methods for using nicAIR camera
 20 data to constrain ash and SO₂ slant column densities (SCDs), quantify plume
 21 ascent speeds, determine SO₂ and fine ash masses for individual explosions,
 22 and estimate mass fluxes of fine ash and SO₂. We apply these techniques
 23 at three active volcanoes exhibiting various styles of degassing and explosive
 24 eruptive activity: Stromboli (Italy), Karymsky (Russia), and Láscar (Chile).
 25 We include brief descriptions of the nicAIR camera and the retrieval methods
 26 for detection and quantification of ash and SO₂ and refer readers to Prata
 27 and Bernardo (2009) and Prata and Bernardo (2014) for further details.

28 The purpose here is not to provide new insights on volcanic processes, but
 29 rather to show the potential applications of the nicAIR camera to characterize
 30 volcanic plumes containing fine ash and/or SO₂ gas. The restriction to fine
 31 ash retrievals is emphasized as the results presented do not include ash mass
 32 estimates for particles with radii greater than approximately 50 μm . Here we
 33 aim specifically to illustrate what can be determined from thermal imagery,
 34 to present some basic methods for processing these data, and to report on
 35 the limitations and difficulties associated with this type of measurement.

36 **2. The nicAIR Infrared imaging camera**

37 The nicAIR is a multispectral IR imaging camera designed for portable
 38 field deployment at volcanoes and in industrial settings (Prata and Bernardo,

39 2009, 2014). The system is capable of detecting and quantifying SO₂ gas and
 40 particulates at sample rates of 0.1–0.5 Hz (or up to 1 Hz for single channel
 41 thermal data). The camera measures thermal radiation emitted and ab-
 42 sorbed by gases and particles in plumes and clouds, and by using specialized
 43 software, the image data can be transformed into SCDs of SO₂ or ash (in
 44 gm⁻², also referred to as ash mass loading in the literature). These SCDs
 45 can be used with feature tracking in the plumes to calculate SO₂ and ash
 46 fluxes (in kg s⁻¹). The sensitivity of the camera detector array is usually as-
 47 sessed by the Noise Equivalent Temperature Difference (NEDT), in Kelvin.
 48 For the broadband channel the NEDT≈50 mK, but for the filtered chan-
 49 nels this increases to 100–500 mK depending on the scene temperature and
 50 filter. Apart from NEDT limitations, the ability of the camera to provide
 51 quantitative estimates of SO₂ or ash SCDs depends on pixel size, distance
 52 to target, the transparency of the atmosphere between the camera and the
 53 plume, plume optical depth, the presence of interfering species (e.g. gases
 54 that absorb or emit in the waveband of 8–12 μm) and problems associated
 55 with non-synchronicity of the multispectral imagery due to movement of the
 56 filter wheel. The minimum errors have been assessed to be ~20% for SO₂
 57 by Prata and Bernardo (2014), but a full assessment of the error for ash re-
 58 trieval has not yet been made. It is likely that due to inaccurate knowledge
 59 of the refractive indices of the ash and of the size and shape distributions
 60 of the particles, in addition to the problems mentioned above, the retrieval
 61 error will be somewhat larger than that for SO₂. The error associated with
 62 ash retrievals from satellites is typically 40–60% (Wen and Rose, 1994), and
 63 this size of error can be expected from the ground-based thermal camera

64 retrievals described here for translucent plumes.

65 In Section 3, the methods used to process the data and obtain retrievals
66 of SO₂ SCDs and fluxes are provided and their accuracies assessed. Ash
67 retrievals are also presented for several cases and an estimate of the flux of
68 fine ash particles ($r < 50 \mu\text{m}$) is made in the case of Stromboli. The system
69 has a 24 hr capability and can be deployed in an automated manner to
70 provide volcano observatories with continuous information on SO₂ and ash
71 fluxes, and the maximum plume height of individual explosions. This last
72 parameter may be useful as an input parameter to initialize dispersion models
73 for forecasting the movement of volcanic ash clouds.

74 *2.1. Technical specifications and deployment scenarios*

75 Table 1 provides detailed technical specifications of the nicAIR camera
76 system. An example of a typical camera deployment in which the camera
77 is viewing SO₂ degassing from Láscar volcano (see Section 4.3) is shown in
78 Figure 1. In this example the camera is located ~ 6.5 km from the degassing
79 vent and tilted slightly upwards to view the plume and some clear sky. In
80 addition to viewing and quantifying daytime emissions, the nicAIR has the
81 significant advantage over ultraviolet (UV) instruments (Kern et al., 2014)
82 of also being able to measure emissions at night. During the Stromboli
83 Workshop (the main topic of this Special Issue), the nicAIR camera was
84 set up to measure continuously and to view both ash and gas emissions
85 during both day and nighttime conditions. Figure 2 shows a sequence of
86 20 broadband IR images, each separated by ~ 20 s that capture discrete
87 explosions from Stromboli Volcano. These explosions could be easily detected
88 and tracked for periods of several minutes using the camera during the day

89 and/or night. Table 2 lists the locations and important deployment logistics
 90 of the nicAIR cameras at the three volcanoes studied.

91 **3. Quantifying SO₂ and ash slant column densities and fluxes**

92 *3.1. SO₂ retrievals*

93 SO₂ gas has a strong IR absorption feature situated near 8.6 μm . This
 94 absorption band has been used by Realmuto et al. (1997) among others
 95 to determine the SCD of SO₂ using multispectral data from airborne and
 96 spaceborne instruments. Prata and Bernardo (2014) have described the
 97 retrieval scheme for determining SO₂ from the nicAIR imaging camera and
 98 we follow that procedure here. The retrieval procedure uses three channels
 99 at central wavelengths of 8.6, 10 and 12 μm (or the broadband channel). The
 100 information regarding SO₂ in the plume is contained in the 8.6 μm channel,
 101 while the 12 μm channel is used to correct for atmospheric effects, and the
 102 10 μm channel, which is the most transparent to water vapor absorption,
 103 is used to estimate the plume temperature. The retrieval scheme estimates
 104 the emissivity of the gas plume and determines the SO₂ SCD (g m^{-2}) from a
 105 laboratory measured absorption coefficient integrated over the filter response
 106 of the channel. The SO₂ SCD m_s , is calculated using the following equation
 107 from Prata and Bernardo (2014):

$$m_s = \rho d = -\frac{1}{k} \cos \theta \cos \phi \ln[1 - \epsilon], \quad (1)$$

108 where ρ is the concentration, d is path length through the plume, k is the
 109 absorption coefficient averaged over the response function of the measurement

channel, θ and ϕ are the camera viewing angles, and ϵ is an effective emissivity of the plume calculated using the method of Prata and Bernardo (2014). Note that m_s and ϵ have the same dimensions as an image frame (e.g. 640 x 512 pixels x lines) but only a subset of these contain SO_2 after processing. Data are screened by differencing images to determine times when plumes are relatively ash-free (see Section 3.3), and only ash-free images are evaluated here for SO_2 SCD. Note that at the time of writing no method exists to correct for the presence of ash while retrieving SO_2 ; a scheme to simultaneously determine both SO_2 and ash is currently under development.

3.2. Ash retrievals

A general retrieval scheme for ash SCD and particle size from multispectral IR camera measurements has been described by Prata and Bernardo (2009). The scheme is applicable in cases where the plumes are semi-transparent, usually at times 10–30 minutes after an ash explosion or at the edges of a rising plume, and is limited to fine-ash with radii $<16 \mu\text{m}$. At the start of an eruption the ash plume is quite often opaque such that retrieving information on the particle size and ash SCD is not possible. For the cases of semi-transparent plumes we have developed a new method that uses three IR channels and an *a priori* model size distribution to estimate total ash mass for particles of radii $<50 \mu\text{m}$, which extends the range of particle sizes detected using the method of Prata and Bernardo (2009) by a factor of three, but is still restricted to fine ash (as defined by Rose and Durant (2009)). It is well known that retrieving microphysical parameters from optical remote sensing data is an ill-posed problem (Liu et al., 1999). The general problem requires inversion of a Fredholm integral of the first kind and can be stated

135 as:

$$\tau(\lambda) = \int_{r_{min}}^{r_{max}} K(r, \lambda; \mathcal{R}) \frac{dn(r)}{d \ln r} dr, \quad (2)$$

136 where $K(r, \lambda; \mathcal{R})$ is a kernel which depends on the particle radius r , the
 137 wavelength λ , and implicitly on the complex index of refraction ($\mathcal{R}$) of the
 138 particles, through an extinction coefficient. The size distribution to be de-
 139 termined, $dn(r)/d \ln r$, appears inside the integral with limits r_{min} and r_{max}
 140 which are set by the spectral range of the measurements and the size range
 141 of particles under study.

142 The discrete version of (2) may be written:

$$\Delta \mathbf{H} = \mathbf{K}^{-1} \mathbf{\Upsilon}, \quad (3)$$

143 where $\Delta \mathbf{H}$ is the frequency distribution which can be related to the particle
 144 size distribution (PSD) from $\Delta \mathbf{H} = \Delta \mathbf{r} \mathbf{H}$, where $\Delta \mathbf{r}$ is the particle size
 145 increment and $\mathbf{H}$ is the PSD. $\mathbf{K}$ is the kernel matrix and $\mathbf{\Upsilon}$ is a set of
 146 measurements of optical depth. The elements of $\mathbf{K}$ may be calculated using
 147 Mie theory using an assumed size distribution of spherical particles and an
 148 assumed complex index of refraction. Recognizing the difficulty of solving
 149 this inverse problem with just three or four spectral measurements, a scheme
 150 has been developed to simplify this problem. This scheme uses the spatial
 151 distribution of the measured optical depths within an ash plume to infer the
 152 PSD, based on an *a priori* model with two free parameters. This assumes
 153 that the extinction does not depend on particle size, but rather the number of
 154 particles in the atmospheric column, and tends to be a better approximation
 155 for larger particles ($r > 16 \mu\text{m}$). Further details of the scheme are provided

156 in the Appendix.

157 3.2.1. *A priori size distribution model*

158 Volcanic ash PSDs have been frequently modeled using a log normal dis-
159 tribution (Wohletz, 1983; Bonadonna and Phillips, 2003; Rose and Durant,
160 2009) for very fine ($r < 30 \mu\text{m}$), fine ($30 < r < 63 \mu\text{m}$) and coarse ($r > 63 \mu\text{m}$)
161 particle sizes. Retrievals for larger particles $r > \sim 50 \mu\text{m}$ and very small par-
162 ticles $r < 1 \mu\text{m}$ are not possible from the spectral measurements used here.
163 The *a priori* log normal distribution model employed here is represented as,

$$\frac{dn(r)}{d \ln r} = \frac{1}{\sigma_l \sqrt{2\pi}} \left[\exp - \left\{ \frac{(\ln r - \ln \mu)^2}{2\sigma_l^2} \right\} \right], \quad (4)$$

164 where μ and $\sigma_l = \ln(\sigma)$, are the first two moments of the distribution and are
165 treated as free parameters to be estimated using a least squares minimization
166 routine. It should be clear that the retrieval depends entirely on the choice of
167 *a priori* size distribution and other choices for the underlying size distribution
168 are possible.

169 3.2.2. *Optical depths*

170 The IR optical depth can be used along with the *a priori* model to es-
171 timate the PSD for $r < 50 \mu\text{m}$. The optical depth (τ_λ) can be estimated
172 using the simple absorption model of Prata and Grant (2001). Assuming
173 no scattering and considering monochromatic radiation with plane-parallel
174 geometry, the IR emissivity at wavelength λ may be written,

$$\epsilon_\lambda = \frac{R_p - R_\lambda}{R_p - R_b}, \quad (5)$$

$$R_p > R_b,$$

175 where R_b is a background atmospheric radiance, R_p is the plume radiance
 176 determined from an estimate of the plume temperature (T_p), and R_λ is the
 177 measured radiance at wavelength λ . The background radiance is determined
 178 from the broadband channel, but the plume radiance is more difficult to
 179 determine (see next subsection).

180 The optical depth is related to the plume emissivity by

$$\tau_\lambda = -\ln[1 - \epsilon_\lambda]. \quad (6)$$

181 Channels at 10 and 11 μm are used to identify pixels in the images that have
 182 an ash signature. The condition for identifying a pixel containing ‘ash’ is,

$$T_{\lambda_1} - T_{\lambda_2} < \Delta T_0, \quad (7)$$

183 where T_{λ_1} and T_{λ_2} are measured brightness temperatures (in this case at
 184 $\lambda_1=10 \mu\text{m}$ and $\lambda_2=11 \mu\text{m}$) and ΔT_0 is a threshold temperature difference,
 185 taken as $\Delta T=0 \text{ K}$. The basis for this ash detection technique is described
 186 in Prata and Bernardo (2009). A third channel (the broadband channel is
 187 suitable) is used to identify the plume and background temperatures needed
 188 to estimate the optical depth (Fig. 3), which is determined at 10 μm .

189 3.2.3. Plume temperature

190 There are several different ways to estimate plume radiance. One is to
 191 look for a very opaque part of the plume and use the broadband measurement
 192 to determine the plume temperature by assuming the plume is a blackbody.

193 This temperature can then be re-inserted into the Planck function and the
 194 plume radiance at the wavelength, λ can be determined. A second way is to
 195 assume that the plume is in thermal equilibrium with its surroundings, and
 196 use an estimate of the air temperature at the level of the plume as the plume
 197 temperature, and again invert the Planck function with this temperature.
 198 A third way is to look for parts of the plume where the radiance at two
 199 wavelengths agree within a specified amount (selected as $\sqrt{2}$ times the sum
 200 of the noise equivalent radiances for the channels); the assumption being that
 201 at these places the plume is behaving as a grey body with a high emissivity.
 202 These parts of the plume must also be opaque, such that the temperature
 203 contrast between the background sky and the plume is as high as possible.
 204 Having calculated the radiance by one of these methods, the radiance must
 205 then be corrected for atmospheric absorption (and possibly emission) for the
 206 path between the plume and the camera. Prata and Bernardo (2014) show
 207 that the size of this correction is typically ± 3 K, and is dependent on the
 208 slant range distance between plume and camera. The optical depth error $\delta\tau_\lambda$
 209 associated with an error in the estimated plume radiance δR_p is:

$$\delta\tau_\lambda = \frac{1}{(R_p - R_b)} \delta R_p. \quad (8)$$

210 The optical depth error is small when the contrast between the plume and
 211 background radiances is large (e.g. thick plume against a cold clear sky). For
 212 measurements at $10\ \mu\text{m}$, and considering a plume temperature of 280 K and
 213 a background sky temperature of 270 K, the resulting optical depth error $\delta\tau_{10}$
 214 would be $\sim \pm 0.2$ for a plume temperature error of ± 2 K. This suggests that
 215 an error in optical depth estimation of at least 10% for optically thick clouds

216 ($\tau \sim 2$) will be due to plume temperature error alone. Figure 3 illustrates the
 217 plume temperature variation for a typical Strombolian ‘ashy’ eruption, where
 218 the ash cloud has detached from its source and is being advected by the local
 219 winds. A vertical transect through one part of the plume (solid line) is com-
 220 pared with a transect outside the plume (vertical dashed line). In this case
 221 the plume temperature is estimated as the maximum temperature along the
 222 transect (~ 290 K) and the background temperature is estimated at the same
 223 height outside the plume (~ 272 K), giving a temperature contrast of ~ 18
 224 K. The shape of the temperature profile inside the plume is due to opacity
 225 changes with height together with an effect due to the atmospheric temper-
 226 ature lapse rate (decreasing temperature with altitude in the troposphere),
 227 which is clearly seen in the profile outside the plume. Horizontal variation of
 228 the plume temperature is assumed to be due only to opacity variations. In
 229 the image shown in Figure 3, the horizontal gradient varies by less than 1 K
 230 for the region of the sky above the plume and by less than 2 K for regions
 231 that include the plume.

232 3.2.4. *Size distribution retrieval*

233 The distribution of optical depths is used to estimate parameters of the
 234 PSD by assuming that the optical depths and PSDs are (1) equivalent and
 235 (2) scaled by a constant factor, the mean effective particle size, set by the
 236 assumed constant mass absorption coefficient (see Section 3.3.6). The mean
 237 effective particle size used here is $10\ \mu\text{m}$. The following steps are taken to
 238 condition the data to ensure stability of the retrieval. First, the images are
 239 shifted by a small amount (typically 1–5 pixels) in pixel/line space to improve
 240 spatial coherence between images at different wavelengths acquired during

241 the same duty cycle but at slightly different times. Because of the finite
 242 time required to move the filter wheel, images are never simultaneous and
 243 plumes typically move a small distance between successive images. Secondly,
 244 images are smoothed with a median filter with a filter length of 3 pixels
 245 and optical depths where $\tau < 0.01$ and $\tau > 5$ are not considered. This last
 246 condition places constraints on the smallest and largest particle sizes that
 247 can reliably be retrieved of $\sim 1 \mu\text{m}$ and $\sim 50 \mu\text{m}^1$, because we have assumed
 248 that the opacity is directly related to the particle size, and there are limits
 249 on the opacity determination. Once a retrieval of the optical depth has
 250 been performed, an elliptical region around the plume is identified and the
 251 histogram of optical depths within this ellipse is computed.

252 3.2.5. Ash slant column densities

253 The following equation shows the relationship between ash SCD (g m^{-2})
 254 and optical depth, τ :

$$m_l(i) = \frac{\tau(i)}{k_{abs}}, \quad (9)$$

$$k_{abs} = \frac{3}{4} \frac{Q_{ext}^\infty}{\rho} \quad (10)$$

256 where $m_l(i)$ is the ash SCD for pixel number i , k_{abs} is a mass absorption
 257 coefficient ($\text{m}^2 \text{g}^{-1}$), ρ is the density of the particle (g m^{-3}), and Q_{ext}^∞ is the
 258 extinction efficiency (m^{-1}), which for large particle sizes may be treated as a
 259 constant, independent of the size distribution. Mie calculations for $r_e > 16$
 260 μm , where r_e is the effective radius, suggest $Q_{ext}^\infty \sim 0.048 \times 10^6 \text{ m}^{-1}$ at $10 \mu\text{m}$

¹These limits are not strict because the opacity retrieval depends also on the clarity of the intervening atmosphere for low opacity clouds and when the opacity is high, a small change in retrieved opacity gives a large change in the retrieved particle size.

261 wavelength for andesitic, spherical ash particles (see Prata (1989), Figure
262 1).

263 Previous work (Prata, 1989; Wen and Rose, 1994; Francis et al., 2012)
264 has shown that the ash composition has an important effect on the mass
265 absorption, but the ash particle shape is less important. The effect of errors
266 in the mass absorption coefficient due to errors in composition will be to
267 scale the retrieved SCDs (and total mass) by some amount. The effect of
268 assuming k_{abs} to be independent of particle size will be more complex. It is
269 certainly not a justifiable assumption for smaller ($r < \sim 16 \mu\text{m}$) particles, but
270 we note that most of the mass resides in the larger particle size range.

271 3.3. Plume rise–optical flow

272 Measurements of plume rise do not require multispectral imagery such
273 that the nicAIR camera can be operated in single-channel mode to increase
274 the frequency of data acquisition up to 1 Hz². Previous studies of plumes
275 using imaging cameras have used feature tracking (Bluth et al., 2007) and
276 edge detection (Lopez et al., 2013). The dynamic and often turbulent nature
277 of volcanic plumes means that rise rates are highly variable over short time
278 scales and spatially within the plume, such that the aforementioned meth-
279 ods will therefore miss some inhomogeneity in motion. Here we present the
280 implementation of an optical flow model over the range of eruptive settings
281 to demonstrate the additional information that can be gathered on plume
282 dynamics from infrared camera imagery.

²It is possible to get single-channel full-frame uncalibrated imagery at ~ 7 Hz; however for this application calibrated data are needed.

Optical flow is defined as the distribution of velocities from the motion of objects, edges or shapes due to relative motion between the observer and the targets (Gibson, 1950). Optical flow methodologies attempt to compute the motion between two image frames at times t and $t+\Delta t$ and are based on the assumption that the brightness (I) of the pixel at (x,y) (for the 2-dimensional plus time case) at time t is equal to the brightness of the pixel at $(x+\Delta x, y+\Delta y)$ and time $t+\Delta t$. Thus:

$$I(x, y, t) = I(x + \Delta x, y + \Delta y, t + \Delta t) \quad (11)$$

A number of methodologies have been developed for solving the above equation and thus calculating optical flow. Six methods, developed prior to 1994 have been compared by Barron et al. (1994), although many more have been developed since (Liu et al., 1997; Jähne et al., 1998; Farneback, 2001).

The examples given here demonstrate an implementation of a dense optical flow algorithm following the method of Farneback (2003) where the optical flow is calculated for every pixel in the scene. The algorithm was implemented using the OpenCV library for Python (Bradski and Kaehler, 2008). The Farneback (2003) estimation utilizes the brightness assumption of Eq. (10) and estimates the motion of each pixel assuming that neighboring pixels undergo similar levels of displacement through the quadratic polynomial expression:

$$F(I) = I\mathbf{A}^T I + b^T I + C, \quad (12)$$

where $\mathbf{A}$ is a symmetric matrix, b is a vector and C is a scalar. The

303 coefficients b and C are estimated from a weighted least squares fit to the
304 signal values for all the pixels in the neighborhood. More weight is given to
305 pixels close to the center of the neighborhood, and pixels outside of the image
306 are ignored. The size of the pixel neighborhood to be considered is chosen by
307 the user and can be made larger to improve robustness but will lose some of
308 the intricacies of the motion (Farneback, 2003). Pyramid layers are used to
309 increase the robustness of the algorithm when dealing with both small and
310 large displacements. Each pyramid layer reduces the pixel resolution of the
311 image by a specified factor, so that the number of pixels becomes fewer with
312 each incremental layer.

313 A popular alternative method of solving the optical flow is that of Lucas
314 et al. (1981) (the so-called *Lucas-Kanade method*). This method is a sparse
315 technique, only calculating flow for significant features within the image.
316 It was found in this application, however, that due to the local nature of
317 the expansion method, it was less effective than the Farneback method in
318 discriminating the flow patterns in the center of the clouds. Results for a
319 discrete Strombolian eruption using the optical flow scheme are shown in
320 Figure 4. An animation illustrating the complex plume dynamics during
321 an explosive eruption of Karymsky Volcano using the optical flow model is
322 shown in Supplementary Material 1.

323 3.4. Mass fluxes (emission/transport rates) of SO_2 and ash

324 Since it is possible to retrieve the SCD of SO_2 and ash, and estimate the
325 two dimensional velocity field using optical flow, it is in principle possible to
326 estimate the mass flux of SO_2 and ash. The mass flux (in $kg\ s^{-1}$), is often
327 referred to as the emission rate or transport rate in the literature, and may

328 be defined as:

$$\vec{J} = \rho \int \int_S \vec{n} \cdot \vec{F} dS, \quad (13)$$

329 where $\vec{J}$ is the flux vector (or emission/transport rate), ρ is the concentration
 330 of SO₂ or ash, dS is an elemental area of the plume surface S , $\vec{n}$ is a unit
 331 vector normal to dS and pointing outwards from the surface, $\vec{F}$ is the velocity
 332 field vector and the integral is performed over a closed surface. The camera
 333 measurements only provide the SCD and the optical flow algorithm only
 334 provides two components of the velocity vector. Thus we can only estimate
 335 the component of the flux in directions defined in a plane with axes oriented
 336 orthogonal to the camera viewing direction. For simplicity if these axes are
 337 oriented so that the x -axis is parallel to the horizontal view angle and the
 338 z -axis is parallel to the vertical viewing direction then:

$$\vec{F} = \langle \vec{u}, 0, \vec{w} \rangle, \quad (14)$$

339 where $\vec{u}$ and $\vec{w}$ are the velocity components in the x and z directions, re-
 340 spectively. In the following examples, the flux calculation is performed by
 341 dividing the plume surface into small elemental areas ($\delta A = \delta x \delta z$), multiply-
 342 ing each element by both $|\vec{F}|$ and the SCD, and finally dividing by δx (or δz ,
 343 as these are chosen to be equal). The magnitude (scalar) flux of material, de-
 344 noted by $|\vec{J}|$ (to distinguish it from the true vector flux, $\vec{J}$) from the volcano
 345 is then calculated as the average of all of these elemental fluxes. Note that we
 346 cannot estimate the true vector flux; only the components orthogonal to the
 347 camera view direction. With two cameras placed orthogonal to each other
 348 and in the same plane it is feasible to estimate $\vec{J}$.

349 4. Applications of nicAIR for volcanic plume characterization

350 To illustrate the capabilities of the camera at determining volcanic plume
351 characteristics, measurements from three different volcanoes are presented
352 including Stromboli (Italy), Karymsky (Russia), and Láscar (Chile) (Fig-
353 ure 5). The nicAIR camera data are highly complementary to other geophys-
354 ical datasets (e.g. infrasound data as in Lopez et al. (2013)) and theoretical
355 models (e.g. for volcanic plume dynamics), and can be used in conjunction
356 with these tools to obtain new insight into volcanic processes. These case
357 studies were selected to illustrate different retrieval products. In the exam-
358 ples from Stromboli and Karymsky we explore the use of optical flow and
359 ash mass retrievals to determine fluxes of fine ash. In the Láscar example we
360 retrieve SO₂ SCDs and calculate fluxes. Each case presents different chal-
361 lenges: the Karymsky eruptions are much more explosive (and opaque) than
362 those at Stromboli, where ash clouds rarely exceeded a few hundred me-
363 ters from the vent and commonly drifted in the light winds. SO₂ emissions
364 at Láscar were much less energetic (low plume rise height) than those at
365 Stromboli making the retrievals more difficult to perform. Maps showing the
366 target volcanoes and the nicAIR location during each measurement period
367 are shown in Figure 5.

368 4.1. *Stromboli*

369 4.1.1. *Volcano setting*

370 Stromboli Volcano (38.789°N, 15.213°E, 924 m) is a potassium-rich basaltic
371 to basaltic andesite volcano located in the Aeolian Archipelago of Italy (Rosi
372 et al., 2000). Stromboli is famous for its persistent, low-level explosive activ-

373 ity that has been occurring since ~ 700 A.D. (Chouet et al., 1974; Barberi
 374 et al., 1993; Rosi et al., 2000). Typical activity at Stromboli consists of
 375 persistent degassing accompanied by discrete explosions occurring approxi-
 376 mately every 5-20 minutes, which emit ash to block-sized scoracious material
 377 to heights ranging from tens to hundreds of meters above the vent (Barberi
 378 et al., 1993; Delle Donne et al., 2006; Lautze and Houghton, 2005; Patrick
 379 et al., 2007; Ripepe et al., 2008). This regular activity is punctuated by
 380 more energetic major explosions (Bertagnini et al., 1999), effusion of lava
 381 flows (Calvari et al., 2011), and less frequently, paroxysmal eruptions, which
 382 are higher energy explosions, producing ash clouds to >10 km altitudes and
 383 erupting large volumes of ejecta (Barberi et al., 1993). Multiparameter ob-
 384 servational datasets and model interpretations suggest that explosions at
 385 Stromboli are due to gas accumulation and ascent within the conduit, cul-
 386 minating in bubble burst at the magma free surface (Blackburn et al., 1976;
 387 Jaupart and Vergnolle, 1988; Harris and Ripepe, 2007a). Emissions of ash
 388 and gas from Stromboli are sourced from three primary vents within the sum-
 389 mit crater terrace (Bertagnini et al., 1999). Previous estimates of total gas
 390 flux during periods of quiescent degassing have been determined using a com-
 391 bination of ground based remote sensing instruments including COSPEC and
 392 open-path FTIR, and direct sampling, and find a value of 5400 ± 860 t day $^{-1}$
 393 (62 ± 10 kg s $^{-1}$), which includes 250 ± 50 t day $^{-1}$ (3.0 ± 0.5 kg s $^{-1}$) SO $_2$ (Allard
 394 et al., 2008). Note that gas fluxes during eruptive periods will be greater,
 395 for example Burton et al. (2009) found an average daily SO $_2$ flux of 620 t
 396 day $^{-1}$ (7.2 kg s $^{-1}$) during the 2007 eruption using the continuous FLAME
 397 network of UV spectrometers; however previous work suggests that quies-

cent degassing represents 95% of total SO₂ degassing at Stromboli (Allard et al., 1994). The combined mass of lava, ash and gas emitted from individual Strombolian explosions has been estimated using visible or thermal image data in a number of studies, including Chouet et al. (1974), Ripepe et al. (1993), Harris and Ripepe (2007a), Delle Donne and Ripepe (2012) and Harris et al. (2013). The most recent of these studies estimate the mass of bombs and gas/ash emitted during individual explosions using thermal remote sensing imagery of the bomb field and ascending plume (Harris et al., 2013) and find minimum bomb and gas/ash masses ranging from 900–31400 kg (mean value of 9113 ± 11466 kg), and 109–1054 kg (mean value of 623 ± 340 kg), respectively. These results are in good agreement with Ripepe et al. (1993), but significantly higher than the other studies (mean value ~ 7 –200 times larger than the minimum values found) (Harris et al., 2013). These estimates use volumes of emitted material (calculated from plume or bomb area within the image) to calculate bomb and gas/ash masses, which depend on significant assumptions regarding the plume composition and density, unlike data from the multichannel camera employed here, which can distinguish fine ash and gas. Additionally, a study by Mori and Burton (2009) find a total mass of SO₂ released during individual Strombolian explosions to range from 15–40 kg using a UV camera.

4.1.2. *Experimental setup*

For several days in June 2013 the nicAIR camera was used to measure SO₂ and ash within plumes associated with explosions from Stromboli. The nicAIR camera was set up approximately 2.6 km from the two main erupting vents at an inclination angle of ~ 14 degrees above the horizontal. With this

423 geometry the average pixel size is $\sim 7 \text{ m}^2$ (Table 2).

424 4.1.3. *Ash results*

425 Ash retrievals were performed over several periods for which isolated
426 small, ash-rich explosions occurred, described by Patrick et al. (2007) as
427 Type 2 Strombolian eruptions. These “puffs” seldom rose to more than a
428 few hundred meters above the vent and dissipated within 5–10 minutes. The
429 plumes were clearly buoyant and were dispersed by light winds. Results of
430 fine ash mass and PSD are presented for 23 June when winds were light (< 5
431 m s^{-1}) and the weather conditions were favorable with only a few meteorological
432 clouds. Figure 6 shows a period from 18:05 UT through to 18:09 UT,
433 when a small ash puff could be tracked as it moved southwestwards from the
434 vent towards the coast. The total mass of fine ash in these ash-rich puffs is
435 estimated to be up to $\sim 2000 \text{ kg}$ ($\sim 2 \text{ t}$), with a bimodal size distribution and
436 median peaks in effective particle radii near to $\sim 20\text{--}30 \text{ }\mu\text{m}$ and $\sim 0.5\text{--}10 \text{ }\mu\text{m}$
437 (Figure 6 and 7). Within five minutes of emission the mass dropped to ~ 500
438 kg ($\sim 0.5 \text{ t}$) and most of the remaining ash particles had radii $< 10 \text{ }\mu\text{m}$, with a
439 median peak of $\sim 0.5 \text{ }\mu\text{m}$ (Figure 6). At the end of the sequence of retrievals,
440 when the puff had all but dissipated, the ratio of pixels with radii less than 10
441 μm to those with radii greater than $10 \text{ }\mu\text{m}$ is $\sim 6\%$. One interesting finding
442 of the IR retrievals is that there appears to be an airborne fraction remaining
443 in the atmosphere with median particle radius $\sim 0.5 \text{ }\mu\text{m}$, or perhaps smaller
444 (Figure 6). The majority of the larger particles (aerodynamic radii $> 100 \text{ }\mu\text{m}$)
445 have fallen to the ground within 30 minutes of the explosion onset, consistent
446 with the timescale for gravitational settling speeds of $\sim 0.3 \text{ m s}^{-1}$. The evolution
447 of the median radius for the two modes is shown in Figure 7, together

448 with the total mass of fine ash. The temporal behavior of the PSDs shows no
449 evidence of particle aggregation in the cases studied, since at all later times
450 the size distribution shifts to smaller particles, eventually reaching sizes too
451 small to be resolved by the IR camera.

452 The mass flux of fine ash within discrete explosions can be estimated using
453 a combination of the ash mass retrievals and the plume velocity determined
454 from the optical flow methodology as described in Section 3.3. Figure 8
455 illustrates an example from 18:24 UT to 18:28 UT on 23 June 2013, where the
456 nicAIR camera observed an ash cloud produced by a discrete explosion that
457 emerged from the vent into the atmosphere. Reliable fine ash mass estimates
458 and plume velocities are obtained within ~ 30 s of the first observation, and
459 up to ~ 5 minutes afterwards. Before ~ 30 s and beyond ~ 5 minutes, the IR
460 optical depth is too high and too low, respectively, for reliable estimates to be
461 made. In this case the optical flow model output has been resampled to a grid
462 of one vector every 6 pixels which returns velocity values at ~ 20 m intervals.
463 The mean and standard deviation of the fine ash flux over the eruption cloud
464 is calculated and plotted for each image in the sequence lasting ~ 5 minutes
465 (Figure 9). A total fine ash mass emitted of ~ 4000 kg (4 t), a mean fine ash
466 mass flux of $\sim 53 \pm 26 \text{ kg s}^{-1}$, and a maximum fine ash mass flux (determined
467 from the peak of the Gaussian) of $\sim 84 \text{ kg s}^{-1}$ are found. Other discrete
468 explosions observed during the field campaign and analyzed in this way gave
469 similar results as those shown in Figure 9. An animation depicting the ash
470 mass flux results for a four minute period corresponding with the Stromboli
471 explosion shown in Figures 8 and 9 can be seen in Supplementary Material
472 2.

4.2. SO_2 results

During the period of measurements at Stromboli, the volcano was degassing SO_2 nearly continuously. SO_2 retrievals were made for sequences of measurements lasting 1–3 hours during the day and night. Figure 10 illustrates the nicAIR SO_2 retrieval that can be directly compared with the results presented in Kern et al. (2014) which includes results for the same plume from six different UV SO_2 camera systems. The range of values found for this particular explosion, by Kern et al. (2014) was 127–209 kg, compared with 161 kg determined from the nicAIR (IR) measurements. The range in average SO_2 masses emitted for four explosions analyzed by Kern et al. (2014) on this day was 51–160 kg, compared to results from Mori and Burton (2009), who used similar techniques in 2006 and found 15–40 kg SO_2 per explosion. Lopez et al. (2013) also did some comparisons between the nicAIR camera and a UV spectrometer system at Karymsky Volcano and found good agreement in SO_2 SCD. Many more comparisons are required before definitive assessments can be made concerning the accuracy of the IR SO_2 retrieval, but the experience at several different volcanoes suggests that the IR sensor is less sensitive to SO_2 (i.e. has a lower signal to noise ratio) in comparison with the UV camera sensors.

4.3. Karymsky

4.3.1. Volcano setting

Karymsky Volcano (54.05°N, 159.45°E, 1536 m) is a predominantly andesitic stratovolcano located in the Eastern Volcanic Front of Russia’s Kamchatka Peninsula. Since its current eruptive cycle began on 2 January 1996 it has been in a nearly continuous state of unrest, with activity consisting

498 of discrete explosive eruptions of ash and gas reaching altitudes up to 6 km,
 499 periodic effusion of lava flows and/or extrusion of summit domes, and various
 500 styles of frequently-audible degassing (e.g. jetting, chugging, etc.) (Johnson
 501 et al., 1998; Johnson and Lees, 2000; Fischer et al., 2002; Izbekov et al., 2004;
 502 Lopez et al., 2013). The various styles of activity exhibited at Karymsky Vol-
 503 cano have been attributed to changes in conduit permeability and magma
 504 recharge from depth (e.g. Fischer et al. (2002); Ozerov et al. (2003); John-
 505 son (2007); Lopez et al. (2013)). The frequent explosions and variable styles
 506 of activity exhibited at Karymsky Volcano have made it a laboratory-type
 507 volcano that has supported numerous multiparameter investigations (e.g.
 508 (Johnson et al., 1998; Johnson and Lees, 2000; Fischer et al., 2002; Lopez
 509 et al., 2013). Measurements of SO₂ fluxes from Karymsky volcano were first
 510 made by Fischer et al. (2002) in 1999 using a COSPEC (Correlation Spec-
 511 trometer) (Stoiber et al., 1983). They observed highly variable SO₂ fluxes,
 512 corresponding with cycles of eruptive activity, with values ranging from near
 513 detection limit to 1.63 kg s⁻¹. Similarly, Lopez et al. (2013) found variable
 514 SO₂ fluxes in 2011 and 2012 ranging from <0.14 kg s⁻¹ to 3.2 kg s⁻¹ using
 515 a FLYSPEC UV spectrometer (Horton et al., 2006) and the nicAIR cam-
 516 era. Lopez et al. (2013) estimated a total eruptive mass of very fine ash
 517 ($r < 16 \mu\text{m}$) for a relatively large explosive eruption at Karymsky observed
 518 on August 22, 2011 of >69000 kg using the nicAIR camera.

519 4.3.2. *Experimental setup*

520 Between 13–24 August 2011 the nicAIR camera was deployed ~3.58 km
 521 southeast of the summit vent (Figure 5). During this study period mea-
 522 surements of Karymsky’s volcanic emissions were collected. The instrument

viewed activity at a slant distance of 3690 m with a $\sim 15^\circ$ inclination angle. This viewing geometry resulted in an average pixel size of $\sim 12 \text{ m}^2$ (Table 2). Volcanic activity was characterized as highly variable, and included gas jetting; pulsatory degassing; discrete ash explosions, producing ash clouds to 500–1500 m above the vent; and explosive eruptions that frequently followed periods of apparent vent sealing and produced ash clouds to $>2000 \text{ m}$ above vent altitudes (Lopez et al., 2013). During this time period, background skies were typically cloud-free and the gas-only plume was primarily translucent. Occasionally, high winds blew the plume down the flank of the volcano towards the camera preventing accurate constraints on SO_2 and ash SCDs and derived fluxes. These non-ideal time periods were excluded from analyses.

4.3.3. *Ash results*

Retrievals of ash SCD and effective particle size were performed for a series of images when Karymsky was exhibiting ash-rich explosions that were relatively vigorous and in many cases produced optically thick plumes. In these cases the particle size retrievals presented here are more reliable than the method described by Prata and Bernardo (2009). Ash eruptions from Karymsky generally gave total fine-ash mass retrievals above 10,000 kg (10 t). Figure 11 shows a sequence of ash retrievals at $\sim 20\text{--}30 \text{ s}$ intervals for an ash-rich explosive eruption. The total masses retrieved (minimums) are 10000, 24000, 32000 and 53000 kg, respectively with increasing time from the start. This final value is not very different from that reported by Lopez et al. (2013) of $\sim 69000 \text{ kg}$. These values are likely to be conservative minimum estimates of the fine ash mass, due to optical depth saturation for thick plumes preventing ash absorption from being accurately constrained. An

548 example of this optical depth saturation can be seen in the panels of Figure 11
 549 at the top half of the growing plume, where the combination of greater path
 550 length and particle numbers leads to high optical depths, and underestimated
 551 ash-mass retrievals. Figure 12 shows an example of a smaller eruption, in
 552 which case the IR image was acquired sometime after the initial explosion,
 553 where the optical depth is lower over almost the entire column of ash, such
 554 that more accurate ash mass retrievals are possible. The fine ash mass for
 555 this smaller eruption is estimated to be ~ 10000 – 20000 kg. For both of these
 556 examples the optical flow method was used to determine plume velocities,
 557 allowing an estimated minimum fine ash mass flux of ~ 150 kg s $^{-1}$ to be
 558 calculated for the large explosion (Figure 11). Figure 13 shows the plume
 559 velocity vectors calculated for the large eruption sequence and an animation
 560 of this sequence is found in Supplementary Material 1. Figure 14 shows the
 561 fine ash mass flux retrieval for the last image in the sequence, for which the
 562 highest flux was found.

563 4.4. *Láscar*

564 4.4.1. *Volcano setting*

565 Láscar Volcano (23.37°S, 67.73°W, 5592 m) is an andesitic stratovolcano
 566 located in the Andean Central Volcanic Zone of Northern Chile, and is con-
 567 sidered the most active volcano in this region (Matthews et al., 1997; Tassi
 568 et al., 2009). Between 1984 and 1996 Láscar underwent a period of cyclic
 569 lava dome growth and subsidence, persistent degassing, and explosive erup-
 570 tions (Matthews et al., 1997). This eruptive period included a sub-Plinian
 571 eruption from 19–20 April of 1993 that produced an ash cloud up to 25 km
 572 altitude and produced over 0.1 km 3 of eruptive material (Matthews et al.,

1997). Explosive activity during this eruptive cycle was attributed to changes in lava dome permeability, which led to dome pressurization and culminated in pressure release during explosion (Matthews et al., 1997). Activity since 1993 has been dominated by persistent degassing and periodic small ash explosions with associated ash clouds to altitudes less than 10 km Tassi et al. (2009).

4.4.2. *Experimental Setup*

Measurements of SO₂ emissions were made using the nicAIR camera on 29-30 November 2012. During this time period volcanic activity was limited to persistent SO₂ degassing. The nicAIR camera was deployed between 6.5 and 6.7 km northwest of the actively degassing crater, giving a pixel size of $\sim 40 \text{ m}^2$ (Table 2).

4.4.3. *SO₂ results*

The SO₂ plume from Láscaar was visible on both days of measurements in November 2011, but generally remained close to the volcano with very little lofting. The plume was clearly seen in visible light, appearing white in color, presumably due to condensed water in the plume. Figure 1 illustrates the typical plume character during the measurement period. Figure 15 shows an 8.6 μm channel image (top) and the associated SO₂ retrieval. The small plume and minimum temperature difference between the cold background sky ($\sim 240\text{--}250 \text{ K}$) and the plume ($\sim 258 \text{ K}$, estimated from the broadband channel), resulted in low signal to noise, which made the retrieval difficult, and increased the error. The calibration of the camera was initially found to give temperatures biased 10-15 K too low. An improved calibration proce-

597 dure which takes into account the effect of self-radiation from the detector
 598 was subsequently introduced which increased the derived temperatures, but
 599 we suspect they may still be biased low by 1–3 K. Self-radiation is an impor-
 600 tant contribution when the focal plane array (FPA) temperature used when
 601 calibrating the camera in the laboratory is very different from the temper-
 602 ature of the scene viewed by the camera in the field. In this case plume
 603 temperatures are $\sim 255\text{--}260$ K, while the FPA temperature was ~ 290 K, so
 604 that the FPA contribution was significant and unaccounted for in the original
 605 calibration. To improve the detection capabilities, the retrieval was focused
 606 on a rectangular region immediately above the vent where the plume signal
 607 was the strongest. The plume temperature was estimated at the base of the
 608 plume in a region of 3×3 pixels, just above the vent and no correction was
 609 made for the temperature lapse rate. Wind speeds were estimated using the
 610 optical flow method described in Section 3.7. The mean flux for this hour is
 611 close to 1.5 ± 0.7 kg s $^{-1}$ (~ 130 t day $^{-1}$), but the variability is high ($\sim 50\%$ of
 612 the signal) as can be seen in Figure 16. We estimate that approximately half
 613 of the variability may be attributed to retrieval error, poor calibration and
 614 errors in estimating the wind speed and plume temperature. Mather et al.
 615 (2004) found fluxes of 28 kg s $^{-1}$ in January 2003, which are much higher than
 616 the values found here. Henney et al. (2012) reported SO $_2$ gas measurements
 617 for 7 December, 2004 from a mini-UV spectrometer and estimated fluxes of
 618 1.9–2.4 kg s $^{-1}$, similar to the mean values we find.

619 5. Conclusions

620 **Ground-based** thermal infrared imaging of volcanic plumes at relatively
621 cool temperatures, close to ambient atmospheric values, is a new technique
622 that offers the potential to infer fine ash and SO₂ SCDs and masses at high
623 temporal resolution. When fine ash mass retrievals are combined with an
624 *a priori* particle size distribution model, total ash mass for the size frac-
625 tion $r < 50 \mu\text{m}$ can be calculated. This provides a potentially significant
626 tool to quickly and remotely estimate plume fine ash fluxes and character-
627 ize eruption size, with important implications for volcano monitoring and
628 hazard mitigation. We note that ash mass retrievals for opaque plumes will
629 represent minimum estimates and error in the resulting PSD will likely be
630 high. An optical flow algorithm was used to determine plume velocities and
631 a scheme was proposed to combine ash and SO₂ SCDs with plume velocities
632 to infer ash and SO₂ fluxes, using a single consistent dataset. The ability to
633 determine these parameters continuously over 24 hours also make the cam-
634 era useful as a monitoring tool and for alerting communities about changes
635 in volcanic activity. Examples of retrievals at Stromboli (Italy), Karymsky
636 (Russia) and Láscar (Chile) volcanoes demonstrate the versatility of the sys-
637 tem; since the instrument has the same capability during the night as the day,
638 the measurements add value and complement those from other volcanological
639 tools (e.g. the UV SO₂ camera). The fine ash mass fluxes retrieved from the
640 camera are in the same range of values determined from other studies using
641 different instruments and methodologies. Future work will involve a robust
642 validation of total ash and SO₂ masses and fluxes, and a scheme will be devel-
643 oped to simultaneously quantify both ash and SO₂, significantly improving

644 the utility of these measurements. Preliminary results suggest errors in fine
645 ash SCDs in the range of 20–50% for semi-transparent plumes (with larger
646 unconstrained error for opaque plumes), and errors of $\sim 20\%$ for SO_2 SCDs,
647 with most of this error due to the fine ash mass and gas retrievals rather
648 than wind velocity errors—usually a significant source of error in determin-
649 ing fluxes from independent wind data. The parameters determined by the
650 camera are ideally suited for use in conjunction with other data to diagnose
651 volcanic activity. In particular, the maximum height of the volcanic column,
652 and fluxes of fine ash and SO_2 gas found from the nicAIR imagery may be
653 used to initialize dispersion models for forecasting atmospheric transport of
654 volcanic ash and gas.

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664 Appendix: Optical depth retrieval

665 The monochromatic optical path between two points, $\mathbf{s}_1$ and $\mathbf{s}_2$ in a cloud
666 of particles may be written,

$$\tau_\lambda(s_1, s_2) = \int_{s_1}^{s_2} \beta_{ext}(\lambda, s) ds, \quad (\text{A-1})$$

667 where β_{ext} is an extinction coefficient with dimensions m^{-1} at wavelength λ .
668 It is often more useful to consider the mass extinction along the path and
669 reformulate (A-1) as,

$$\tau_\lambda(s_1, s_2, r) = \int_{s_1}^{s_2} \int_0^\infty \pi r^2 Q_{ext}(\lambda, r) \frac{dN}{dr}(r) dr ds, \quad (\text{A-2})$$

670 where now the dimensionless extinction efficiency factor $Q_{ext}(\lambda, r)$ replaces
671 $\beta_{ext}(\lambda, r)$. The mass extinction is an integral over the size distribution $\frac{dN}{dr}$
672 where N is the number of particles per unit volume and the particle radius
673 is r . It is now assumed that the size distribution and mass extinction are
674 independent of position ($\mathbf{s}$):

$$\tau_\lambda = L \int_0^\infty \pi r^2 Q_{ext}(\lambda, r) \frac{dN}{dr}(r) dr, \quad (\text{A-3})$$

675 where $L=|\mathbf{s}_1 - \mathbf{s}_2|$. It is commonly found in studies of volcanic plumes that
676 the particle sizes follow a log-normal distribution :

$$\frac{dN}{dr} = \frac{1}{r} \frac{N_o}{\sqrt{2\pi \ln \sigma}} \exp \left\{ - \left(\frac{\ln r - \ln \mu}{\ln \sigma} \right)^2 \right\}, \quad (\text{A-4})$$

677 where μ and σ are the geometric mean and standard deviation of the dis-
678 tribution. The factor $1/\sqrt{2\pi}$ ensures that the integral of the distribution
679 from $r=0$ to $r=\infty$ equals N_o , the total number of particles per unit vol-
680 ume. Equation (A-2) shows that, in principle, from a set of measurements
681 of optical depths it is possible to infer the particle size distribution. The
682 measurements may be spectral, for example from an FT-IR device, or they
683 may be at one wavelength but at different angles, for example in a device
684 that is able to perform zenith scans. In our case, the measurements consist of
685 a large number ($0.01\text{--}0.1 \times 10^6$) of image pixels, each measuring a different
686 part of the same ash plume.

687 Let us now consider a monochromatic (λ) thermal image of M (l lines x
688 p pixels) measurements of a cloud of ash for which the size distribution is
689 log-normal with $\mu=\mu_k$ and $\sigma=\sigma_k$. If the optical depth image is defined as
690 $\tau_\lambda(L_i)$ then in matrix notation:

$$\mathbf{\Upsilon} = \mathbf{KH}^T, \quad (\text{A-5})$$

and

$$\mathbf{\Upsilon} = \begin{pmatrix} \tau_i \\ \tau_{i+1} \\ \vdots \\ \tau_M \end{pmatrix},$$

691 is a vector of measurements. $\mathbf{H}$ is a vector representing the size distribution,
692 with each element given by:

$$H_j = h_j dr, \quad (\text{A-6})$$

where dr is the particle size increment and $j=1 \cdots J$. The kernel $\mathbf{K}$ is a matrix with elements

$$\mathbf{K} = \begin{pmatrix} q_{1,1} & q_{1,2} & q_{1,3} & \cdots & q_{1,J} \\ q_{2,1} & q_{2,2} & q_{2,3} & \cdots & q_{2,J} \\ \vdots & \vdots & \vdots & \vdots & \\ q_{M,1} & q_{M,2} & q_{M,3} & \cdots & q_{M,J} \end{pmatrix},$$

693 where each element may be interpreted as an extinction factor for a particular
694 pixel and particle size. Figure 17 shows the variation of the kernels with
695 particle size for different wavelengths and Figure 18 shows the variation of
696 the integrand as a function of particle size for several different values of μ_k
697 with $\sigma=1.5$. These functions have been determined using a Mie scattering
698 calculation for spherical andesitic particles and log-normal size distributions.
699 The natural way to solve (A-5) is simply to write:

$$\mathbf{H} = \mathbf{K}^{-1}\mathbf{\Upsilon}, \quad (\text{A-7})$$

700 however this is known to be an ill-posed problem and it is clear that the
701 columns of $\mathbf{K}$ may be linearly dependent. As an illustration suppose the
702 kernel function is chosen to be a constant, q_0 independent of particle size.
703 Then the integral in (A-3) can be evaluated immediately to give:

$$\tau_i = L_i \pi q_o N_o \exp\{2(\ln \mu + \ln^2 \sigma)\}. \quad (\text{A-8})$$

704 Given q_o and assuming that L_i are known independently, then parameters of

705 the (assumed) size distribution are uniquely determined. Unfortunately each
 706 of the L_i are linearly dependent, leading to multiple solutions. The retrieval
 707 scheme chosen here relies on the correspondence between the distribution of
 708 particle sizes and the distribution of optical depths as shown in (A-2). Rather
 709 than attempting to retrieve particle sizes directly by inverting $\mathbf{K}$, through de-
 710 termining its elements using Mie scattering, the underlying form of the PSD
 711 is determined by fitting log-normals to the optical depth distribution. In
 712 this scheme there is a scaling between the particle size parameters (μ and σ)
 713 proportional to $\mathbf{K}^{-1}$. This scaling is a constant factor in the current scheme
 714 and is chosen based on the premise that the particle sizes are likely to be
 715 large, *viz.* $r > 16 \mu\text{m}$. Prata (1989) showed that the extinction efficiency for
 716 large particles tends towards a constant value, while more recently Francis
 717 et al. (2012) show that for several different ash compositions and for three
 718 wavelengths (*viz.*, 10.8, 12.0 and 13.4 μm) the scaled mass extinction coeffi-
 719 cient tends towards a value of $\sim 0.02 \text{ m}^2 \text{ g}^{-1}$. Figure 17 also shows that for
 720 particle sizes greater than $\sim 30 \mu\text{m}$, $Q_{ext} \rightarrow 2.0$, a constant. The assumption
 721 of an underlying log-normal size distribution can be changed to assume that
 722 the distribution is the linear sum of several log-normals, or indeed any PSD.
 723 The choice can be informed by examining the distribution of the measured
 724 optical depths, as has been done here.

725 References

- 726 Allard, P., Aiuppa, A., Burton, M., Caltabiano, T., Federico, C., Salerno, G.,
 727 La Spina, A., 2008. Crater gas emissions and the magma feeding system

728 of Stromboli volcano. The Stromboli Volcano: An Integrated Study of the
729 2002-2003 Eruption, 65–80.

730 Allard, P., Carbonnelle, J., Metrich, N., Loyer, H., Zettwoog, P., 1994. Sul-
731 phur output and magma degassing budget of Stromboli volcano. *Nature*
732 368 (6469), 326–330.

733 Barberi, F., Rosi, M., Sodi, A., 1993. Volcanic hazard assessment at Strom-
734 boli based on review of historical data. *Acta Vulcanologica* 3, 173–187.

735 Barron, J. L., Fleet, D. J., Beauchemin, S. S., 1994. Performance of optical
736 flow techniques. *International Journal of Computer Vision* 12 (1), 43–77.

737 Bertagnini, A., Coltelli, M., Landi, P., Pompilio, M., Rosi, M., 1999. Violent
738 explosions yield new insights into dynamics of Stromboli volcano. *Eos,*
739 *Transactions American Geophysical Union* 80 (52), 633–636.

740 Blackburn, E., Wilson, L., Sparks, R., 1976. Mechanisms and dynamics of
741 Strombolian activity. *Journal of the Geological Society* 132 (4), 429–440.

742 Bluth, G., Shannon, J., Watson, I., Prata, A., Realmuto, V., 2007. Develop-
743 ment of an ultra-violet digital camera for volcanic SO₂ imaging. *Journal*
744 *of Volcanology and Geothermal Research* 161 (1), 47–56.

745 Bonadonna, C., Phillips, J., 2003. Sedimentation from strong volcanic
746 plumes. *Journal of Geophysical Research* 108 (B7), 2340.

747 Bradski, G., Kaehler, A., 2008. Learning OpenCV: Computer vision with the
748 OpenCV library. O'Reilly Media, Inc.

749 Burton, M., Caltabiano, T., Murè, F., Salerno, G., Randazzo, D., 2009. SO_2
750 flux from Stromboli during the 2007 eruption: Results from the flame net-
751 work and traverse measurements. *Journal of Volcanology and Geothermal*
752 *Research* 182 (3), 214–220.

753 Calvari, S., Spampinato, L., Bonaccorso, A., Oppenheimer, C., Rivalta, E.,
754 Boschi, E., 2011. Lava effusion – A slow fuse for paroxysms at Stromboli
755 volcano? *Earth and Planetary Science Letters* 301 (1), 317–323.

756 Carter, A. J., Ramsey, M. S., Belousov, A. B., 2007. Detection of a new
757 summit crater on Bezymianny Volcano lava dome: satellite and field-based
758 thermal data. *Bulletin of Volcanology* 69 (7), 811–815.

759 Chouet, B., Hamisevicz, N., McGetchin, T. R., 1974. Photoballistics of vol-
760 canic jet activity at Stromboli, Italy. *Journal of Geophysical Research*
761 79 (32), 4961–4976.

762 Delle Donne, D., Lacanna, G., Marchetti, E., Ripepe, M., Ulivieri, G., 2006.
763 Monitoring explosive volcanic activity using thermal images, Stromboli
764 volcano, Italy. In: *American Geophysical Union Fall Meeting Abstracts*.
765 Vol. 1. p. 1795.

766 Delle Donne, D., Ripepe, M., 2012. High-frame rate thermal imagery of
767 Strombolian explosions: Implications for explosive and infrasonic source
768 dynamics. *Journal of Geophysical Research: Solid Earth* 117 (B9).

769 Farneback, G., 2001. Very high accuracy velocity estimation using orientation
770 tensors, parametric motion, and simultaneous segmentation of the motion
771 field. In: *Computer Vision, 2001. International Conference on Computer*

- 772 Vision 2001. Proceedings. Eighth Institute of Electrical and Electronic En-
773 gineers International Conference. Vol. 1. IEEE, pp. 171–177.
- 774 Farnebäck, G., 2003. Two-frame motion estimation based on polynomial ex-
775 pansion. In: Image Analysis. Springer, pp. 363–370.
- 776 Fischer, T. P., Roggensack, K., Kyle, P. R., 2002. Open and almost shut case
777 for explosive eruptions: vent processes determined by SO₂ emission rates
778 at Karymsky volcano, Kamchatka. *Geology* 30 (12), 1059–1062.
- 779 Francis, P. N., Cooke, M. C., Saunders, R. W., 2012. Retrieval of physical
780 properties of volcanic ash using Meteosat: A case study from the 2010
781 Eyjafjallajökull eruption. *Journal of Geophysical Research* 117 (D00U09).
782 URL <http://dx.doi.org/10.1029/2011JD016788>
- 783 Gibson, J., 1950. *The Perception of the Visual World*. Oxford, England:
784 Houghton Mifflin. (1950). xii 242 pp.
- 785 Harris, A., Delle Donne, D., Dehn, J., Ripepe, M., Worden, A., 2013. Vol-
786 canic plume and bomb field masses from thermal infrared camera imagery.
787 *Earth and Planetary Science Letters* 365, 77–85.
- 788 Harris, A., Ripepe, M., 2007a. Synergy of multiple geophysical approaches
789 to unravel explosive eruption conduit and source dynamics—A case study
790 from Stromboli. *Chemie der Erde-Geochemistry* 67 (1), 1–35.
- 791 Harris, A., Ripepe, M., 2007b. Temperature and dynamics of degassing at
792 Stromboli. *Journal of Geophysical Research: Solid Earth* 112 (B3).

793 Harris, A. J., Dehn, J., Calvari, S., 2007. Lava effusion rate definition and
794 measurement: a review. *Bulletin of Volcanology* 70 (1), 1–22.

795 Henney, L., Rodríguez, L., Watson, I., 2012. A comparison of SO₂ retrieval
796 techniques using mini-UV spectrometers and ASTER imagery at Láscar
797 volcano, Chile. *Bulletin of Volcanology* 74 (2), 589–594.

798 Horton, K. A., Williams-Jones, G., Garbeil, H., Elias, T., Sutton, A. J.,
799 Mougini-Mark, P., Porter, J. N., Clegg, S., 2006. Real-time measurement
800 of volcanic SO₂ emissions: validation of a new UV correlation spectrometer
801 (FLYSPEC). *Bulletin of Volcanology* 68 (4), 323–327.

802 Izbekov, P. E., Eichelberger, J. C., Ivanov, B. V., 2004. The 1996 eruption of
803 Karymsky volcano, Kamchatka: historical record of basaltic replenishment
804 of an andesite reservoir. *Journal of Petrology* 45 (11), 2325–2345.

805 Jähne, B., Haussecker, H., Scharr, H., Spies, H., Schmundt, D., Schurr,
806 U., 1998. Study of dynamical processes with tensor-based spatiotemporal
807 image processing techniques. In: *Computer Vision European Conference*
808 *on Computer Vision 1998*. Springer, pp. 322–336.

809 Jaupart, C., Vergnolle, S., 1988. Laboratory models of Hawaiian and Strom-
810 bolian eruptions. *Nature* 331 (6151), 58–60.

811 Johnson, J., Lees, J., 2000. Plugs and chugs: seismic and acoustic observa-
812 tions of degassing explosions at Karymsky, Russia and Sangay, Ecuador.
813 *Journal of Volcanology and Geothermal Research* 101 (1), 67–82.

814 Johnson, J. B., 2007. On the relation between infrasound, seismicity, and

815 small pyroclastic explosions at Karymsky Volcano. *Journal of Geophysical*
816 *Research: Solid Earth* 112 (B8).

817 Johnson, J. B., Lees, J. M., Gordeev, E. I., 1998. Degassing explosions
818 at Karymsky volcano, Kamchatka. *Geophysical Research Letters* 25 (21),
819 3999–4002.

820 Kern, C., Lbcke, P., Bobrowski, N., Campion, R., Mori, T., Smekens,
821 J.-F., Stebel, K., Tamburello, G., Burton, M., Platt, U., Prata, F., 2014.
822 Intercomparison of so₂ camera systems for imaging volcanic gas plumes.
823 *Journal of Volcanology and Geothermal Research* (0), –.
824 URL <http://www.sciencedirect.com/science/article/pii/S0377027314002662>

825 Lautze, N. C., Houghton, B. F., 2005. Physical mingling of magma and
826 complex eruption dynamics in the shallow conduit at Stromboli volcano,
827 Italy. *Geology* 33 (5), 425–428.

828 Liu, H., Hong, T.-H., Herman, M., Chellappa, R., 1997. A general motion
829 model and spatio-temporal filters for computing optical flow. *International*
830 *Journal of Computer Vision* 22 (2), 141–172.

831 Liu, Y., Arnott, W. P., Hallett, J., 1999. Particle size distribution retrieval
832 from multispectral optical depth: Influences of particle nonsphericity and
833 refractive index. *Journal of Geophysical Research: Atmospheres* 104 (D24),
834 31753–31762.

835 Lopez, T., Fee, D., Prata, F., Dehn, J., 2013. Characterization and inter-
836 pretation of volcanic activity at Karymsky Volcano, Kamchatka, Russia,

837 using observations of infrasound, volcanic emissions, and thermal imagery.
838 Geochemistry, Geophysics, Geosystems 14 (12), 5106–5127.

839 Lucas, B. D., Kanade, T., et al., 1981. An iterative image registration tech-
840 nique with an application to stereo vision. In: International Joint Confer-
841 ence on Artificial Intelligence. Vol. 81. pp. 674–679.

842 Marchetti, E., Ripepe, M., Harris, A., Delle Donne, D., 2009. Tracing the
843 differences between Vulcanian and Strombolian explosions using infrasonic
844 and thermal radiation energy. Earth and Planetary Science Letters 279 (3),
845 273–281.

846 Mather, T., Tsanev, V., Pyle, D., McGonigle, A., Oppenheimer, C., Allen,
847 A., 2004. Characterization and evolution of tropospheric plumes from Las-
848 car and Villarrica volcanoes, Chile. Journal of Geophysical Research: At-
849 mospheres 109 (D21).

850 Matthews, S. J., Gardeweg, M. C., Sparks, R. S. J., 1997. The 1984 to
851 1996 cyclic activity of Lascar Volcano, northern Chile: cycles of dome
852 growth, dome subsidence, degassing and explosive eruptions. Bulletin of
853 Volcanology 59 (1), 72–82.

854 Mori, T., Burton, M., 2009. Quantification of the gas mass emitted during
855 single explosions on stromboli with the SO_2 imaging camera. Journal of
856 Volcanology and Geothermal Research 188 (4), 395–400.

857 Oppenheimer, C., Yirgu, G., 2002. Thermal imaging of an active lava
858 lake: Erta’Ale volcano, Ethiopia. International Journal of Remote Sensing
859 23 (22), 4777–4782.

- 860 Ozerov, A., Ispolatov, I., Lees, J., 2003. Modeling strombolian eruptions
861 of Karymsky volcano, Kamchatka, Russia. *Journal of Volcanology and*
862 *Geothermal Research* 122 (3), 265–280.
- 863 Patrick, M., 2007. The gas content and buoyancy of Strombolian ash plumes.
864 *Journal of Volcanology and Geothermal Research* 166 (1), 1–6.
- 865 Patrick, M. R., Harris, A. J., Ripepe, M., Dehn, J., Rothery, D. A., Calvari,
866 S., 2007. Strombolian explosive styles and source conditions: insights from
867 thermal (FLIR) video. *Bulletin of Volcanology* 69 (7), 769–784.
- 868 Prata, A., 1989. Infrared radiative transfer calculations for volcanic ash
869 clouds. *Geophysical Research Letters* 16 (11), 1293–1296.
- 870 Prata, A., Bernardo, C., 2009. Retrieval of volcanic ash particle size, mass
871 and optical depth from a ground-based thermal infrared camera. *Journal*
872 *of Volcanology and Geothermal Research* 186 (1), 91–107.
- 873 Prata, A., Bernardo, C., 2014. Retrieval of sulphur dioxide from a ground-
874 based thermal infrared imaging camera. *Atmospheric Measurements and*
875 *Techniques Discussions* 7, 1153–1211.
- 876 Prata, A., Grant, I., 2001. Retrieval of microphysical and morphological
877 properties of volcanic ash plumes from satellite data: Application to Mt
878 Ruapehu, New Zealand. *Quarterly Journal of the Royal Meteorological So-*
879 *ciety* 127 (576), 2153–2179.
- 880 URL <http://dx.doi.org/10.1002/qj.49712757615>
- 881 Realmuto, V., Sutton, A., Elias, T., 1997. Multispectral thermal infrared
882 mapping of sulfur dioxide plumes: A case study from the East Rift Zone

883 of Kilauea Volcano, Hawai'i. *Journal of Geophysical Research: Solid Earth*
884 102 (B7), 15057–15072.

885 Realmuto, V. J., Abrams, M. J., Buongiorno, M. F., Pieri, D. C., 1994. The
886 use of multispectral thermal infrared image data to estimate the sulfur
887 dioxide flux from volcanoes: a case study from Mount Etna, Sicily, July
888 29, 1986. *Journal of Geophysical Research: Solid Earth* 99 (B1), 481–488.

889 Ripepe, M., Donne, D. D., Harris, A., Marchetti, E., Ulivieri, G., 2008.
890 Dynamics of Strombolian activity. *Geophysical Monograph Series* 182, 39–
891 48.

892 Ripepe, M., Harris, A. J., Carniel, R., 2002. Thermal, seismic and infra-
893 sonic evidences of variable degassing rates at Stromboli volcano. *Journal*
894 *of Volcanology and Geothermal Research* 118 (3), 285–297.

895 Ripepe, M., Harris, A. J., Marchetti, E., 2005a. Coupled thermal oscillations
896 in explosive activity at different craters of Stromboli volcano. *Geophysical*
897 *Research Letters* 32 (17).

898 Ripepe, M., Marchetti, E., Ulivieri, G., Harris, A., Dehn, J., Burton, M.,
899 Caltabiano, T., Salerno, G., 2005b. Effusive to explosive transition during
900 the 2003 eruption of Stromboli volcano. *Geology* 33 (5), 341–344.

901 Ripepe, M., Rossi, M., Saccorotti, G., 1993. Image processing of explosive
902 activity at Stromboli. *Journal of Volcanology and Geothermal Research*
903 54 (3), 335–351.

904 Rose, W., Durant, A., 2009. Fine ash content of explosive eruptions. *Journal*
905 *of Volcanology and Geothermal Research* 186 (1–2), 32–39.

- 906 Rosi, M., Bertagnini, A., Landi, P., 2000. Onset of the persistent activity at
907 Stromboli volcano (Italy). *Bulletin of Volcanology* 62 (4-5), 294–300.
- 908 Stoiber, R., Malinconico, L., Williams, S., 1983. Use of the correlation spec-
909 trometer at volcanoes. *Forecasting Volcanic Events* 1, 425–444.
- 910 Tassi, F., Aguilera, F., Vaselli, O., Medina, E., Tedesco, D., Huertas, A. D.,
911 Poreda, R., Kojima, S., 2009. The magmatic-and hydrothermal-dominated
912 fumarolic system at the Active Crater of Lascar volcano, northern Chile.
913 *Bulletin of Volcanology* 71 (2), 171–183.
- 914 Wen, S., Rose, W. I., 1994. Retrieval of sizes and total masses of particles
915 in volcanic clouds using AVHRR bands 4 and 5. *Journal of Geophysical*
916 *Research* 99 (D3), 5421–5431.
- 917 Wessels, R. L., Vaughan, R. G., Patrick, M. R., Coombs, M. L., 2013. High-
918 resolution satellite and airborne thermal infrared imaging of precursory
919 unrest and 2009 eruption at Redoubt Volcano, Alaska. *Journal of Vol-*
920 *canology and Geothermal Research* 259, 248–269.
- 921 Wohletz, K. H., 1983. Mechanisms of hydrovolcanic pyroclast formation:
922 grain-size, scanning electron microscopy, and experimental studies. *Journal*
923 *of Volcanology and Geothermal Research* 17 (1), 31–63.

Table 1: nicAIR technical specifications.

Total field-of-view	32° x 24°
Optics	35 mm F1.4 Ge lens
Image size	640 x 512 pixels
Number of filters	up to 4
Channel 1 (SO ₂)	8.6 μm
Channel 2 (Plume temperature)	10.0 μm
Channel 3 (Ash)	11.0 μm
Channel 4 (Ash)	12.0 μm or 8–13 μm (broadband)
Sampling rate (max for single channel;)	~1 Hz
Detector	Uncooled microbolometer
NE Δ T Filter 1	500 mK @ 250 K
NE Δ T Filter 2	200 mK @ 250 K
NE Δ T Filter 3	200 mK @ 250 K
NE Δ T Filter 4	200 mK @ 250 K
NE Δ T Broadband	50 mK @ 250 K
Requirements	12 V, 3 A, 40 W peak
Accuracy (SO ₂)	$\pm 0.2 \text{ g m}^{-2}$ (or $\pm 20\%$, whichever is higher)
Accuracy (silicate particles) ¹	$\pm 0.5 \text{ g m}^{-2}$ (or $\pm 40\text{--}60\%$, whichever is higher)
Detection range	~10 km
Operating temperature range	-10 °C to +50 °C
Weight	8 kg
Dimensions	300 mm x 160 mm x 160 mm (LxWxH)

¹ For particles in the range 1–16 μm radius.

Table 2: Logistical information on the sites and measurements made at four different volcanoes.

Volcano	Stromboli	Karymsky	Láscar	Láscar
Date	Jun 23, 2013	Aug 13–24, 2011	Nov 29, 2012	Nov 30, 2012
Vent Latitude	38.793°N	54.048°N	23.363°S	23.363°S
Vent Longitude	15.211°E	159.443°W	67.732°W	67.732°W
Height [†] (m.a.s.l.)	800	1500	5320	5320
Camera Latitude	38.8068°N	54.0244°N	23.3046°S	23.3083°S
Camera Longitude	15.2346°E	159.4803°W	67.7483°W	67.7546°W
Camera Altitude (m.a.s.l.)	176	605	4386	4283
R-distance* (m)	624	900	930	1040
H-distance [‡] (m)	2560	3580	6710	6510
S-distance* (m)	2630	3690	6780	6600
Inclination Angle (°)	13.7	14.9	7.9	9.0
Center Pixel Size (m ²)	7	12.0	40	39
Measurement type	SO ₂ , ash, ϱ_s	ϱ_a	SO ₂ , ϱ_s	SO ₂ , ϱ_s

[†]Highest Altitude of Volcano between Camera and Plume (m)

*Relief to Plume Base (m) [‡] Horizontal Distance between Camera and Vent (m)

*Slant Distance between Camera and Plume (m) ϱ_s is the SO₂ flux in g s⁻¹.

ϱ_a is the ash flux in g s⁻¹.

Figure 1: Photograph of the nicAIR camera viewing the SO₂ plume from Láscar. The character of the plume, low altitude and close to the volcano was typical of its behavior during the measurements on 29–30 November 2012.

Figure 2: Broadband (8–13 μm) infrared images of Strombolian ash explosions. Each image is separated in time by ~ 20 s; a time stamp is shown at the bottom right of each image frame. Images are shown sequentially from top-left (earliest) to bottom right (latest). Data acquired on 23 June 2012 during the Stromboli Plume Workshop.

Figure 3: Broadband (8–13 μm) infrared image showing an ‘ashy’ Strombolian plume. Vertical temperature profiles inside the plume (black solid line) and outside the plume (dashed line) illustrate the plume and environmental vertical temperature variation. The plume temperature is taken to be the maximum of the profile and is corrected for a constant lapse rate determined from the profile outside the plume. Horizontal temperature variations in the plume are assumed to be solely due to opacity variations.

Figure 4: Motion vectors determined using broadband IR imagery at ~ 1 min intervals for an ash explosions at Stromboli. During the study period the ash plumes tended to reach neutral buoyancy within ~ 2 – 5 minutes and then migrated towards the southeast in the light winds. In this example three separate explosions can be seen.

Figure 5: Topographic data for the three volcanoes studied: A) Stromboli (Italy), B) Karymsky (Russia), and C) Láscar (Chile). The location of the camera is indicated on each map by a white square and the location of the active vent by a red triangle.

Figure 6: Four image pairs of an ash explosion at Stromboli on 23 June 2013: left-hand panels are the broadband temperature image and the right-hand panels show the optical depth retrievals for the elliptical regions shown. A two parameter log-normal size distribution fit to the data is shown for each image in the inset graphic. (a) 18:05 UT, (b) 18:06 UT, (c) 18:07 UT, and (d) 18:09 UT.

Figure 7: Temporal evolution of the two size modes found for the Stromboli ash eruptions. Mode 1 contains particles with particle radii $>10 \mu\text{m}$. Mode 2 is a fine mode with mean particle radii $\sim <10 \mu\text{m}$. Both modes decrease in size from the start to the end of the sequence. The fine ash mass determined from the data (grey colored bars) is $<2000 \text{ kg}$ for this small ash explosion. The time is minutes past 18:00 UTC on 23 June 2013.

Figure 8: (a) Broadband image, (b) ash SCD and, (c) fine ash mass flux, derived from infrared image data for a discrete Strombolian eruption at 18:27:09 UT on 23 June 2013.

Figure 9: Temporal evolution of the fine ash mass (grey-color bars) and flux (black filled circles) for a single explosion from Stromboli over the observed lifetime from first appearance in the atmosphere until last appearance when the optical depth drops below the nicAIR detection capability. Also, shown as a dashed line is a Gaussian profile centered at 26.5 minutes with a spread of 1.1 minutes and a maximum mass flux of 84 kg s^{-1} . The time is minutes past 18:00 UT on 23 June 2013.

Figure 10: SO₂ mass determined from the nicAIR camera at $\sim 16:04$ UT on 26 June 2012, when six UV cameras were operating. See Kern et al. (2014) for results from the UV retrievals. Units are given as ppm*m for consistency with the results presented in Kern et al. (2014).

Figure 11: Sequence of four images of a large ash-rich explosion from Karymsky Volcano on 22 August 2011. The left-hand panel in each of the sequences shows the brightness temperature in the broadband channel, and the right-hand panel shows the optical depth with an inset panel showing the size distribution and a two-parameter log-normal fit. (a) 08:22:06 UT, (b) 08:22:34 UT, (c) 08:22:53 UT and (d) 08:23:31 UT.

Figure 12: Two nicAIR images of a small ash explosion from Karymsky on 18 August 2011. The images show the ash plumes some time after the initial explosion. In these cases the optical depth of the column is mostly small and the total fine ash masses retrieved is more reliable. (a) Image acquired at 07:23:46 UT, and (b) image acquired at 07:26:31 UT.

Figure 13: Eruption velocity vectors determined using the optical flow algorithm for the same Karymsky ash eruption on 22 August 2011 as Figure 11. The times of acquisition of each image are shown in each panel.

Figure 14: Fine ash mass flux determined for the large Karymsky eruption determined using the plume velocity and fine ash mass retrievals for the last image in the sequence shown in Figure 13.

Figure 15: *Top:* 8.6 μm image of the Láscair SO_2 plume on 29 November 2012. *Bottom:* SO_2 retrieval for a section of the image that includes the plume. The total mass of SO_2 in this section is $\sim 64 \pm 20$ kg.

Figure 16: Time series of SO₂ fluxes (top panel), wind speed (m s⁻¹) (middle panel) and plume temperature (K) (bottom panel), for Láscair on 29 November 2011. The data have been smoothed with a filter over a ~ 1 min interval. The wind speeds are determined approximately every 1 minute and then smoothed to reduce noise. The mean and standard deviation of the flux were determined over 1 hr of measurements. Error bars are shown for all of the parameters. The plume temperature error is assumed to be ± 3 K and the wind speed error is estimated at ± 1 ms⁻¹, or $\sim \pm 20\%$ of the magnitude of the wind speed.

Figure 17: Extinction kernels for different monochromatic wavelengths as a function of particle size. The shaded region (grey) shows the particle size range of validity. Outside these limits the optical depth retrieval is inaccurate.

Figure 18: Variation of the product of the kernel function and size distribution (the integrand in equation A-3) as a function of particle size for several different log-normal PSDs with different means but the same standard deviation of $\sigma=1.6 \mu\text{m}$, computed at $\lambda=12 \mu\text{m}$. The shaded region (grey) shows the particle size range of validity.