

# Ash sedimentation by fingering and sediment thermals from wind-affected volcanic plumes

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1	Ash Sedimentation by Fingering and Sediment Thermals from
2	Wind-Affected Volcanic Plumes
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18	ABSTRACT
19	Ash fallout and volcanic plume dispersion represent critical hazards for local and global human
20	populations for minutes to years after the onset of an eruption. Understanding the key processes

- 21 governing the sedimentation of ash particles is a major challenge in modern volcanology from
- 22 modelling and risk management perspectives. Recent experiments predict that sedimentation

23 from eruption clouds rich in fine-grained ash can be driven by convective phenomena in the 24 form of ~100 m to km-scale ash fingers related to the intermittent formation and detaching of 25 ash-rich particle boundary layer. Remote sensing observations of mammatus and cloud veils from numerous eruptions over recent years as well as field observations of spatially 26 27 discontinuous ash deposits are consistent with this prediction. Indeed, this mode of ash 28 sedimentation is predicted to be the predominant mechanism of fine ash removal for a 29 significant fraction of eruptions in the geological records. Here, we use a novel combination of 30 3 millimeter-wavelength Doppler radar observations and time series of optical disdrometer data 31 to characterize for the first time in real-time the time-varying structure and sedimentation 32 properties of volcanic ash plumes under steady wind conditions. As a case study, we apply this 33 new method to weak short-lived plumes from Stromboli Volcano. 96% of the disdrometer 34 proximal sedimentation data highlight pulsatory phases of increased sedimentation rate that are 35 20-60 s apart and characterized by particle size distribution variation with bulk concentrations 36 up to  $681 \text{ mg/m}^3$ . Radar data also record intermittent periods of higher reflectivity (*i.e.* a factor 37 of 3 in mass concentration) inside the ash sedimentation interspersed by 30 to 50 s and 38 interpreted as ash fingers crossing the radar beam. From time series of radar signals and ground-39 based disdrometer measurements, together with simple analog experiments, we develop a 40 conceptual model for intermittent sedimentation from wind-affected ash plumes. In particular, 41 we show that when wind speeds are comparable to or greater than ash settling velocities, the dynamics of wind-driven rolls in ash clouds can govern the production of gravitational 42 43 instabilities and the occurrence and timing of ash fingers that form descending sediment 44 thermals in turn. This study suggests that ambient wind should affect the maximum size and 45 minimum concentration of ash particles needed to form fingers but also control where and when 46 fingers form at the base of wind-drifted ash plumes. These novel predictions highlight the need for future work aimed at refining our understanding of ash finger formation under windy
conditions from the perspectives of in-situ characterization, numerical modelling and risk
assessment of fine ash dispersal from the most frequent style of eruptions on Earth.

50 Highlights:

- Novel observations of time-dependent ash sedimentation using a 3 mm-wavelength
   Doppler radar and a ground-based disdrometer.
- Wind-induced counter-rotating rolls lead to the formation of ash fingers evolving into
   sediment thermals.

Wind controls when and where ash fingers form at the base of wind-drifted ash plumes.
Keywords: volcanic plume; ash sedimentation; ash finger; sediment thermal; Doppler radar;
optical disdrometer

#### 58 **1 Introduction**

59 Ash plumes produced by explosive eruptions, such as the 2010 Eyjafjallajökull event, 60 disrupt air travel and affect infrastructure and surrounding human populations (Horwell & Baxter, 2006; Tomašek et al., 2016; Wilson et al., 2012). Volcanic Ash Transport and 61 62 Dispersion models (VATDMs) are used to forecast ash dispersal in the atmosphere and to 63 predict impacted areas (Bonadonna et al., 2011; Folch, 2012; Mastin et al., 2009). VATDMs are initialized with the source parameters of an eruption, which include the Mass Eruption Rate 64 65 (MER), plume heights, eruption duration, the Total Grain Size Distributions (TGSD) and the 66 vent location. In addition, VATDMs are initialized using scenarios from past eruptions (Scollo 67 et al., 2009, 2010, 2013) or a calculation of mass load estimates from post-eruption deposit analyses (Bonadonna et al., 1998; Bonadonna & Houghton, 2005; Bonadonna & Costa, 2012, 68 69 2013; Andronico et al., 2014; and references therein). Although some models consider ash

density variations and aggregation processes (Costa *et al.*, 2010; Carazzo & Jellinek, 2012;
Folch *et al.*, 2016), most of them assume ash settles via continuous individual particle settling,
with progressively smaller particles settling out with increasing distance from the eruptive
source (Pfeiffer *et al.*, 2005; Costa *et al.*, 2006; Folch, 2012; Bonadonna *et al.*, 2015). In these
models, the TGSD, the ash concentration and, in turn, terminal settling velocity of ash particles
control the description of ash sedimentation.

76 Many studies have shown that ash sedimentation from ash clouds can vary in space and time. For example, intermittent sedimentation, via ash fingers and mammatus, has been 77 78 observed at Mount Saint Helens in 1980 (Hobbs et al., 1991; Durant et al., 2009; Carazzo & 79 Jellinek, 2012), Soufrière Hills volcano in 1997 (Bonadonna et al., 2002) and Eyjafjallajökull 80 in 2010 (Manzella et al., 2015). More generally, Carozzo & Jellinek (2012, Figure 8) show that 81 the dynamics giving rise to fingers probably governed ash sedimentation in nearly half the 82 deposits in geological record form which grain-size data is available. However, although analogue laboratory experiments and scaling theory constrain the conditions favoring the 83 84 formation and evolution of ash fingers (Carazzo & Jellinek, 2012; Manzella et al., 2015; Del 85 Bello et al., 2017; Scollo et al., 2017), a critical knowledge gap is that their dynamics have not 86 been characterized quantitatively in situ.

Accordingly, we use a novel 3 mm wavelength (95 GHz) Doppler radar together with an optical disdrometer to characterize volcanic cloud sedimentation dynamics during a series of eruptions at Stromboli in 2015, when the eruptive activity was mostly characterized by shortlived weak ash plumes (**Figure 1A**). Guided by time series radar data from this campaign, and qualitative observations of analog experiments, we propose that sedimentation through the production of ash-laden fingers evident in **Figure 1B** (Patrick *et al.*, 2007) at Stromboli, is a consequence of convective instabilities (*e.g.* Hoyal *et al.*, 1999; Bonadonna *et al.*, 2002; Bush

94 et al., 2003; Durant et al., 2009; Carazzo & Jellinek, 2012, 2013; Manzella et al., 2015; Scollo 95 et al., 2017) modified by wind effects including the production of wind counter-rotating rolls 96 (Turner, 1960). We will argue that resulting overturning and stirring of the ash-air mixture 97 enhances the production of Particle Boundary Layers (PBLs) and the formation of ash fingers. This class of time-dependent sedimentation also predicts deposits with spatially-varying 98 99 textural characteristics that are readily identifiable in the field (Branney, 1991; Carazzo & 100 Jellinek, 2012). Finally, our results suggest that finger formation is not only a common feature 101 of sedimentation from transient and weak volcanic plumes, consistent with expectations from 102 Carazzo & Jellinek (2012, figure 8), but also that their formation depends strongly on the 103 character of wind-driven stirring.



104

Figure 1: A) Photograph of an ash plume emitted by the Southwest crater of Stromboli. B) Strombolian
ash plume drifted toward the North by the wind and forming an ash finger. The wind direction is shown
by the red arrow. C) Instrument setup at Stromboli Island. Upper right panel: 95 GHz radar at Punta
Labronzo (100 m a.s.l.) 2.1 km in slant distance from the vents (red line: radar beam direction). Lower
right panel: Disdrometer near Pizzo Sopra La Fossa at 900 m a.s.l., 350-450 m ESE of the summit
craters (SW crater, red triangle; NE Crater, blue triangle).

Our paper is organized in the following way: in Section 2, we present the methodology we applied at Stromboli as a case study, between the 28<sup>th</sup> of September 2015 and the 4<sup>th</sup> of October 2015, for identifying and characterizing distinct sedimentation features of volcanic plumes. In section 3, we present the combined radar and disdrometer results and highlight principal observations. In section 4, we use observations and a small suite of analog experiments to build a conceptual model for intermittent sedimentation from weak volcanic plumes constrained by our principal observations before concluding in section 5.

118 2 Methodology

#### 119 2.1 The measurement campaign at Stromboli

120 The goal of our measurement campaign is to characterize quantitatively the 121 sedimentation behavior of volcanic plumes with three techniques: (i) a 3 mm wavelength 122 Doppler radar (95 GHz), designed to track ash in volcanic plumes (Figure 1C); (ii) an optical disdrometer Parsivel<sup>2</sup> (Figure 1C); and (iii) analyses of ground deposit samples acquired during 123 124 the campaign (Freret-Lorgeril et al., 2019). In this study, it is important to note that each method 125 detects a local Particle Size Distribution (PSD). Such size distributions represent only a part of 126 the whole erupted mixture defined as the TGSD. From the disdrometer PSDs and phase delays 127 in radar return, we determine the modal settling velocity of ash particles.

During 39 hours of monitoring, 181 short-lived ash plumes (*i.e.*, from type 2a and 2b explosions; Patrick *et al.*, 2007) were recorded by the radar and 26 sedimentation events by the optical disdrometer. The geometry of the instrument locations (**Figure 1C**), which was the same during the whole campaign, did not allow for simultaneous measurement of the same ash sedimentation event as the wind was either blowing the sedimentation towards the radar ortowards the disdrometer.

134 2.2 The optical disdrometer

The optical disdrometer Parsivel<sup>2</sup> designed by OTT HydroMet (Figure 1C) can measure 135 136 particle settling velocities and diameter directly by converting the signal from a 780 nm 137 wavelength laser beam, emitted from a transmitter to a receiver, into a voltage signal (Figure 138 2A and 2B). The laser obscuration time is used to determine the settling velocity of falling particles, while the absorption amplitude, portion of the 54 cm<sup>2</sup> laser sheet surface obscured by 139 140 the particle, measures the particle size (Löffler-Mang & Jürg, 2000; Figure 2B). Measurable settling velocity and size ranges are between 0.05-20.8 m/s and 0.250-26 mm. Disdrometers 141 142 were originally designed for detecting hydrometeor precipitation. Nevertheless, Freret-Lorgeril 143 et al. (2019) and Kozono et al. (2019) have highlighted the reliability of such instruments for 144 detecting volcanic ash fallout with a high temporal resolution. Usually, disdrometer 145 measurements are time-averaged either to avoid stochastic variations of measurements due to 146 complex hydrometeor shapes (Löffer-Mang & Jürg, 2000) or to obtain stable quantitative data, 147 e.g. rain-rates, for long-lasting meteorological events as opposed to short ash fallout events. 148 Accordingly, we used the disdrometer maximum sampling frequency of 0.1 Hz (Tokay et al., 149 2014) in order to capture short time-varying sedimentation processes.



150

151 Figure 2: A) Sketch of the disdrometer detection procedure. A transmitter emits a laser sheet, i.e. in 152 red, which is measured by a receiver. B) Signal of the laser as a function of time. The amplitude and 153 time of the particle-induced laser obscuration are used to determine the size and the settling velocity of 154 ash particles, respectively.

Because of its lower detection limit and for accessibility reasons, the Parsivel<sup>2</sup> was set 155 156 up next to Pizzo Sopra La Fossa at ~900 m a.s.l, 370 m NE of the SW crater (Figure 1C). This 157 location allowed recording of the proximal coarse ash sedimentation. In addition, ground 158 tarpaulins were placed next to the disdrometer to collect ash samples for each event recorded 159 by the disdrometer. Freret-Lorgeril et al. (2019) demonstrated the influence of ash particle 160 shape on disdrometer terminal settling velocity and size measurements. Since the disdrometer 161 is designed to measure the shape of spherical droplets for meteorological applications, the shape 162 of settling ash particles were constrained and determined with a morpho-grainsizer Morphologi G3<sup>®</sup> by Freret-Lorgeril *et al.* (2019). Assuming particles fall parallel to their shortest axis, the 163 disdrometer detects particle with circle-equivalent diameter  $D_{CE} \ge 0.230$  mm (Freret-Lorgeril 164 165 et al., 2019). The particle detection resolution (*i.e.* the disdrometer size classes) is equal to 0.115 mm in  $D_{CE}$ . 166

167 We calculate particle numbers over the range of measured particle  $D_{CE}$  and ash 168 concentrations  $C_{ash}$  (kg/m<sup>3</sup>) with (Freret-Lorgeril *et al.*, 2019):

169 
$$C_{ash} = \frac{\pi \rho_{ash}}{6} \int_{D_{min}}^{D_{max}} N(D_{CE}) D_{CE}^{-3} dD_{CE} , \qquad (1)$$

170 where  $N(D_{CE})$  is the particle number density (m<sup>-3</sup>/mm) of a size class crossing the beam,  $dD_{CE}$ 171 (mm) is the diameter class width, and  $\rho_{ash} = 2739 \pm 50$  kg/m<sup>3</sup> the average ash particle density 172 measured by water pycnometry (Freret-Lorgeril *et al.*, 2019).

#### 173 2.3 Radar observations of cloud structure and sedimentation

174 The 3 mm wavelength Doppler radar is a compact version of the prototype of Bistatic 175 Radar System for Atmospheric Studies (BASTA, Delanoë et al., 2016), using smaller antennas 176 (30 cm wide). It was located at Punta Labronzo (Figure 2A), about 2 km NNE from Stromboli 177 summit craters. We chose this location to capture plume emissions and their subsequent 178 sedimentation that was preferentially dispersed towards the North due to ambient wind. The 179 radar beam (0.8° wide) aimed at an area 100 m above the SW crater. Particle radial (along-180 beam) velocity components  $v_r$  and reflectivity factors Z are measured inside each probed volume 181 of the beam as a function of time and range with resolutions of 2 s and 12.5 m (*i.e 12.5 m mode*), 182 respectively (See Table S1 in Supplemental information).

183 Radial velocities  $v_r < 0$  characterize particles with a motion component toward the radar, 184 as expected for particles falling through the upward inclined beam (Sauvageot, 1992). The 185 maximum non-ambiguous (*i.e* measurable by the radar without aliasing) velocity is  $\pm 9.87$  m/s (resolution of 0.01 m/s) in the 12.5 m mode used in this study. When detecting ash 186 187 sedimentation, it is possible to estimate the settling velocity of detected particles  $V_T$  from their 188 radial velocities. Under the assumption of no vertical component to the ambient wind,  $V_T$  can 189 be estimated from the beam elevation angle ( $\theta = 21.4^{\circ}$ ) and the range-time slope  $v_p$  of a 190 sedimentation pattern crossing the radar beam. However, horizontal winds, which were present 191 during the campaign, induce a velocity component along the radar beam axis and need to be 192 deconvolved from the radar measurements. We were unable to obtain reliable wind data during 193 the measurement campaign. Consequently, we exploit the radar beam geometry to constrain the 194 average horizontal wind velocity, and in turn, a more accurate estimate of  $V_T$ .



## Direction towards the radar

**Figure 3**: Schematic representation of measured and inferred velocity components of ash and wind, given a prescribed radar beam geometry.  $v_w$  is the horizontal wind velocity presumably inducing the along-beam (radial) displacement speed of the whole sedimentation pattern,  $v_p$ , measured from its range-time slope in reflectivity time series.  $v_r$  are the measured internal radial velocities of the recorded pattern. The measured differential velocity  $v_{r,a} = |v_r| - |v_p|$ , is assumed to be the radial velocity component related only to ash particles. The settling velocity  $V_T$  of particles in a sedimentation pattern is then calculated using **Equation 3**.

195

Figure 3 shows each velocity component related to a sedimentation feature detected inside the radar beam. Under trigonometric considerations, the unknown horizontal wind velocity  $v_w$ , giving rise to the values of  $v_p$ , can be estimated with  $|v_p|/\cos\theta$ . Additionally, we can calculate the radial component velocity of the sedimentation,  $v_{r,a}$ , in the radar beam as:

207 
$$v_{r,a} = \left| v_r \right| - \left| v_p \right|, \tag{2}$$

Finally, from the vertical component of  $v_{r,a}$  the mean settling velocity of all the ash particles  $V_T$  is:

210 
$$V_T = \frac{|v_r| - |v_p|}{\sin\theta} .$$
(3)

Assuming that radar returns are in the Rayleigh scattering limit, the radar reflectivity factor, *Z*, characterizes the quantity of targets crossing the radar-beam and can also be calculated for the local PSD measured by the disdrometer, as (Sauvageot, 1992):

214 
$$Z = \int_{D_{\min}}^{D_{\max}} N(D) D^6 dD, \qquad (4)$$

where Z is in mm<sup>6</sup>/m<sup>3</sup>, N(D) is the particle number density (mm<sup>-1</sup>/m<sup>3</sup>), D is the particle size in mm and dD is the range of particle size classes used (*e.g.* 0.115 mm resolution from disdrometer size classes). Z is expressed in logarithmic units (dBZ) as:

218 
$$Z(dBZ) = 10 \times \log_{10}(Z(mm^6/m^3)).$$
 (5)

219 Strictly, **Equations 4** are valid if we assume that the ash-air mixture is a dilute and 220 homogeneous suspension of spherical particles with diameters less than about  $\lambda/4$  (Gouhier & 221 Donnadieu, 2008).

#### 222 **3 Results**

#### 223 3.1 Intermittent sedimentation from disdrometer records

We observed 24 sedimentation events in the optical disdrometer data associated with single ash plume emissions blown toward the instrument by an easterly wind. **Figure 4** shows a time series of five sedimentation events recorded by the disdrometer on October 3, 2015 between 12:35:02 and 14:50:02 UTC. These events, lasting between 110 and 340 s, are distinguished by time breaks during which no particles were detected. The multi-peaked variations in sedimentation rate, *i.e.* the mass of particle falling as a function of time and per unit area, shown for the five events in **Figure 4** are characteristic of 96% of the 24 events recorded by the disdrometer.



Figure 4: Time series of the number of falling particles detected by the optical disdrometer on October
3, 2015 between 12:35:02-12:50:02 UTC and 14:25:02-14:50:02 UTC. The arrows show the first time
that the radar detects the ash plumes rising above the vent, which occurs a few seconds after the start
of each explosion. The ash sedimentation events are recorded several minutes afterwards.

232

Figure 5 shows another 8 min-long sedimentation event comprised of 2,514 detected ash particles on October 3, 2015 between 09:49:00 and 09:59:00 UTC, originating from an ash plume emitted 7 minutes prior at 09:41:52 UTC. The PSD (Figure 5A) ranges from 0.230 <

 $D_{CE} < 0.690$  mm (Class 1-4) and has a modal value between 0.345 and 0.460 mm (Class 2). 240 241 Settling velocities (Figure 5B) range between  $0.7 < V_T < 3.2$  m/s with a modal value between 242 1.4 to 2 m/s. These are consistent with maximum settling velocities (cf., Ganser, 1993) of 1.8  $< V_T < 2.4$  m s<sup>-1</sup> for Class 2 particles using a measured sphericity of 0.75 and average sizes 243 244 between the longest and the intermediate axes of the particles (Freret-Lorgeril et al., 2019). Figure 5B shows the widespread distribution of settling velocities as a function of particle size 245 246 that is characteristic of all detected sedimentation events. In Figure 5C, the number of ash 247 particles falling through the laser beam over time, and hence the sedimentation rate, shows 248 significant fluctuations affecting all size classes. Six pulses of up to 145 particles occur with 249 variations of 98% of the maximum pulse amplitude. The sedimentation can be divided into two 250 main phases of 4 min 40 s and 3 min 10 s, respectively, that are separated by a 20 s interval with very few detected particles. Successive pulses within each sedimentation phase occur at 251 intervals of 20 to 60 s and in between pulses, the sedimentation rate decreases for less than 10 252 253 s to one minute before increasing again.

254 Figure 5C shows expected radar reflectivity factor values derived from the 255 sedimentation rate measured by the disdrometer using Equations 4-5. Z varies from -26.29 <256 Z < +12.85 dBZ for this event and, using **Equation 1**, we estimate average and maximum ash concentrations of  $0.64 \times 10^{-4}$  and  $1.88 \times 10^{-4}$  kg/m<sup>3</sup>, respectively. For all sedimentation events 257 258 observed in this study, the disdrometer data shows relatively constant PSD modes of  $0.403 \pm$ 259 0.058 mm comprising on average 46% of the total number of detected particles, ranging from 260 45 to 1,261. The ash concentrations inferred from Z values on the whole set of disdrometer data range between  $7.61 \times 10^{-7}$  to  $6.81 \times 10^{-4}$  kg/m<sup>3</sup>. 261



Figure 5: A) Disdrometer-derived local Particle Size Distribution of the ash sedimentation event between 09:49:02 and 09:59:02 UTC on Oct. 3 2015. Each size class ( $\pm$  0.05 mm) is color-coded. **B**) Settling velocities ( $V_T$ ) measured by the disdrometer as a function of circle-equivalent diameters ( $D_{CE}$ ). Bin colors represent number of particles. **C**) Temporal series of particle number of each class (colored lines), the total number of particles (dashed crossed line) and calculated reflectivity factor Z (red triangles). The grey areas and the black arrows represent the two sedimentation phases and the 6 sedimentation pulses, respectively.

271 Between October 2 and 3, 2015, weak northerly winds advected transient ash plumes 272 towards the Doppler radar location at Punta Labronzo. The 3 mm wavelength Doppler radar 273 detected the emission of ash plumes ~100 m above the vent and 15 subsequent ash 274 sedimentation events over a ~2100 m slant distance. Figure 6A shows the range-time profiles 275 of ash reflectivity factor, Z, with white arrows marking detected ash plumes rising above the 276 vent at ~2000 m range. The first two detected ash plumes are followed by the detection of ash 277 fallout, appearing as Long Duration Reflectivity Features (LDRFs) that grow in size as they 278 move towards the radar site (white line in **Figure 6A**). The width of the first LDRF, detected at 279 15:23:00 UTC, is ~100 m until 15:29:00 UTC. During the following 3 min, it steadily widens 280 to ~600 m. The second LDRF, detected at 15:47:00 UTC, steadily widens over the following 4 281 min from <100 m to ~200 m. For the remainder of the detection, its width thins and widens 282 multiple times. The LDRFs do not appear to be homogeneous in Z along the radar beam axis 283 and show small-scale range-time trends spaced quasi-periodically within their large-scale 284 pattern (black arrows in **Figure 6A**). The slopes of the *small-scale reflectivity trends* are lower 285 than the slope of the large-scale pattern.

The overall Doppler velocities,  $v_r$ , within the first LDRF show a decrease from ~0 m/s to ~-2 m/s around 15:26:00 UTC (**Figure 6B**). At about 15:28:00 UTC, the velocities vary spatially and intermittently above and below 0 m/s within the LDRF at later time and closer range. The second LDRF shows a similar evolution in  $v_r$  over time and range of detection.



Figure 6: A) Ash reflectivity factor (Z) time series recorded on October 2 2015 (15:15-16:10 UTC) as
a function of time (x-axis) and along-beam slant distance (y-axis: range). Five ash plumes are pointed
by white arrows at crater distance (2100 m) while two Long Duration Reflectivity Features (LDRFs;
indicated by white lines), lasting ~20 min each, are tracked over 2 km toward the radar location. Black
arrows indicate example of small-scale range-time trends B) Range-time profiles of radial velocities of
the same eruptions.

297 In **Figure 7**, we show the sedimentation occurring within 1000 m of the SW Crater. The 298 parent ash plume, which drifts to the north, is observed at slant distances of 2069-2119 m. As 299 in Figure 6A, the range-time Z measurements show the ash plume rise, followed by a 300 sedimentation event split into two phases. The first 20 s-long impulsive phase shows the ash 301 plume detected 100 m above the vent at 14:26:44 UTC. Initially, it is < 50 m wide and then 302 widens to 150 m in the N-S direction 16 s after detection (Figure 7A). 12 s after the detection 303 onset, a maximum of 16.52 dBZ is reached in the bin at 2081 m. The following 158 s-long 304 decrease in reflectivity until 14:29:42 UTC marks the ash plume exiting the radar beam. Radial 305 velocities  $(v_r)$  in the emitted plume alternate in time between negative and positive values with 306 a period of 24 s between the first and second yellow bands (see white arrows at top left in 307 Figure 7B) and 40 s between the second and third yellow bands. The variations of the ash 308 plume's internal velocities and reflectivity do not appear correlated.

After the initial ash plume exits the radar beam in **Figure 7**, there is a period of 123 s without ash detection. We detect ash again between 14:31:45 and 14:40:01 UTC at a lower altitude in downwind bins closer to the radar (994-1944 m) with a maximum reflectivity of  $Z_{max}$ = -4 dBZ, which is weaker than that of the initial ash plume. The width of this LDRF increases from 100 to 300 m from 14:33:30 to 14:34:45 UTC and its average slope indicates an average speed of -2.3 m/s towards the radar. Similar LDRFs, between 15:22:25-15:43:46 UTC and 15:47:02-16:08:09 UTC in **Figure 6A**, are detected within 2000 m of the radar location.

In **Figure 7**, three continuous features, with -14.23 < Z < -4.72 dBZ, emerge within the LDRF 160 s after its initial detection. These features are 25-100 m thick, 50 m on average and have range-time slopes that are distinct from the average slope of the overall LDRF signal. At 14:35:07 UTC, we observe three local maxima in *Z* values located in the middle of these features at different distances that have a relatively constant spacing of about 85 m (80 to 95 m) and intervals of 30 to 50 s (**Figures 7A** and **7C**). These small-scale reflectivity trends are best described by a constant along-beam slope, indicating a constant velocity component towards the radar. Thus, the reflectivity of the emitted ash plumes is characterized by an initial maximum decaying continuously, whereas the LDRFs detected closer to the radar are characterized by the emergence of multiple small-scale and evenly spaced reflectivity trends (**Figure 6A** and **Figure 7A**).



328 Figure 7: A) Range-time profile of measured radar reflectivity factor, Z, of a single ash plume emission 329 and subsequent LDRF. Red arrows indicate the small-scale reflectivity trends inside the LDRF. Inset: 330 sketch of the wind direction according to the visually observed trajectory of the ash plume relative to 331 the radar beam (at 2100 m and about 14:26 UTC on Oct. 2 2015). B) Range-time profiles of radial 332 velocities of the same eruption. White arrows indicate the pattern of positive Doppler velocities in the 333 emission. Inset: Histogram of Doppler (along-beam) velocities (filtered with threshold Z > -13 dBZ) 334 inside the whole LDRF pattern. Doppler velocities mainly range between -5.5 to 1.5 m/s with a mode 335 between -3 and -2.5 m/s. C) Z-profile across the three main small-scale range-time trends (peaks marked 336 by crosses) at 1506 m from the radar between 14:32:30 and 14:38:30 UTC (black dashed line in A). D)

Log-Log plot of the normalized reflectivity ratio  $(Z^*/Z_0^*)$  as a function of the normalized time scale for wind-driven overturning of rolls observed within the ash cloud  $t/(H_0/v_w)$  (see text). Here,  $H_0$  is the width of the parent plumes where the maximum reflectivity is detected in Figure 6 and 7A, and  $v_w$  is the horizontal wind velocity. The three events recorded by the radar on 2 October at 14:26 UTC (black dots), 15:17 UTC (green dots) and 15:44 UTC (red dots) show a decrease in normalized reflectivity factor.

343 In total, 93% of the Doppler velocities within the LDRF in Figure 7 are marked by negative values within  $-1 < v_r < -5$  m/s (inset in **Figure 7B**). They show an abrupt increase in 344 345 overall  $v_r$  from  $\leq -3$  m/s to -0.7 m/s around 14:34:00 UTC and a return to  $\leq -3$  m/s closer to the radar around 14:36:30. The horizontal component of the LDRF's range-time slope in Figure 346 347 **7A** corresponds to a horizontal wind velocity,  $v_w = 2.47$  m/s. From the LDRF's internal modal  $v_r$ , -2.5 <  $v_r \le -3$  m/s (inset in Figure 7B), a modal radial velocity  $v_{r,a}$  between -0.2 and -0.7 m/s 348 349 is found. The corresponding  $V_T$  ranges from 0.55 to 1.92 m/s (Equation 3), indicating detected 350 particle sizes between 0.1 and 0.3 mm (Ganser, 1993; Freret-Lorgeril et al., 2019).

351 A final observation was done on October 2, from 12:19 to 12:25 UTC, when LDRFs emerging in the radar data were accompanied by simultaneous visual observation of the 352 353 formation of ash fingers under a dilute weak plume (Figure 8; Supplemental information S2 354 and video S2). The same features observed in Figures 6 and 7 are highlighted in Figure 8. 355 Small-scale reflectivity trends are detected within a LDRF over a distance of 550 m (i.e. from 356 1256 m to 706 m along the radar beam) and during 284 s (Figure 8). The small-scale features 357 show reflectivity ranging from -16 to -7 dBZ (Figure 8A). Finally, the LDRF is characterized 358 by negative Doppler velocities alternating between -4 and -1 m/s (Figure 8B).



Figure 8: Simultaneous observation of ash fingering (black arrows) at Stromboli from visible
(snapshots) and radar imagery on October 2, 2015 between 12:19 and 12:25 UTC. A) Range-time
profile of ash reflectivity factor Z. The white dashed line indicates the detected LDRF. B) Range-time
profile of radial velocities v<sub>r</sub>.

#### 364 4 Discussion

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The LDRFs detected by the radar over a range of 150 to 2000 m downwind from the vent (**Figures 6**, **7** and **8**) are interpreted as ash sedimentation from the October 2015 eruption 367 clouds. We interpret the temporal correlations between pulses in the sedimentation rate 368 recorded proximally on the ground by the disdrometer (Figures 4 and 5) and the small-scale 369 radar reflectivity trends (Figures 6A and 7A) as evidence that the ash sedimentation is pulsed 370 and quasi-periodic with recurrence intervals of 20-60 s in the disdrometer data and 30-50 s in 371 the radar reflectivity data. On the basis of this consistency, the observations of finger-like 372 structures in Figure 1B, Figure 8 and the observations of ash curtains below Strombolian 373 plumes in Patrick *et al.* (2007), we hypothesize that the sedimentation pulses are ash fingers 374 arising from the same mechanism, which we now explore in detail.

# 4.1 A hypothesis for the origin, geometry and evolution of the ash fingers: Wind-driven rolls, convective fingering and sediment thermals

377 Before comparing the PSD measured by the disdrometer to that detected in the 378 sedimentation targeted by the radar, several challenges must be addressed. First, the local PSD 379 measured by the disdrometer only represents the fraction of the PSD falling at the location of 380 the disdrometer (Figure 2). The largest pyroclastic particles are likely to fall out of the ash 381 cloud before reaching the disdrometer. The disdrometer has also a lower detection limit of 0.230 382 mm and thus records a local PSD that is not only biased against the largest particles but also 383 biased against fine ash, the dynamics of which probably control finger formation (Carazzo & 384 Jellinek, 2012). Nevertheless, Freret-Lorgeril et al. (2019) have shown that the disdrometer 385 local PSDs matched 75-94% (i.e., in vol%) of the ground PSDs sampled at the same location. 386 The total concentration measured by the disdrometer consequently sets a lower limit for the 387 total ash concentration settling at its location. Additionally, transient non-dilute mixture effects 388 such as enhanced or hindered settling (Del Bello *et al.*, 2017), potentially complicate  $V_T$  derived 389 from  $v_r$ . Bearing this in mind, we proceed with caution, to investigate the origin of the 390 sedimentation features.

391 Processes that drive the formation of fingers must explain four principal observations392 from the combined disdrometer and radar data:

- 393 1. Quasi-periodic sedimentation pulses with wavelengths of 80-95 m and periods of 30-50 s
  394 (Figure 7A).
- 395 2. An increase in size over time of the initial sedimentation pulses detected (Figures 6A and
  396 7A)
- 397 3. The 120 360 s time scale for fingers to emerge (Figure 6A and 7A) after initial detection
  398 of ash sedimentation.
- 399 4. The overall increase of v<sub>r</sub> within the sedimentation after 14:33:30 and 12:15:00 in Figure 400
  7B and Figure 8B, respectively (after 15:28:00 for the first LDRF in Figure 6B).

401 Explosive volcanic eruptions, no matter their size and nature, are commonly 402 characterized by unsteady mass fluxes at the vent (Ripepe et al., 1993; Donnadieu, 2012; 403 Scharff et al., 2015; Chevalier & Donnadieu, 2015). If the source mass flux fluctuates 404 significantly over time, we would expect to see such fluctuations in the radar data acquired 405 above the vent. However, the reflectivity of the ascending ash plume detected by the radar 406 (Figure 7A) does not show such time-dependent behavior during the emission. Nonetheless, 407 the radar data do show an alternating pattern of positive and negative Doppler velocities (see 408 white arrows top left of this Figure 7B), which we interpret as internal motions within the 409 plume.

410 With an approximately constant mass eruption rate (at least over the time scale for 411 sediment fingers to form in our data), one mechanism for the quasi-periodic formation of the 412 fingers evident in Figure 1B and Figure 8 is the intermittent growth and detachment of ash-413 rich particle boundary layers (Carazzo & Jellinek, 2012; Scollo et al., 2017). If sufficiently 414 strong, the wind will influence the timing of these convective effects in the Stromboli ash cloud 415 in Figure 1 and Figure 8. If we assume that the wind is horizontal, from the crater to the radar 416 site, consistent with qualitative observations made in the field, we estimate the wind speed as 417 the horizontal component of the overall range-time slope for the sedimentation in radar 418 reflectivity data as  $v_w = 2.47$  m/s (Figure 7A). This background wind is larger than the 419 maximum inferred settling speeds for particles forming the median grain size in the PSD 420 (Figure 5C) and will consequently affect their dynamics at leading order. For a cloud width  $H_0$ 421 (*i.e* where the maximum reflectivity is detected), wind-driven organization of the ash plume 422 will occur over a time scale  $t_w = H_0/v_w$ . Because this time scale is comparable to or smaller than 423 the time scale for particle settling  $t_p = H_0/V_T$  for much of our inferred PSD, an initial prediction 424 for particle buoyancy-driven sedimentation is that it will begin approximately at or shortly after 425  $\sim 1 t_w$ , which is consistent with **Figure 7D**. We consequently infer that there is strong particle-426 fluid coupling and these mixture buoyancy effects are more important than particle loss by their 427 ejection from the rolls (i.e., so-called Stokes number effects). Moreover, if this inference is 428 incorrect, the change in concentration in Figure 7D would begin at a different part of the 429 evolution than where  $t/(H_0/v_w) = 1$ .

Although consistent with **Figure 7D**, this heuristic expectation provides little insight into the underlying mechanisms governing how particles become concentrated into mixtures sufficiently dense to form fingers (Carazzo & Jellinek, 2012). **Figure 7D** shows only that progressive particle loss occurs after ~1  $t_w$ . Nonetheless, observations of fingers in the images of the wind-advected plume taken on October 2, 2015 (**Figure 1B**) indicate that the small-scale reflectivity trends in **Figure 6A** and **Figure 7A** represent ash fingers. If ash sedimentation is 436 only controlled by individual particle settling, we would expect to see an LDRF with no small-437 scale reflectivity features within.

438 The maximum finger length,  $h_f$  below the parent ash cloud is  $h_f = t_w V_f$ , where  $V_f$  is the 439 finger fall velocity (see Equation 23 in Carazzo & Jellinek, 2012). If we assume that particle-440 fluid mixtures forming fingers descend faster than individual particles (*i.e.*, a condition for 441 fingers to form in the first place) then we can take  $V_f \approx V_T$  detected in the LDRFs in **Figure 7B**. 442 With  $H_0 = 100$  m we find  $189 < h_f < 300$  m, in agreement with estimates from visual 443 observations of fingers in Figure 1B and Figure 8. Additionally, the initial increase in width 444 followed by a decrease in overall  $v_r$  of the LDRFs in Figure 6A, Figure 7A and Figure 8B 445 indicate that the volume of the fingers increases as they fall, which is followed by a decrease in 446 settling velocity soon after, consistent with the dynamics of sediment thermals (Bush et al., 447 2003). On this basis, we propose that these are observations of fingers evolving into sediment 448 thermals (Ernst et al., 1994; Bush et al., 2003; Carazzo & Jellinek, 2012; Suzuki & Koyaguchi, 449 2013; Del Bello et al, 2017; Scollo et al., 2017

#### 450 4.2 A conceptual model for the generation of fingers forming sediment thermals

451 To build qualitative understanding of the sedimentation dynamics below wind-affected 452 volcanic plumes, we conduct simple experiments (described in the Supplemental Information 453 S3) in which we introduce dilute warm mixtures of 0.030, 0.190, 0.225 and 2 mm diameter 454 silica particles into water with an imposed constant crossflow velocity to form an advected 455 analog for the cloud shown in Figure 1. We choose ad hoc properties in the experiments to 456 capture the range in  $t_w/t_p$  inferred for our PSD for a wind speed of 2.47 m/s, and to explore 457 sensitivity of the cloud structure to this ratio of time scales (Figure 9). In Figure 9A, counter-458 rotating rolls induced by the experimental crossflow develop along with regularly spaced 459 sedimentation features in the form of particle-laden fingers. They detach to form sediment 460 thermals but retain the regular spacing of the features. Descending sediment thermals entrain 461 ambient water and their buoyancy increases. This entrainment causes them to slow such that 462 entrained particles sediment from their basal edges to form a second generation of smaller 463 sediment thermals. The second generation of sediment thermals disperses and the dispersed 464 clouds of both generations continue to settle to the tank floor at the terminal settling velocity of 465 the modal particle size (0.03 m/s and 0.200 mm respectively) (Bush et. al, 2003; Carazzo & 466 Jellinek, 2012).

467 Guided by the experimental observations from Figure 9A, previous studies (Ernst et al., 468 1994; Bush et al., 2003; Carazzo & Jellinek, 2012; Scollo et al., 2017), and the field 469 observations presented in this study we present the conceptual model. We key the main features of this model to 4 time slices ( $t_1$ - $t_4$  in Figure 9B), which indicate the sedimentation pulses in 470 471 Figure 7A and those inferred in Figure 9A. The sketches of Figure 9C depict the formation of 472 ash fingers over a time  $\sim t_w$  evolving, in turn, into sediment thermals at the base of the volcanic 473 ash cloud, as seen by the radar. The development of counter-rotating rolls in the particle cloud 474 enhances the delivery of particles to its base, particularly where the flow in the rolls converge 475 (Front view in **Figure 9C**). This relatively dense mixture undergoes gravitational instabilities to form fingers, in turn (Carazzo & Jellinek, 2012, 2013; Scollo et al., 2017) (t<sub>2</sub> in Figure 9C; 476 477 see also figures 3, 5, 6 and 9 in Scollo et al., 2017). Fingers descend, entrain ambient fluid and 478 decelerate (Figure 9A, t<sub>3</sub>). This basic evolution agrees with the increase in cloud width (point 479 2 in section 4.1) and overall decrease of cloud settling velocity (point 4 in section 4.1) occurring 480 simultaneously around 14:33:30. However, Figure 7B shows that the overall settling velocity 481 increases after 14:36:30, which could imply that a second generation of sediment thermals has 482 formed, similar to that observed at t<sub>4</sub> in the Figure 9A.



484 Figure 9: A) Photo of polydisperse particle cloud experiment exhibiting regularly spaced particle-laden 485 fingers transforming into sediment thermals. Crossflow is from left to right, nearside counter-rotating 486 roll indicated by number 1 with red arrows showing fluid motion and sediment thermals are marked by 487 numbers 2 and 3 with red arrows showing overturning head fluid motion. t<sub>4</sub> corresponds to instance in 488 A), which represents the second generation of sediment thermals. B) Zoomed view of sedimentation 489 pulses in Figure 7A with black bars representing four instances in time  $(t_1-t_4)$ . C) Model sketches of 490 parent ash emission with crosswind generating horseshoe eddies that induce counter-rotating rolls in 491 resulting ash cloud.  $t_1$  and  $t_2$  show sedimentation behavior at instances in time corresponding to  $t_1$  and 492  $t_2$  in A), with radar beam in red.  $t_1$  shows side and front views of ash cloud with arrows indicating flow 493 direction for counter-rotating rolls and initial finger instability. Darker particles at base of ash cloud 494 in Front View and t<sub>2</sub> highlight the Particle Boundary Layer formation enhanced by counter-rotating 495 rolls above.  $t_2$  shows overturning motions at head of finger as it evolves into a sediment thermal (Ernst

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496 et al., 1994; Bush et. al, 2003; Carazzo & Jellinek, 2012, 2013; Suzuki & Koyaguchi, 2013; Del Bello
497 et al., 2016; Scollo et al., 2017).

498 Predictions from the conceptual model in Figure 9 satisfy the list of observations we499 identify in section 4.1 and make additional predictions, in line with our observations:

- 500 1. Fingers transitioning into sediment thermals: Periodic structure of the sedimentation
  501 pulses in space and time (Figures 4, 5, 6, 7 and 9).
- 5022. Entrainment and finger/thermal dilution: The volumetric growth of sediment thermals503through entrainment is consistent with observed increases in size in Figures 6 and 7A,504and the decrease in reflectivity density of the initially detected sedimentation pulses in505Figure 7D. Dilution through entrainment is also consistent with the overall increase of506 $v_r$  within the sedimentation, indicating a decrease in settling velocity, occurring after50714:34:30 in Figure 7B and after 12:22:00 and 12:23:30 in Figure 8B, respectively.
- 5083.Multiple generations: The entrance of later forming fingers into the radar beam, after509the initial finger enters the radar beam, explains the emergence of periodic510sedimentation structure after the first sedimentation pulse is detected (Figure 7A). In511particular, the particle loading supplied by a second generation of sediment thermals512may explain the curious decrease of  $v_r$  (between 15:26:00 and 15:28:00 in Figure 6B;513after 14:36:30 in Figure 7B), which indicates an increase in settling velocity, later in514the sedimentation.

515 4. Sediment thermals, entrainment and individual particle settling: The dispersal of sediment 516 thermals leads to individual particle settling, which is captured by the radar and can 517 explain the modal  $v_r$  in the sedimentation corresponding to a smaller size range of 518 particles than that detected by the disdrometer. The faster fall velocity of sediment 519 thermal mixtures vs the terminal settling velocity of the particles that compose them can 520 explain the extreme values of  $v_r \le -3$  m/s detected in the sedimentation (Figure 7B). 521 5. Theoretical plausibility of sediment fingers: The modal value of the local PSD measured 522 at the disdrometer (0.345-0.460 mm) and the modal value derived from  $v_r$  (0.100-0.300 523 mm) are both consistent with theoretical constraints indicating a 0.060-0.250 mm particle 524 size limit for the formation of fingers for plausible ash concentrations (Carazzo & Jellinek, 525 2012, Figure 8). Moreover, the maximum ash concentration derived from the disdrometer measurements, 0.681 g/m<sup>3</sup>, is comparable to the minimum concentration required (1 g m<sup>-</sup> 526 527 <sup>3</sup>) for the development of gravitational instabilities determined from scaling arguments. 528 (Carazzo & Jellinek, 2012; Manzella et al., 2015). Furthermore, radar measurements of 529 ash cloud thickness, finger fall and ambient wind velocity predict a finger length,  $189 < h_f$ 530 < 300 m, which is consistent with the observations in **Figure 1B**.

#### 531 4.3 Wind-affected ash finger sedimentation

532 Ash fingering requires the formation of PBLs, which are expected to govern ash 533 sedimentation in highly explosive eruptions (Carazzo & Jellinek, 2012), as well as the relatively 534 weak events at Stromboli in 2015 (i.e., type 2 explosions of Patrick et al., 2007; section 2.1). It 535 is important to note that the ash clouds observed in this study are not related to large MERs or 536 high concentrations of particles, *i.e.* two mandatory conditions for the formation of gravitational 537 instabilities in addition to the presence of fine ash, after Scollo et al., 2017. Hence, our new 538 finding at Stromboli that will enter in consideration for how and when ash fingers form 539 generally is that wind-driven stirring can modulate the growth and timing of PBLs. In particular, 540 unsteady wind-driven motions can form spatially local zones of fine-grained ash concentration 541 where counter-rotating vortices meet (section 4.2). We expect the occurrence of these wind effects on sedimentation to be common features of any ash cloud for which  $t_w \sim t_p$  (section 4.1). 542

- 543 Furthermore, these effects will be enhanced in clouds with higher particle concentration and a
- 544 greater proclivity for PBL formation (Carazzo & Jellinek, 2012).



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Figure 10: Ash plume snapshots of A) the Eyjafjallajökull long-lasting plume on 4 May 2010 (figure modified from Manzella et al., 2015) and B) the 23 November 2013 paroxysmal tephra plume at Mount Etna (Figure modified from Scollo et al., 2017, source: <u>www.tboeckel.de/EFSF/efsf etna/Etna2013/</u> Etna 11 13/volcano etna 11 2013 e.htm). White arrows indicate ash finger location. Are reported on both figures the horizontal plume and wind speed ( $v_{plume}$  and  $v_w$ ), the plume width ( $H_0$ ), the distance from the vent where the first finger is observed (d) or is expected due to wind-driven organization (d<sub>0</sub>) and the maximum distance  $d_0^{max}$  assuming  $v_w = v_{plume}$ .

553 However, among the studies relating to ash finger formation (Carazzo & Jellinek, 2012, 554 2013; Manzella et al., 2016; Scollo et al., 2017; and reference therein), the potential for wind-555 driven dynamics of ash finger sedimentation and potential impact on ground in a natural system 556 have only been explored by Manzella et al. (2015) during the Eyjafjallajökull eruption in 2010. 557 This study provides detailed observations of a long-lasting plume emitted on 4 May 2010. The 558 plume horizontal velocity  $v_{plume}$  was ~7.9 m/s with a horizontal  $v_w$  at its base of 11 m/s (Figure 559 10A). Considering a plume width  $H_0$  of ~2100 m obtained from Figure 10A, we derive a wind-560 driven organization time (see section 4.1)  $t_w$  of 191 s. This value, multiplied by  $v_{plume}$ , leads to a distance  $d_0$  of ~1500 m being similar to the ~1400 m from which fingers are continuously 561

562 observed (Manzella et al., 2015). Therefore, our results for transient weak Strombolian plumes 563 help to reconcile key observations of sedimentation from the relatively continuous 564 Eivafjallajökull event: Fingers appear approximatively or shortly after ~1 wind-driven organization time  $t_w$ . Also, if we take  $v_{plume} = v_w$ , we obtain a maximum distance at which the 565 first finger should form with  $d_0^{max} = 2100$  m, approximately the width of the plume. Hence, 566 evidence of ash fingers on ground deposit should emerge at a distance greater than  $d_0^{max}$ . 567 568 depending on the horizontal speeds of the fingers and their settling velocity (Manzella et al., 569 2015).

570 We have applied the same reasoning for a weak plume emitted during the 23 November 571 2013 paroxysm at Mount Etna (Figure 10B). For this eruption, less information (e.g. distance 572 of finger formation and plume horizontal speed) is available except for a wind speed of ~40 m/s 573 and a plume width of ~3000 m (Corradini et al., 2016; Montopoli, 2016). Considering the 574 maximum plume velocity to be equal to  $v_w$  leads to  $d_0^{max}$ =3000 m. Although less precise than 575 the previous calculation for Ejyafjallajökull, this simple estimate seems reasonable at Etna 576 regarding the visually observed location of the first fingers inside La Valle del Bove from 577 Figure 10B.

Predictions based on our observations and conceptual model constitute a step forward in understanding the location, timing and overall character of ash sedimentation driven by windaffected PBL convective dynamics. Nevertheless, quantitative observations of ash finger formation remain poorly constrained for most eruptive events. Future field observations should include high space-time resolution instruments similar to our Doppler radar to more precisely constrain ash finger dynamics from transient and weak volcanic plumes *in situ*. Finally, given the major control of wind on the 3-dimensional structure and dynamics of PBLs, future experiments should address finger formation in the presence of background wind velocity fieldswith varying strength and direction.

#### 587 5 Conclusions

Quantifying the processes leading to ash finger formation *in situ* is a key challenge to understand their impacts on ground deposits and how fine ash is dispersed in the atmosphere. Our joint radar-disdrometer study aims to characterize, quantitatively, the behavior of heterogeneous ash sedimentation from volcanic plumes at high spatial and temporal resolution. As a case study, the campaign carried out at Stromboli in 2015 demonstrates an effective approach for monitoring volcanic ash plumes and for characterizing the internal dynamics of their sedimentation.

595 Our observations of ash sedimentation show that ash sedimentation is time dependent, 596 potentially periodic, and well explained as a consequence of wind-affected gravitational 597 instabilities to ash-air mixtures in the form of ash fingers. Our results support a hypothesis that 598 wind-induced counter-rotating rolls may concentrate ash particles such that the formation and 599 release of relatively dense ash-air mixtures occurs over a time scale that is small in comparison 500 to individual particle settling.

In more detail, we have provided evidence that the maximum particle size and minimum concentration favoring fingering dynamics at the base of volcanic plumes (Carazzo & Jellinek, 2012, Scollo *et al.*, 2017) may be influenced by ambient wind. Estimates of the *in-situ* ash size distribution and concentration across the fingers are, however, required for a more thorough comparison of these data with existing theoretical constraints on the basis of laboratory experiments. Time-dependent wind-induced counter-rotating rolls in ash plumes are not a new phenomenon and have been observed for a wide spectrum of eruptive activity (Ernst *et al.*, 608 1994; Suzuki & Koyaguchi, 2013). However, their role in ash finger formation was not 609 previously quantified. Accordingly, we show that ash fingers may form approximatively at or 610 shortly after one wind-driven organization time. Finally, if counter-rotating rolls, fingering and 611 sediment thermal dynamics emerge even for very dilute and short-lived ash plumes observed 612 in this study, we expect these physical processes to be common features of wind-affected 613 explosive volcanic plumes with higher fine ash concentrations that have a greater proclivity for 614 particle boundary layer-related dynamics to occur (Carazzo & Jellinek, 2012). Our newly 615 proposed model for the effect of wind on ash finger formation can be tested with future 616 experimental studies and may need to be parameterized for ash cloud dispersal and fallout 617 modelling in the framework of ash hazard assessment following explosive eruptions.

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626 **REFERENCES CITED** 

Andronico, D., Scollo, S., Cristaldi, A., Lo Castro, M. D., 2014. Representivity of incompletely
sampled fall deposits in estimating eruption source parameters: a test using the 12-13
January 2011 lava fountain deposit from Mt. Etna volcano, Italy. Bull. Volcanol.
76(10):861, doi:10.1007/s00445-014-0861-3.

- Bonadonna, C., Houghton, B.F., 2005. Total grain-size distribution and volume of tephra-fall
  deposits. Bull. Volcanol. 67, 441–456, doi:10.1007/s00445-004-0386-2.
- 633 Bonadonna, C., Ernst, G. G. J., Sparks, R. S. J., 1998. Thickness variations and volume
- 634 estimates of tephra fall deposits: the importance of particle Reynolds number. J. Volcanol.
- 635 Geotherm. Res. 81, 173–187.
- Bonadonna, C., Macedonio, G., Sparks, R.S.J., 2002. Numerical modeling of tephra fallout
  associated with dome collapses and Vulcanian explosions: Application to hazard
  assessment in Montserrat, in: Druitt, T.H., and Kokelaar, B.P., (Eds.), The eruption of
  Soufrière Hills Volcano, Montserrat, from 1995 to 1999: Geological Society of London
  Memoir 21, pp. 517–537, doi:10.1144/GSL.MEM.2002.021.01.23.
- Bonadonna, C., Folch, A., Loughlin, S., Puempel, H., 2011. Future developments in modelling
  and monitoring of volcanic ash clouds: Outcomes from the first IAVCEI-WMO workshop
  on Ash Dispersal Forecast and Civil Aviation. Bull. Volcanol. 74(1), 1–10,
  doi:10.1007/s00445-011-0508-6.
- Bonadonna, C., Costa, A., 2012. Estimating the volume of tephra deposits: A new simple
  strategy. Geology 40(5), 415–418, doi:10.1130/G32769.1.
- Bonadonna, C., Costa, A., 2013. Plume height, volume, and classification of explosive volcanic
  eruptions based on the Weibull function. Bull. Volcanol. 75(8), 1–19, doi:10.1007/s00445013-0742-1.
- 650 Bonadonna, C., Costa, A., Folch, A., Koyaguchi, T, 2015. Tephra Dispersal and Sedimentation.
- 651 *In:* Sigurdsson, H., Houghton, B., McNutt, S., Rymer, H., Stix, J. (Eds.), The Encyclopedia
- of Volcanoes, Elsevier Science, pp. 587–597, http://dx.doi.org/10.1016/B978-0-12385938-9.00033-X.

- Branney, M.J., 1991. Eruption and depositional facies of the Whorneyside Tuff Formation,
  English Lake District: an exceptionally large-magnitude phreatoplinian eruption. Geol.
- 656 Soc. Am. Bull. 103, 886–897.
- Bush, J.W.M., Therber, B.A., Blanchette, F., 2003. Particle clouds in homogeneous and
  stratified environments. J. Fluid Mech. 489, 29–54, doi:10.1017/S0022112003005160
- Carazzo, G., Jellinek, A. M., 2012. A new view of the dynamics, stability and longevity of
  volcanic clouds. Earth Planet. Sci. Lett. 325–326, 39–51.
- 661 Carazzo, G., Jellinek, A. M., 2013. Particle sedimentation and diffusive convection in volcanic
  662 ash-clouds. J. Geophys. Res. 118(4), 1420–1437, doi:10.1002/jgrb.50155.
- Chevalier, L., Donnadieu, F., 2015. Considerations on ejection velocity estimation from
  infrared radiometer data: a case study at Stromboli volcano. J. Volcanol. Geotherm. Res.
  302, 130–140, doi:10.1016/j.jvolgeores.2015.06.022.
- Costa, A., Macedonio, G., Folch, A., 2006. A three-dimensional Eulerian model for transport
  and deposition of volcanic ashes. Earth Planet. Sci. Lett. 241, 634–647,
  doi:10.1016/j.epsl.2005.11.019.
- Costa, A., Folch, A., Macedonio, G., 2010. A model for wet aggregation of ash particles in
  volcanic plumes: Implications for tephra dispersion models, J. Geophys. Res. B: Solid
  Earth 115, B09201, doi:10.1029/2009JB007175.
- 672 Delanoë, J., Protat, A., Vinson, J.P., Brett, W., Caudoux, C., Bertrand, F., Parent du Chatelet,
- J., Hallali, R., Barthes, L., Haeffelin, M., Dupont, J.C., 2016. BASTA: A 95-GHz FMCW
- 674Doppler Radar for Cloud and Fog Studies. J. Atmos. Ocean. Technol. 33, 1023–1037, doi:
- 675 10.1175/JTECH-D-15-0104.1.
- 676 Del Bello, E., Taddeucci, J., Michieli Vitturi, M., Scarlato, P., Andronico, D., Scollo, S.,
- 677 Kueppers, U., Ricci, T., 2017. Effect of particle volume fraction on the settling velocity of

- volcanic ash particles: insights from joint experimental and numerical simulations. Sci.
  Reports 7:39620, doi:10.1038/srep39620.
- 680 Donnadieu, F., 2012. Volcanological applications of Doppler radars: A review and examples
- from a transportable pulse radar in L-Band, in: Bech, J., and Chau, J.L., (Eds.), Doppler
- Radar Observations: Weather Radar, Wind Profiler, Ionospheric Radar, and Other
  Advanced Applications: Intech, ISBN 979–953–307–352–0, doi:10.5772/35940.
- 684 Durant A.J., Rose, W.I., Sarna-Wojcicki, A.M., Carey, S., Volentik, A.C.M., 2009.
- 685 Hydrometeor-enhanced tephra sedimentation: Constraints from the 18 May 1980 eruption
- 686 of Mount St. Helens. J. Geophys. Res. 114, B03204, doi:10.1029/2008JB005756.
- Ernst, G.G.J, Davis, J.P., Sparks, R.S.J., 1994. Bifurcation of volcanic plumes in a crosswind.
  Bull. Volcanol. 56, 159–169.
- Folch, A., 2012. A review of tephra transport and dispersal models: Evolution, current status,
  and future perspectives. J. Volcanol. Geotherm. Res. 235-236, 96–115,
  doi:10.1016/j.jvolgeores.2012.05.020.
- Folch, A., Costa, A., Macedonio, G., 2016. FPLUME-1.0: An integral volcanic plume model
  accounting for ash aggregation. Geoschi. Model Dev. 9, 431–450, doi :10.5194/gmd-9431-2016.
- Freret-Lorgeril V., Donnadieu, F., Eychenne, J., Soriaux, C., Latchimy, T., 2019. *In situ* terminal settling velocity measurements at Stromboli: input from physical characterization
- 697
   of ash. J. Volcanol. Geotherm. Res. 374, 62–79,

   698
   https://doi.org/10.1016/j.jvolgeores.2019.02.005.
- Ganser, G.H., 1993. A rational approach to drag prediction of spherical and non-spherical
  particles. Powder Technol. 77, 143–152, doi:10.1016/0032-5910(93)80051-B.

- Gouhier, M., Donnadieu, F., 2008. Mass estimations of ejecta from Strombolian explosions by
  inversion of Doppler radar measurements. J. Geophys. Res. 113(10), 1–17.
  doi:10.1029/2007JB005383.
- Hobbs, P.V., Radke, L.F., Lyons, J.H., Ferek, R.J., Coffman, D.J., 1991. Airborne
  measurements of particle and gas emissions from the 1990 volcanic eruptions of Mount
  Redoubt. J. Geophys. Res. 96, 18,735–18,752.
- Horwell C.J., Baxter P.J., 2006. The respiratory health hazards of volcanic ash: a review for
  volcanic risk mitigation. Bull. Volcanol. 69(1), 1–24, doi:10.1007/s00445-006-0052-y.
- Hoyal, D.C.J.D., Bursik, M.I., Atkinson, J.F., 1999. Settling-driven convection: A mechanism
  of sedimentation from stratified fluids. J. Geophys. Res. 104, 7953–7966,
  doi:10.1029/1998JC900065.
- Kozono, T., Iguchi, M., Miwa, T, Maki, M., Maesaka, T., Miki, D., 2019. Characteristics of
  tephra fall from eruptions at Sakurajima volcano, revealed by optical disdrometer
  measurements. Bull. Volcanol. 81:41, doi:10.1007/s00445-019-1300-2.
- Löffler-Mang, M., Jürg, J., 2000. An Optical Disdrometer for Measuring Size and Velocity of
  Hydrometeors. J. Atmos. Ocean. Technol. 17, 130–139.
- Manzella, I., Bonadonna, C., Phillips, J.C., Monnard, H., 2015. The role of gravitational
  instabilities in deposition of volcanic ash. Geology 43, 211–214, doi:10.1130/G36252.1.
- 719 Mastin, L.G., Guffanti, M., Servranckx, R., Webley, P., Barsotti, S., Dean, K., Durant, A.,
- Ewert, J.W., Neri, A., Rose, W.I., Schneider, D., Siebert, L., Stender, B., Swanson, G.,
- 721 Tupper, A., Volentik, A., Waythomas, C.F., 2009. A multidisciplinary effort to assign
- realistic source parameters to models of volcanic ash-cloud transport and dispersion during
- ruptions. J. Volcanol. Geotherm. Res. 186, 10–21, doi:10.1016/j.volgeores.2009.01.008.

- Montopoli, M., 2016. Velocity profiles inside volcanic clouds from three-dimensional scanning
  microwave dual-polarization Doppler radars. J. Geophys. Res. Atmos. 121, 7881-7900,
  doi:10.1002/2015JD023464.
- Patrick, M.R., Harris, A.J.L., Ripepe, M., Dehn, J., Rothery, D.A., Calvari, S., 2007.
  Strombolian explosive styles and source conditions: Insights from thermal (FLIR) video.

729 Bull. Volcanol. 69(7), 769–784, doi:10.1007/s00445-006-0107-0.

- Pfeiffer, T., Costa, A., Macedonio, G., 2005. A model for the numerical simulation of tephra
  fall deposits. J. Volcanol. Geotherm. Res. 140, 273–294,
  doi:10.1016/j.volgeores.2004.09.001.
- Ripepe, M., Rossi, M., Saccorotti, G., 1993. Image processing of explosive activity at
  Stromboli. J. Volcanol. Geotherm. Res. 54(3-4), 335–351, doi:10.1016/03770273(93)90071.
- 736 Sauvageot, H., 1992. Radar meteorology, Artech House, ISBN 0890063184, Boston.
- Scharff, L., Hort, M., Varley, N.R., 2015. Pulsed Vulcanian explosions: A characterization of
  eruption dynamics using Doppler radar. Geology 43(11):G36705.1.
- Scollo, S., Prestifilippo, M., Spata, G., D'Agostino, M. & Coltelli, M., 2009. Monitoring and
  forecasting Etna volcanic plumes. Nat. Hazard. Earth Sys. Sci. 9, 1573–1585.
- Scollo, S., Prestifilippo, M., Coltelli, M., Peterson, R.A. & Spata, G., 2010. A statistical
  approach to evaluate the tephra deposit and ash concentration from PUFF model forecasts.
- 743 J. Volcanol. Geotherm. Res. 200, 129–142, doi:10.1016/j.volgeores.2010.12.004.
- 744 Scollo, S., Coltelli, M., Bonadonna, C., Del Carlo, P., 2013. Tephra hazard assessment at Mt.
- Etna (Italy). Nat. Hazard. Earth Sys. Sci. 13, 3221–3233, doi:10.5194/nhess-13–3221–
  2013.

- Scollo, S., Bonadonna, C., Manzella, I., 2017. Settling-driven gravitational instabilities
  associated with volcanic clouds: new insights from experimental investigations. Bull.
  Volcanol. 79:39, doi:10.1007/s00445-017-1124-x.
- 750 Suzuki, Y.J., Koyaguchi, T., 2013. 3D numerical simulation of volcanic eruption clouds during
- the 2011 Shinmoe-dake eruptions. Earth Planet. Space 65, 581–589,
  doi:10.5047/eps.2013.03.009.
- Tokay, A., Wolff, D.B., Petersen, W.A., 2014. Evaluation of the New Version of the LaserOptical Disdrometer, OTT Parsivel<sup>2</sup>. J. Atmos. Ocean. Tech. 31, 1276–1288,
  doi:10.1178/JTECH-D-13-00174.1.
- Turner, J.S., 1960. A comparison between buoyant vortex rings and vortex pairs. J. Fluid Mech.
  7(3), 419–432, https://doi.org/10.1017/S0022112060000189.
- Wilson, T.M., Stewart, C., Sword-Daniels, V., Leonard, G.S., Johnston, D.M., Cole, J.W.,
  Wardman, J., Wilson, G., Barnard, T.S., 2012. Volcanic ash impacts on critical
- 760 infrastructure. Phys. Chem. Earth 45-46, 5–23, https://doi.org/10.1016/j.pce.2011.06.006.