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1 **Influence of atmospheric stability on the size-distribution of the vertical dust flux**
2 **measured in eroding conditions over a flat bare sandy field**

3
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16
17 **Key Points:**

- 18 • The size-resolved dust flux was measured during eight erosion events by the gradient
19 method
- 20 • The size-distribution of the dust flux depends on the surface layer stability conditions
- 21 • The vertical dust flux is enriched in submicron particles in thermally unstable conditions
22

23 Abstract

24 In spite of their importance for the modeling of the atmospheric cycle of mineral dust,
25 measurements of the intensity and size-distribution of the dust emission flux produced by wind
26 erosion in natural conditions remain rare. During the WIND-O-V's (WIND erOsion in presence
27 of sparse Vegetation) 2017 experiment, 8 major erosion events having occurred on a sandy flat
28 field of southern Tunisia were documented. Consistent with the small size (90 μm) of the
29 erodible sand grains and the low aerodynamic roughness length ($Z_0 < 0.079$ cm), the threshold
30 for wind erosion was low ($u_{*t} = 22$ cm s⁻¹). The classical gradient method was applied to assess
31 the size-resolved vertical dust flux, and the stability of the atmosphere quantified by the means of
32 the Richardson number (Ri) as well as of its shear stress (u_*) and thermal gradient ($\partial\theta/\partial z$)
33 components. The vertical dust flux increased with u_* following a power law but the number size-
34 distribution of the dust flux was found to be significantly richer in submicron particles in
35 thermally unstable than in stable periods. This challenges the usual assumption that,
36 independently of their size, the particles smaller than 10 μm follow equally the movements of the
37 air masses in which they are embedded and that the thermal stratification of the surface layer
38 does not affect the size-distribution of the surface flux when measured a few meters above the
39 ground. Finally, we propose a simple empirical method for taking this influence of the thermal
40 instability into account.

41

42 1 Introduction

43 When the stress exerted by the wind becomes large enough, loose particles present on the
44 surface of arid and semi-arid soils are set into motion. Within approximately the first meter
45 above the surface, the size of these wind-eroded particles ranges from a fraction of micrometer to
46 a few millimeters. However, only the particles with a diameter smaller than 20 μm (PM₂₀)
47 (Gillette, 1981; Shao, 2008) have a residence time in the troposphere long enough to be
48 transported hundreds or thousands kilometers away from their source (e.g., Prospero, 1999;
49 Schütz, 1980). According to current estimates (e.g., Boucher et al., 2013), between 1000 to 4000
50 Tg of mineral particles are thus injected yearly into the atmosphere.

51 As compared to the composition of the surface horizon of the parent soils, PM₂₀ are
52 enriched in soil nutrients and their export has negative effects on the fertility of agricultural soils
53 (e.g., Sterk et al., 1996; Webb et al., 2012). Conversely, at the other end of their atmospheric
54 cycle, deposition of eroded particles can constitute a precious source of nutrients for depleted
55 continental (Okin et al., 2004; Swap et al., 1992) or marine (e.g. Duce et al., 1991; Jickells et al.,
56 2005) areas. During their stay in the atmosphere, particles impact the Earth's climate directly by
57 scattering and absorbing solar and terrestrial radiations (e.g., Sokolik et al., 2001; Sokolik &
58 Toon, 1996), but they can also affect it indirectly by favoring the formation of clouds (e.g.,
59 DeMott et al., 2003; Sassen et al., 2003) or modifying their micro-physical properties (e.g., Min
60 et al., 2009; Weinzierl et al., 2017). Finally, airborne mineral dust was identified to impact
61 negatively human health worldwide (e.g., Hashizume et al., 2010; de Longueville et al., 2013;
62 Morman & Plumlee, 2013).

63 Several physical schemes have been developed to predict the characteristics of the PM₂₀
64 emission flux as a function of both the wind shear stress and the state of the soil surface.
65 Basically, two mechanisms are generally considered as being responsible for the emission of

66 these particles: 1) direct aerodynamic entrainment (e.g., Klose & Shao, 2013), and 2) the
67 combination of saltation and sandblasting.

68 In most arid and sparsely vegetated semi-arid soils, the amount of sand-sized grains
69 easily set into motion by the wind is not a limiting factor and the second mechanism is assumed
70 to be predominant (Shao et al., 2011). For wind friction velocities above the threshold of motion,
71 the heavy sand grains start moving in approximately the first meter above the surface (saltation
72 process) and a fraction of their kinetic energy is used for releasing the finer PM₂₀ particles
73 (sandblasting process). Thus, saltation appears as a necessary prerequisite for the emission of the
74 PM₂₀, which is in good agreement with most observations performed either in laboratory wind-
75 tunnels (Alfaro et al., 1997, 1998; Shao et al., 1993) or in outdoor conditions on naturally
76 eroding fields (Gillette, 1977; Gillette et al., 1997; Gillette & Walker, 1977; Gomes et al., 1990;
77 Ishizuka et al., 2014; Sow et al., 2009). Because of the large size of the grains entering saltation,
78 this process is the easiest to observe and in the wake of the pioneering work of Bagnold (1941)
79 an abundant literature has been dedicated first to its parametrization (see the review of Greeley &
80 Iversen (1985)) then to its physical modeling (e.g., Marticorena & Bergametti, 1995; Owen,
81 1964; Shao & Li, 1999). Much fewer theories, in fact only three, have been proposed for
82 explaining the emission of PM₂₀ in presence of saltating sand grains and predicting their size-
83 distribution. Based on wind-tunnel observations, Alfaro & Gomes (2001) considered that the
84 PM₂₀ preexisted in the saltating aggregates or in the soil and that a fraction of the kinetic energy
85 of the saltators was used to counterbalance the binding energies of these fine particles and
86 eventually release them. Shao (2001) also considered that the impact of the saltators on the soil
87 lead to its partial disaggregation. In spite of their differences, the two models agreed on the fact
88 that the size-distribution of the ejected particles should depend on the energy of the saltators, and
89 therefore on wind speed. This point was challenged by Kok (2011a) who proposed the ‘brittle’
90 theory in which it was considered that the PM₂₀ size-distribution did not depend on the speed of
91 the wind (Kok, 2011b). Whatever the exact nature of the process, dust particles are released once
92 the kinetic energy of saltating particles exceeds their binding energy. Then, turbulent diffusion
93 contributes to their transport towards higher levels thus opposing the effect of gravity that tends
94 to maintain them close to the surface.

95 Thus, the vertical flux (F_v) in the Surface atmospheric Boundary Layer (SBL) is the sum
96 of the upward diffusive flux and the downward settling one and is often referred to as the ‘net’
97 flux. It is found to increase with the wind stress following a power law similar to that proposed
98 by Gillette & Passi (1988):

$$99 \quad F_v = C(u_*)^n \left(1 - \frac{u_{*t}}{u_*}\right) \quad (1)$$

100 In this equation, u_* is the wind friction velocity, u_{*t} the threshold wind friction velocity
101 above which saltation occurs, and C and n are constants. In their original work, Gillette & Passi
102 (1988) suggested that the exponent n should be equal to 4 but field measurements indicate that it
103 is actually more variable (e.g., Ishizuka et al., 2014).

104 Two methods exist for measuring F_v . One aims at applying the eddy covariance method
105 to the determination of the size-resolved vertical dust flux (Fratini et al., 2007; Porch & Gillette,
106 1977). Though promising, its application to dust particles remains complicated because of the
107 current lack of instruments being at the same time able to acquire at high frequency the
108 fluctuations of the very low number concentrations of the coarsest dust particles and small
109 enough not to disturb the micrometeorological measurements performed in parallel at the same

110 point. The second method, usually referred to as the ‘gradient method’, was designed by Gillette
 111 et al. (1972). In the framework of the AMMA (African Monsoon Multidisciplinary Analysis;
 112 Redelsperger et al., 2006) international campaign, this method was improved by the use of an
 113 Optical Particle Counter (OPC) to document the effect of u_* on the size-resolved dust emission
 114 flux over a bare agricultural field in Niger (Sow et al., 2009). The method was also used for the
 115 same purpose during the Japan-Australia Dust Experiment (JADE; Ishizuka et al., 2014). This
 116 method is well adapted to the measurement of the dust flux size distribution up to diameters of
 117 approximately 20 μm (Gillette et al., 1972).

118 Basically, the ‘gradient method’ is an adaptation of the K-theory, which considers that, in
 119 the constant flux layer above the ground, the vertical turbulent flux of a scalar is simply
 120 proportional to the vertical gradient of its mean concentration. When applied to dust, this gives:

$$121 \quad F_v = -K_p \frac{dC}{dz} \quad (2)$$

122 with K_p the turbulent eddy diffusivity of the dust particles and C the dust concentration
 123 (in particles m^{-3} or kg m^{-3}).

124 Then, assuming neutral conditions and that the vertical transport of particles with a
 125 diameter $\leq 20 \mu\text{m}$ is similar to that of momentum, Gillette et al. (1972) obtained:

$$126 \quad F_v = u_* k \frac{C_l - C_h}{\ln\left(\frac{z_h}{z_l}\right)} \quad (3a)$$

127 where k is the von Karman’s constant ($k = 0.4$), and C_l and C_h are the dust concentrations
 128 measured in the SBL at a low (z_l) and high (z_h) heights (in m).

129 Because conditions are rarely neutral in natural conditions, Equation (3a) can be
 130 corrected using the stability function ψ_m (Businger et al., 1971) calculated using the Monin-
 131 Obukhov length (L):

$$132 \quad F_v = u_* k \frac{C_l - C_h}{\ln\left(\frac{z_h}{z_l}\right) - \psi_m\left(\frac{z_h}{L}\right) + \psi_m\left(\frac{z_l}{L}\right)} \quad (3b)$$

133 Again, this correction assumes that the vertical transport of particles and momentum are
 134 similar. However, relatively ancient (Businger, 1986; Businger et al., 1971) and more recent
 135 (Abbasi et al., 2017; Li & Bou-Zeid, 2011; Smedman et al., 2007) studies indicate that even in
 136 slightly unstable conditions the ratio of the eddy diffusivity of scalars (water vapor, heat) to that
 137 of momentum increases dramatically, which means that these scalars are transported more
 138 efficiently than momentum. For particles, experimental data are missing but Freire et al. (2016)
 139 used Large Eddy Simulation (LES) to show that even in idealized conditions (spatially and
 140 temporally constant emission flux) atmospheric stability and particle size (in the range 1-30 μm)
 141 should be taken into account in the derivation of the surface net flux from the vertical profiles of
 142 concentration measured in the surface boundary layer. Indeed, the simulations indicate that as for
 143 other scalars the vertical uptake of the finest particles by turbulence should be favored and this
 144 particularly in unstable conditions.

145 Therefore, we hypothesize that atmospheric instability could also play an important role
 146 in the more variable ‘real life’ conditions and have a quantifiable effect on the 1) magnitude and
 147 2) size-distribution of the vertical dust flux in the surface layer. In order to validate or infirm
 148 these assumptions, we will use the results of the measurements performed in the frame of the

149 WIND-O-V's (WIND erOsion in presence of sparse Vegetation) 2017 experiment conducted
150 over a flat bare field of southern Tunisia. The description of the experimental site and the details
151 of the instrumentation and methods are given in the following 'Materials and methods' section;
152 then the impact of atmospheric stability on the magnitude and size-distribution of the vertical
153 flux is analyzed.

154

155 **2 Materials and Methods**

156 2.1 Site description

157 The experiment was carried out from 1 March 2017 to 15 May 2017 in the Dar Dhaoui
158 experimental range (latitude 33°17'45"N, longitude 10°46'57"E) of the Institut des Régions
159 Arides (IRA) of Médenine (Tunisia). The site, described in more details in Dupont et al. (2018,
160 2019), is located in the arid region of the Jeffara Plain. The parent soil is aeolian fine sand
161 deposit generally lying on calcrete (Labiadh et al., 2013). Six composite samples, made of about
162 15 individual samples of loose soil surface particles (about 1 cm depth) were collected each ten
163 meters along radii of the plot, and their size-distributions analyzed. Practically, this was done in
164 two ways. On the one hand, composite soil samples were dry sieved following the methodology
165 established by Chatenet et al. (1996). On the other hand, the soil organic matter was removed
166 using H₂O₂ (H₂O₂ 15% 24 h at 50°C and 72 h at 25°C), the samples were carefully rinsed with
167 demineralized water, disaggregated for 1 minute with ultrasonic agitation and sodium
168 hexametaphosphate added in order to maintain finer particles in suspension. A laser sizer (LS
169 230 Laser Diffraction Particle Size Analyser - Beckman Coulter[®]) was used to measure the
170 totally dispersed soil distribution. In both cases, the median of the size distribution is 90 µm (Fig.
171 1). The mean mass of the soil fraction lower than 10 µm (corresponding approximately to the
172 PM₁₀ studied in the paper) represents 3.25% (Standard Deviation = 0.49%) of the soil sample
173 (Fig. 1b). In order to meet the conditions of an ideal flat bare soil without soil crust or ridges, the
174 surface was tilled with a disc plough and levelled with a wood board before the beginning of the
175 experiment. During the experiment, the soil surface tended to evolve either because of a relative
176 increase of its rough elements, of some rain events, or because of the beginning of vegetation
177 growth. Thus, to maintain the ideal conditions we were seeking the surface was levelled again on
178 13-14 March and 19 April, and tilled and levelled on 29 April-1 May. The measurements were
179 performed at the center of the southern limit of the experimental plot approximating a flat half-
180 circle of 150 m radius (Fig. 2a). In the North, the fetch was slightly longer (about 200 m). The
181 plot was surrounded by small bushes in the North-West (0.34±0.08 m height and 0.58±0.20 m
182 diameter) and young olive trees arranged in a square pattern (about 1.7±0.3 m height, 1.5±0.4 m
183 diameter, and 26 m spaced) in the North-East. The interested reader will find in supplementary
184 material the study showing that, in spite of the relatively short fetch, the instruments located at
185 the 4 m height are still in the constant flux layer, which ensures the applicability of the gradient
186 method.

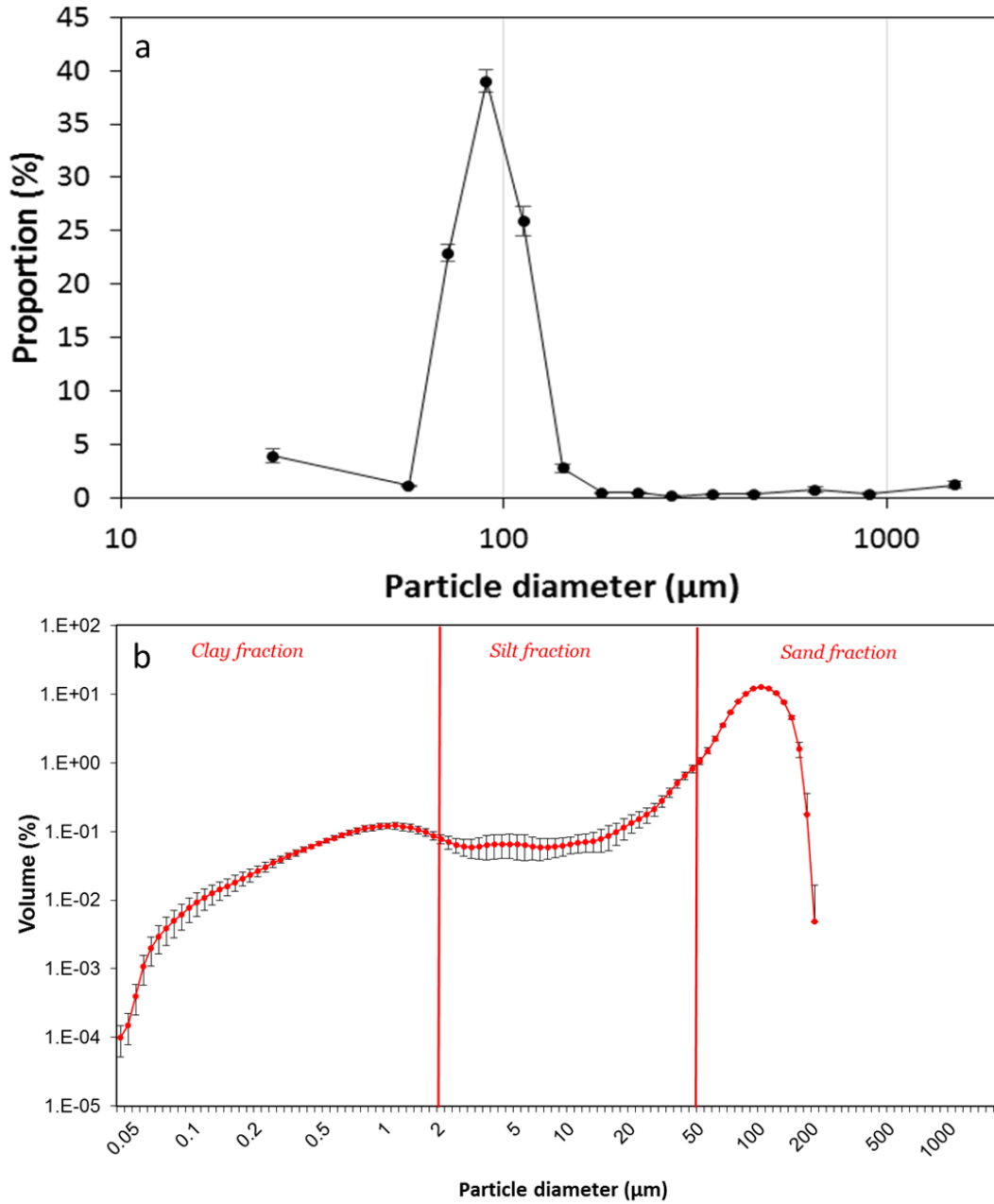


Figure 1. (a) Mean size distribution of the dry sieved soil. (b) Mean size distribution of the totally dispersed soil. Error bars represent the standard deviation of the mean of 6 composite soil samples.

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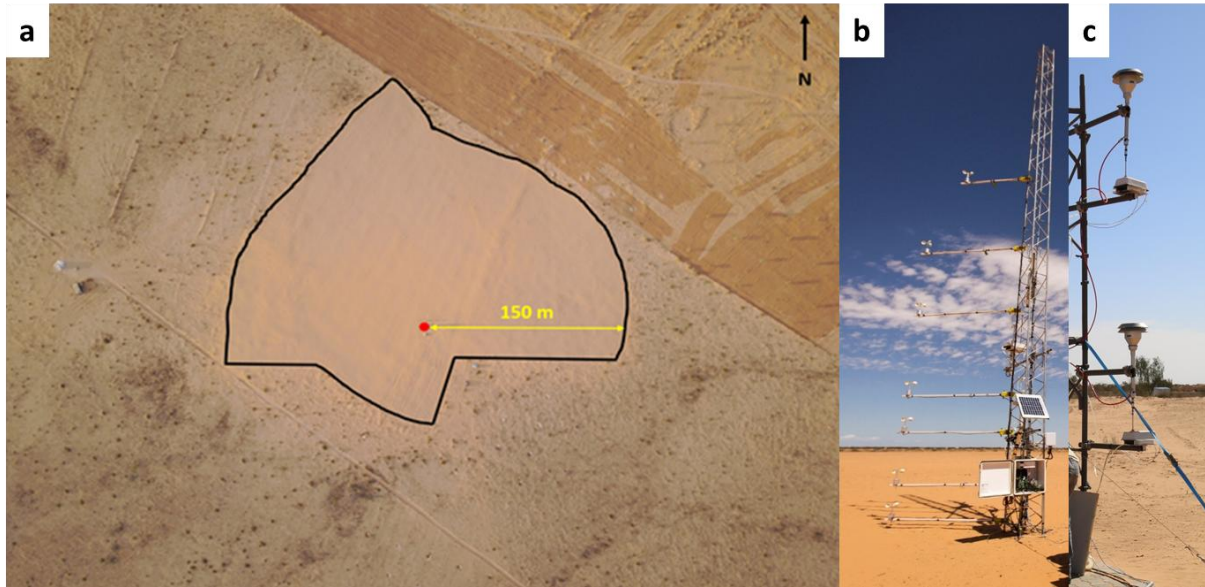


Figure 2. (a) Aerial view of the experimental plot (©IRD – IRA – Christian Lamontagne). The red dot indicates the location of the instruments; (b) View of the meteorological mast with the 7 cup anemometers; (c) View of the two OPCs with their TSP sampling heads.

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2.2 Measurements

197
198 The calculation of the vertical dust flux based upon the gradient method requires to
199 measure (1) the dynamical parameters characterizing the structure of the SBL (the wind friction
200 velocity, u_* , surface roughness length, Z_0 , and Monin-Obukhov's length, L) and (2) the dust
201 concentrations at two heights (Eq. 3). The required instrumentation deployed during the WIND-
202 O-V's 2017 experiment is described in this paragraph.

2.2.1 Meteorological parameters

203
204 A meteorological mast (Fig. 2b) was equipped with 7 cup anemometers (A100R Vector
205 Instrument[®]; resolution = 0.25 m s^{-1}) positioned at 0.22, 0.65, 1.33, 1.83, 3.01, 4.00, and 5.24 m
206 above ground level (agl), and 4 thermocouples (type T copper/Constantan; resolution = 0.2°C)
207 positioned at 0.48, 1.66, 3.83, and 5.07 m agl to measure simultaneously at 0.1 Hz the wind
208 speed and air temperature vertical profiles, respectively. These meteorological measurements
209 allowed the computation of the aerodynamic parameters necessary to apply the gradient method
210 (u_* , Z_0 , and L) as well as other quantities such as the Richardson number allowing to characterize
211 the SBL stability.

212 Before being installed in the gradient position, anemometers were intercompared to test
213 the consistency of their response in natural conditions. In all, 2542 data (5-min averages) were
214 acquired in parallel by the 7 anemometers and their correlation to the mean measurements was
215 excellent ($R^2 > 0.99$, with a slope ranging from 0.99 to 1.02).

216 Wind direction was derived from the measurements performed with a 3D sonic
217 anemometer. Relative humidity was measured using an HMP60 probe (Campbell[®] Scientific
218 Instruments).

219 2.2.2 Dust number concentrations

220 To measure the size resolved dust concentration, we used two OPCs (welas[®], White
 221 Light Spectrometer 2300, PALAS) positioned at 2.04 and 4.10 m agl (Fig. 2c). These
 222 instruments are spectrometers deriving the air-suspended particles number concentration in
 223 predefined size classes from the light they scatter at 90° when illuminated by a white light
 224 source. The suction is done at a controlled flowrate of 5 L min⁻¹ and each OPC was equipped
 225 with an omnidirectional Total Suspended Particles (TSP) sampling head (BGI by Mesa Labs,
 226 Butler, NJ USA). Vanderpool et al. (2018) conducted wind tunnel tests of six commonly used
 227 omnidirectional, low-volume inlets to determine their actual size-selective performance under
 228 conditions of expected use using solid, polydisperse aerosols at wind speeds of 2, 8, and 24 km/h
 229 (i.e., 0.56, 2.2, and 6.7 m s⁻¹, respectively). For the TSP inlet, the estimated total mass
 230 concentration collection efficiency was shown to increase with wind speed. For the coarsest
 231 particle size distribution (mass median diameter of 15 µm with a proportion of 93%), the
 232 estimated total mass concentration collection efficiency varied from 79% at 0.56 m s⁻¹ to 99% at
 233 2.2 m s⁻¹, and 102% at 6.7 m s⁻¹. Considering the range of wind velocity measured during the
 234 emission periods (> 5 m s⁻¹ at about 2 m), we can expect that the use of the TSP inlet will not
 235 induce any significant loss of the coarse dust particles. The nominal flowrate of the TSP inlet is
 236 16.7 L min⁻¹. So this flow was isokinetically split in two: a main flow immediately directed to
 237 the OPC through a short vertical tube to minimize the particle loss before measurement, and an
 238 auxiliary flow (11.7 L min⁻¹) controlled by an automated volumetric flow controller (MCR-
 239 50SLPM, Alicat Scientific). The time resolution of the measurements was 1 min.

240 The calibration was done by the company. The OPC calibration parameters were adjusted
 241 at the beginning of the experiment by following the routine calibration described in the welas[®]
 242 operating manual using the monodisperse caldust 1500 composed of silicon dioxide with a
 243 refractive index of $m = 1.43 - 0i$ in the visible wavelengths and a diameter of 1.28 µm so that the
 244 measured particle diameters correspond to optical diameters of spherical particles having an
 245 extinction efficiency equivalent to that of silica. We checked at the end of the experiment that the
 246 calibration parameters had not changed for the 2 OPCs. The particles are sorted into 15 diameter
 247 bins of equal logarithmic width ($d\log(D)=0.125$) ranging from 0.237 to 17.78 µm and centered
 248 on the mean geometric diameters (noted $D_{g,i}$ in the following) reported in Table S1 of the
 249 supplementary material. As recommended by the manufacturer, the number of particles counted
 250 in the smallest size class ($D_{g,1} = 0.27$ µm) was considered as unreliable and thus discarded.

251 At the beginning and at the end of the campaign, the two OPCs were set up at the same
 252 level, 4.10 meters agl and along a west/east axis, to test the similarity of their response. The
 253 measured concentrations were found to be sufficiently large to allow a relevant inter-comparison
 254 only during the second period, which lasted from 10/05/2017 to 18/05/2017. For this
 255 comparison, data collected when wind directions were in the 80-100° and 260-280° sectors were
 256 rejected to avoid one OPC being shadowed by the other. When relative humidity was greater
 257 than 80%, particle concentration measurements were also rejected to avoid the risk of performing
 258 the inter-comparison in presence of fog droplets to which the instruments are also sensitive but
 259 not calibrated for. The correlation coefficients between the measurements of the two OPCs (16-
 260 min. average concentrations) are presented in Table S1 in supplementary material. It was > 0.9
 261 (for $n = 160$) in all size-classes but the two coarsest in which the number concentrations were
 262 quite low ($D_{g,14} = 11.55$ µm and $D_{g,15} = 15.40$ µm, for which R^2 were 0.87 and 0.74,
 263 respectively). Therefore, these 2 size classes were discarded. The strong correlation existing in

264 the remaining size-classes allows correcting easily one instrument according to the other. In
 265 order to avoid attributing more weight to the measurements performed by any which one of the
 266 two OPCs, they were successively chosen as the reference and the final concentration was
 267 determined as the arithmetical average of the two corrected concentrations.

268 In order to locate temporally the erosion events, we needed to be able to determine when
 269 the concentration at the lower level was significantly larger than that measured at the higher
 270 level. For this, we used the scatter around the mean of the difference of concentrations measured
 271 by the two instruments positioned at the same height and found that this threshold corresponded
 272 to be a relative difference of 23% in the 1.15 μm size class.

273 2.3 Dust flux computation

274 The wind friction velocity (u_*), surface roughness length (Z_0), and Monin-Obukhov's
 275 length (L) can be determined using the measured wind speed and temperature profiles according
 276 to the Monin-Obukhov theory (Monin & Obukhov, 1954):

$$277 \quad U(z) = \frac{u_*}{k} \left[\ln\left(\frac{z}{Z_0}\right) - \psi_m\left(\frac{z}{L}\right) + \psi_m\left(\frac{Z_0}{L}\right) \right] \quad (4)$$

$$278 \quad \Delta\theta = \theta(z_h) - \theta(z_l) = \frac{\theta^*}{k} \left[\ln\left(\frac{z}{Z_0}\right) - \psi_h\left(\frac{z}{L}\right) + \psi_h\left(\frac{Z_0}{L}\right) \right] \quad (5)$$

279 where θ^* is the characteristic dynamical temperature (in K), and $\theta(z_l)$ and $\theta(z_h)$ the
 280 temperature (in K) at measurement heights z_l and z_h (in m), respectively, and ψ_h the heat stability
 281 function.

282 In this study, u_* , Z_0 , and L were obtained using the iterative procedure optimized by
 283 Frangi & Richard (2000) for fitting the wind speed and temperature measurements to Eqs. (4)
 284 and (5), with the stability functions ψ_m and ψ_h respectively defined as:

285 - for $z/L < 0$ (unstable cases; Dyer & Hicks, 1970; Hicks, 1976; Paulson, 1970):

$$286 \quad \psi_m\left(\frac{z}{L}\right) = 2 \ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2 \arctan(x) + \frac{\pi}{2} \quad (6)$$

$$287 \quad \psi_h\left(\frac{z}{L}\right) = 2 \ln\left(\frac{1+y}{2}\right) \quad (7)$$

288 With (Dyer, 1974):

$$289 \quad x = \left[1 - 15 \frac{z}{L}\right]^{1/4} \quad \text{and} \quad y = \left[1 - 15 \frac{z}{L}\right]^{1/2} \quad (8)$$

290 - for $z/L > 0$ (stable cases; (Webb, 1970):

$$291 \quad \psi_m = \psi_h = -5 \frac{z}{L} \quad (9)$$

292 Noteworthy, the gradient method used for the inversion of the dynamical parameters
 293 assumes the use of at least 10-min wind speed averages so that the major time scales of
 294 turbulence occurring in the SBL are integrated (Wieringa, 1993). In this study, we used 16-min
 295 averages for wind speed and air temperature following the results obtained by Dupont et al.
 296 (2018) on the same field.

297 The quality of the inversion of the dynamical parameters was carefully examined by
 298 using criteria similar to those defined by Marticorena et al., (2006):

- 299 1. Computation of the dynamical parameters were only done for directions where the
 300 experimental device could not perturb the measurements, i.e. between [265°, 360°] and
 301 [0°, 95°] (Fig. 2a).
- 302 2. Although the minimal wind speed that anemometers can measure is 0.25 m s⁻¹, we
 303 considered that a minimum wind speed of 1 m s⁻¹ was required to have a sufficient
 304 precision to obtain a correct wind profile. Thus, the computation of the aerodynamic
 305 parameters were performed only when all the anemometers measured a wind speed
 306 greater than 1 m s⁻¹.
- 307 3. The quality of the inversion procedure was considered acceptable only if the difference
 308 between the computed and measured profiles was less than 5% for wind speeds and less
 309 than 0.2 K for temperatures.
- 310 4. To eliminate the possible cases of free convection, the profiles for which the fitted wind
 311 friction velocity u_* was lower than 0.2 m s⁻¹ were also rejected.

312

313 In the “gradient configuration”, a dust emission event is defined as a period during which
 314 1) the concentration of the lowest OPC in one of the most abundant size-classes (1.15 μm) is
 315 greater than 5 particle cm⁻³ and 2) the relative difference $((C_l - C_h)/C_l)$ between the bottom and
 316 upper concentrations is positive and larger than the threshold defined from the intercomparison
 317 of the two OPCs. For the same reason as during the OPC intercomparison, particle concentration
 318 measurements were also rejected when relative humidity was greater than 80%. The vertical dust
 319 flux was calculated only when all these conditions were fulfilled.

320 The total number flux, $F_{v,n}$, over the size range 0.32-10.00 μm in diameter is thus
 321 obtained by summing the dust fluxes calculated in each of the twelve retained individual size
 322 classes:

$$323 \quad F_{v,n} = \sum_{i=2}^{13} \left(\frac{dF_{v,i}}{d\log(D_{g,i})} \right) d\log(D_{g,i}) \quad (10)$$

324 The derivation of the total mass flux, $F_{v,m}$, from the total number flux, $F_{v,n}$, was done (i)
 325 assuming spherical particles and (ii) using the equivalent mass density measured by Sow et al.
 326 (2009) during dust emission events in Niger ($\rho_d = 2.38 \pm 0.24$ g cm⁻³). The hypothesis of
 327 particle sphericity is generally adopted even though several studies suggest that the aspect ratio
 328 for African dust particles is probably close to 1.5-1.7 (e.g., Chou et al., 2008; Kandler et al.,
 329 2007; Klaver et al., 2011). The equivalent mass density we used is slightly lower than the density
 330 of pure quartz (2.65 g cm⁻³) generally adopted for dust particles in modelling studies but the
 331 value of 2.38 g cm⁻³ appears reasonable because African dust particles are known to be complex
 332 clusters of various minerals, mainly clays, calcite, quartz, feldspars and iron oxides (e.g., Kandler
 333 et al., 2009; Nowak et al., 2018) with voids in-between.

334 Noteworthy, the flux yielded by the gradient method (Equation 3b) is actually the net
 335 vertical flux (positive upward), which is to say the sum of the diffusive (F_d) and the negative
 336 gravitational settling (F_s) fluxes.

$$337 \quad F_v = F_d + F_s \quad (11)$$

338 In the gradient theory, F_d is the flux assumed to be constant from the surface to the top of
 339 the surface boundary layer. Therefore, F_d is supposed to represent the flux of particles released
 340 from the surface before it is modified by upward transport and sedimentation.

341 F_s , which is simply the product of the gravitational settling velocity (given by the Stokes
 342 formula) and concentration for a given particle size, also depends on the altitude. Therefore, F_v is
 343 also elevation-dependent in the SBL and should be corrected to compensate the effect of
 344 gravitation (Shao, 2008) to retrieve F_d . However, in their exploitation of the JADE campaign,
 345 Shao et al. (2011) showed that this size-dependent correction is relatively limited as it was less
 346 than 2% for particles smaller than 3 μm and did not exceed 15% for particles with sizes up to
 347 8.4 μm for $\Delta z \sim 2$ m. We also computed the correction for all size classes and found it to be less
 348 than 1% for particles smaller than 3.5 μm and less than 4% for particles with sizes ranging from
 349 4 μm to 8.5 μm . Though small, this correction was applied to all the vertical dust fluxes we
 350 computed.

351 2.4 Atmospheric stability

352 Atmospheric stability is a result of the combination of thermal and dynamical effects: in a
 353 convective (unstable) boundary layer, turbulence is generated by buoyancy; in stable boundary
 354 layer, turbulence is suppressed by buoyancy while in neutral boundary layer, turbulence is
 355 generated by wind shear (Stull, 1988). Several parameters (see for instance Golder, 1972) can be
 356 used to quantify the stability of the atmosphere: the Monin-Obukhov length, the Richardson
 357 number, the bulk Richardson number, the Pasquill and Turner categories.... In this study, we
 358 selected the Richardson number, Ri (Richardson, 1920), which compares directly the
 359 atmospheric thermal source of instability to the dynamical one:

$$360 \quad Ri = \frac{g}{T} \frac{\partial\theta/\partial z}{(\partial u/\partial z)^2} \quad (12)$$

361 where g is the acceleration of gravity (in m s^{-2}), T the absolute air temperature (in K), θ
 362 the potential temperature (in K), and u the mean wind speed (in m s^{-1}).

363 The sign of $\partial\theta/\partial z$ (equivalently that of Ri) indicates whether the atmosphere is
 364 thermally stable ($Ri > 0$) or unstable ($Ri < 0$), and the magnitude of Ri gives information on the
 365 predominance of dynamical vs. thermal effects on atmospheric stability. Note that at the very
 366 low altitude above sea-level of our experimental site (50 m), the potential (θ) and absolute (T) air
 367 temperatures can be approximated to be equal.

368 Practically, we have computed Ri using the equations proposed by Arya (1988) who
 369 expressed it as a function of the Monin-Obukhov length, L :

370 For $Ri < 0$:

$$371 \quad \frac{Z_m}{L} = Ri(Z_m) \quad (13)$$

372 For $0 \leq Ri < 0.2$:

$$373 \quad \frac{Z_m}{L} = \frac{Ri(Z_m)}{1-5Ri(Z_m)} \quad (14)$$

374 with Z_m (=1.56 m) the geometric mean of the minimum (Z_{min} =0.48 m) and maximum
 375 (Z_{max} =5.07 m) heights of the air temperature measurements.

376 **3 Results**

377 3.1 The dust emission events

378 The two OPCs were set in the gradient position from 5 March to 9 May 2017, and 8
 379 periods were found to correspond to dust emission events identified according to the previously
 380 defined criteria (Tab. 1). Their durations vary from approximately 5 to 6 hours for Events 2, 5
 381 and 6 to more than 9 hours for Events 4, 7 and 8. During the majority of these events, the wind
 382 blew rather constantly either from NNW (Events 1, 2, 3, and 5) or from NE (Events 4, 7, and 8).
 383 It only turned progressively from NE to NW during Event 6.

384 The minimal value of u_* appearing in Table 1 corresponds to the smallest vertical upward
 385 flux quantifiable by the gradient method. This minimum is 0.21 m s^{-1} for the whole measurement
 386 period, which is consistent with the threshold value ($u_{*t}=0.22 \text{ m s}^{-1}$) yielded by the sand
 387 movement detector (saltiphone) operated on the same site (Dupont et al., 2018). Noteworthy, the
 388 maximal value achieved by u_* (Tab. 1) is always well above u_{*t} . This ensures that a wide range
 389 of variation of the erosion conditions was monitored during each one of the 8 events of this
 390 study.

391 The lowest roughness ($Z_0 \sim 0.03 \cdot 10^{-3} \text{ m}$) was measured during periods when the saltation
 392 was minimum. This value is in agreement with the surface roughness lengths measured in
 393 absence of saltation over flat bare surfaces (e.g., Marticorena et al. (2006) and references
 394 therein). When saltation develops, the range of variation of Z_0 is enlarged to about $3 \cdot 10^{-3} \text{ m}$,
 395 which is also consistent with the values measured over flat bare eroding soils (Dupont et al.,
 396 2018; Gomes et al., 2003; Rajot et al., 2003; Shao et al., 2011; Sow et al., 2009). Indeed, in this
 397 case Z_0 , which is then frequently referred to as the ‘apparent roughness length’, increases due to
 398 the presence of saltating particles and becomes larger than the actual roughness of the surface
 399 (Gillette et al., 1998; Owen, 1964; Raupach, 1991).

400

401 **Table 1.** Date, beginning and duration of the 8 dust events. The ranges of variation of u_* , and the
 402 wind direction (clockwise and with 0 to the north) are also indicated as well as the minimum Z_0
 403 values.

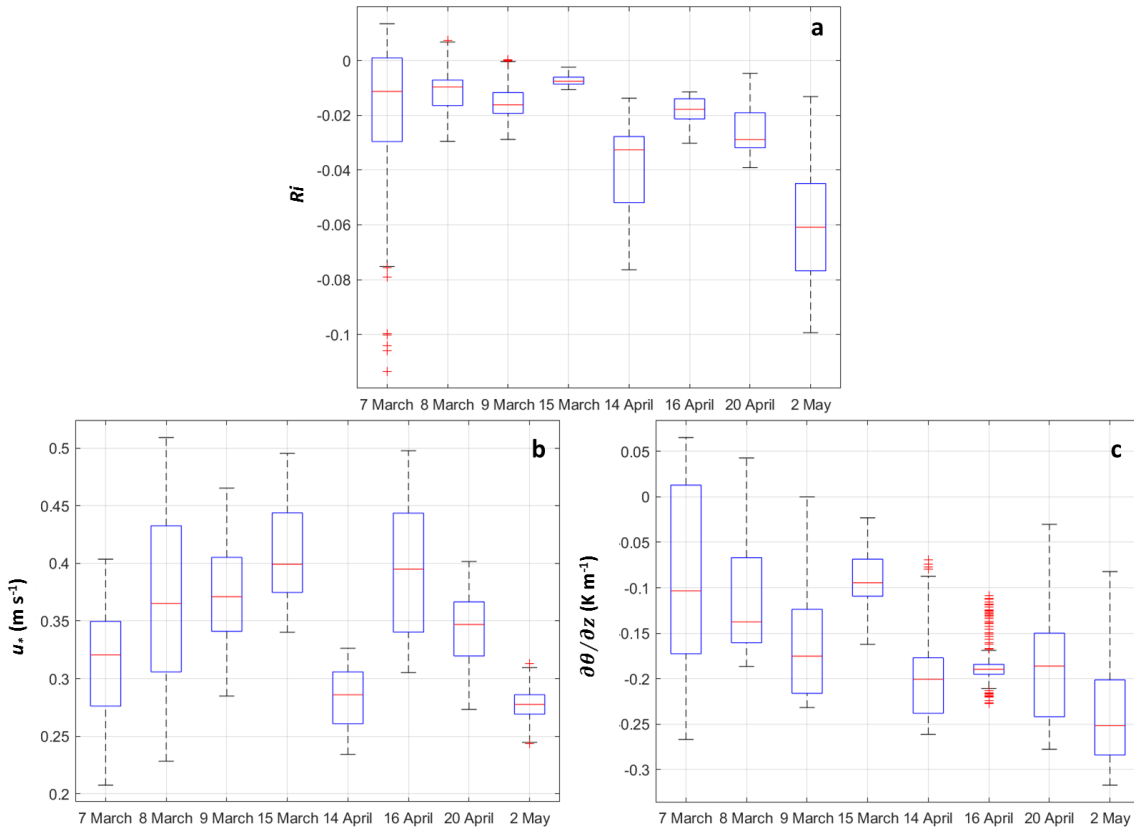
Event details				u_* (m s^{-1})		Z_0 (10^{-3} m)	Wind Dir. ($^\circ$)	
N	Date	Beg. (UTC)	Dur. (h)	Min	Max	Min	Min	Max
1	07/03/2017	10:20	7.2	0.21	0.40	0.03	288	348
2	08/03/2017	11:34	5.4	0.23	0.51	0.10	322	10
3	09/03/2017	08:16	8.4	0.28	0.47	0.16	318	344
4	15/03/2017	06:56	9.3	0.34	0.50	0.10	55	72
5	14/04/2017	10:20	5.6	0.23	0.33	0.03	273	292
6	16/04/2017	09:51	5.9	0.31	0.50	0.13	325	53
7	20/04/2017	07:31	9.1	0.27	0.40	0.05	32	69
8	02/05/2017	06:54	9.8	0.24	0.31	0.02	48	75

404

405 We also assessed the range of variability of the instability conditions during the 8
 406 periods, by examining the range of variations of Ri (Fig. 3a), and its mechanical stress (u_* ; Fig.

407 3b) and thermal ($\partial\theta/\partial z$; Fig. 3c) components. As expected for erosion events during which the
 408 wind gradient is by definition large, Ri remained small ($-0.12 < Ri < 0.02$) and its range of
 409 variation limited (Fig. 3a) as for instance during the 15 March event that remained constantly
 410 close to neutrality. However, with their slightly more negative Ri , some events such as 14 April
 411 and 2 May appear as relatively more unstable than the others. This instability results from the
 412 combination of a relatively strong negative thermal (Fig. 3c) and a weak wind ($u_* < 0.32 \text{ m s}^{-1}$;
 413 Fig. 3b) gradients.

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415

416 **Figure 3.** Boxplots of the variations of the Richardson number Ri (a), wind friction velocity u_*
 417 (b), and thermal gradient $\partial\theta/\partial z$ (c) during the 8 dust emission events. Whiskers are drawn from
 418 the ends of the interquartile ranges to the furthest observations within the whisker length that is
 419 equal to 1.5 times the interquartile range at a maximum. Values beyond this limit (red crosses)
 420 are defined as outliers.

421

3.2 Influence of the stability conditions on the vertical dust flux

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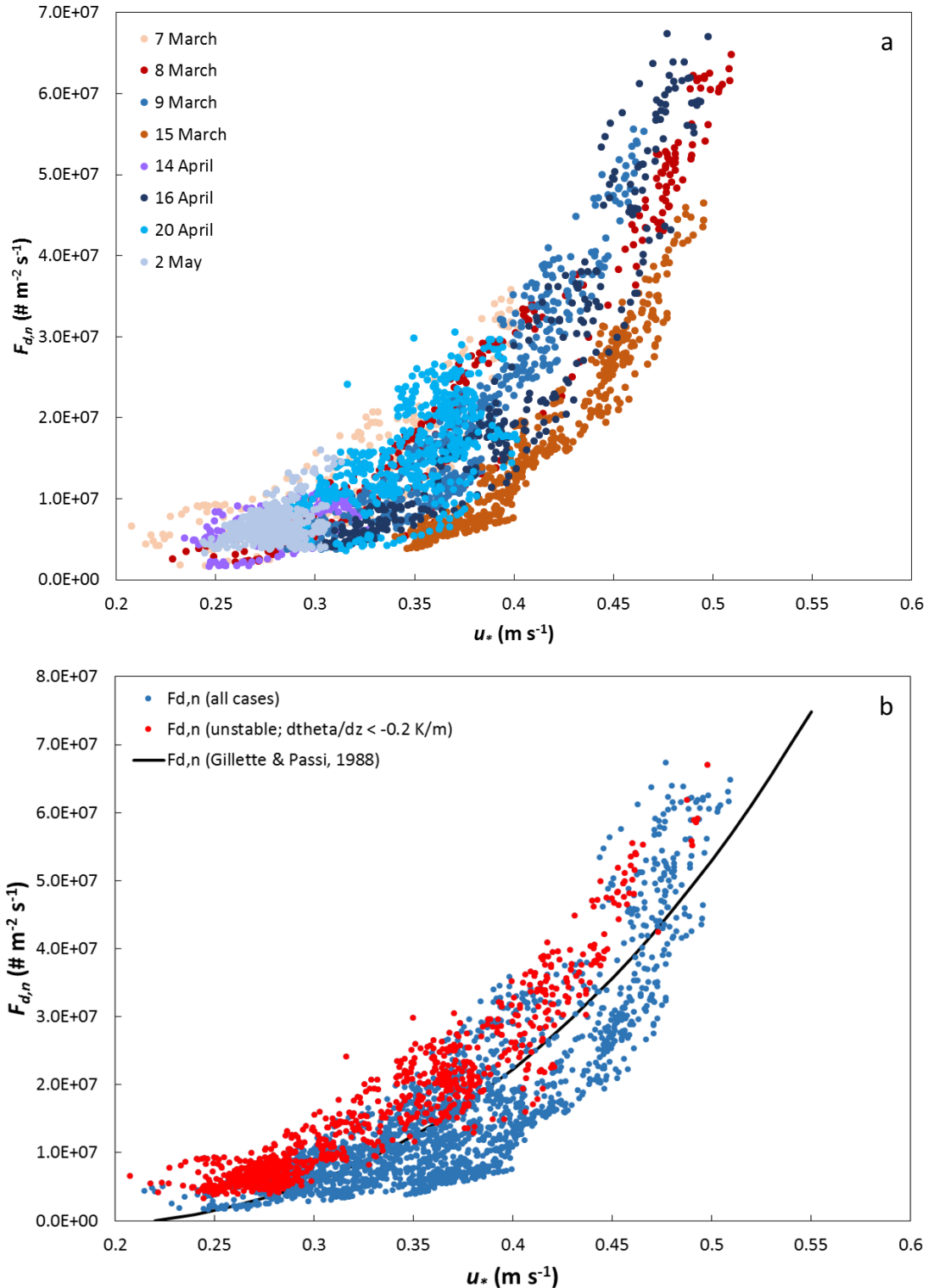
3.2.1. Vertical number dust flux

423

424 Figure 4a illustrates the variations of $F_{d,n}$ with u_* for each of the eight dust emission
 425 events. $F_{d,n}$ increased from nearly 0 to more than $6 \cdot 10^7 \text{ particles m}^{-2} \text{ s}^{-1}$ when u_* increased from
 426 u_{*t} (0.21 m s^{-1}) to 0.51 m s^{-1} . As expected from the smallness of the gravitational settling flux
 (F_s), the general behavior of the diffusive flux (F_d) is consistent with the power law (Equation

427 (1)) describing the evolution of F_v with u_* . The constants C and n ($7.1 \cdot 10^8$ and 2.90,
428 respectively) were determined by the means of a least square iterative routine and the adjustment
429 found to be satisfactory (slope=0.93, vertical intercept= $1.73 \cdot 10^6$, $R^2=0.76$, $N=3135$), thus
430 confirming that the wind friction velocity is the main driver of the vertical dust flux. However, a
431 relatively large scatter of the experimental data around the power law can still be observed. For
432 instance, for similar u_* in the range $0.4\text{-}0.5 \text{ m s}^{-1}$, $F_{d,n}$ is approximately 2 to 3 times larger on 9
433 March than on 15 March. This shows that u_* is not the sole predictor of the magnitude of the
434 vertical dust flux.

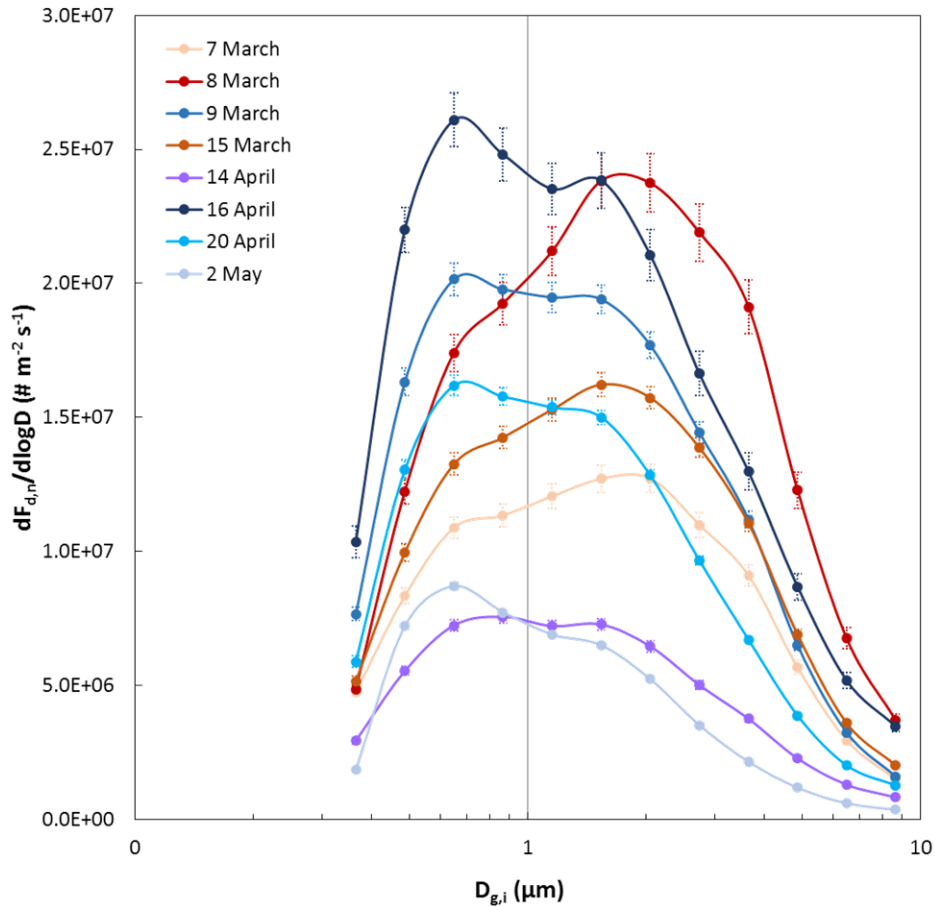
435 In the introduction, we hypothesized that stability conditions could have an effect on the
436 efficiency of the vertical transfer of the emitted dust. This assumption is roughly confirmed by
437 our measurements. Indeed, when the $F_{d,n}$ data acquired in thermally ‘unstable’ conditions (i.e.,
438 when $\partial\theta/\partial z < -0.2 \text{ K m}^{-1}$) are distinguished from the others (Fig. 4b), it appears that the total
439 vertical dust number flux is quasi-systematically above the value predicted by the average
440 ‘power’ law even if the two categories seem to converge at large u_* ($> 0.45 \text{ m s}^{-1}$), which is to
441 say when the production of turbulence by wind shear is likely to predominate largely over that by
442 thermal instability.



443
 444 **Figure 4.** Total number vertical dust flux ($F_{d,n}$ in particles $\text{m}^{-2} \text{ s}^{-1}$) as a function of the wind
 445 friction velocity (u_* , in m s^{-1}) during the 8 wind erosion events (a), and sorted according to their
 446 thermal stability (b) (red points were acquired when $\partial\theta/\partial z$ was $< -0.2 \text{ K m}^{-1}$; the blue points
 447 represent the remainder of the dataset). In (b), the black solid line corresponds to the adjustment
 448 of the “power law” of Gillette & Passi (1988) to the entire dataset (see coefficients in the text).

449 3.2.2. Size distribution of the vertical number dust flux

450 Figure 5 displays the size distributions of the vertical number dust flux averaged over the
 451 duration of each dust event. These size distributions appear to be mainly bi-modal with one fine
 452 mode centered on $0.6 \mu\text{m}$ and another coarser and more variable one around $1.5 \mu\text{m}$ (Fig. 5). The
 453 relative proportions of these two modes vary from one event to the other. For instance, the coarse
 454 mode is more abundant during events 1 (7 March), 2 (8 March) and 4 (15 March), than during
 455 the other ones. As denoted by the length of the error bars (standard error of the mean), the
 456 variability of the size-distribution inside a given event is quite large.

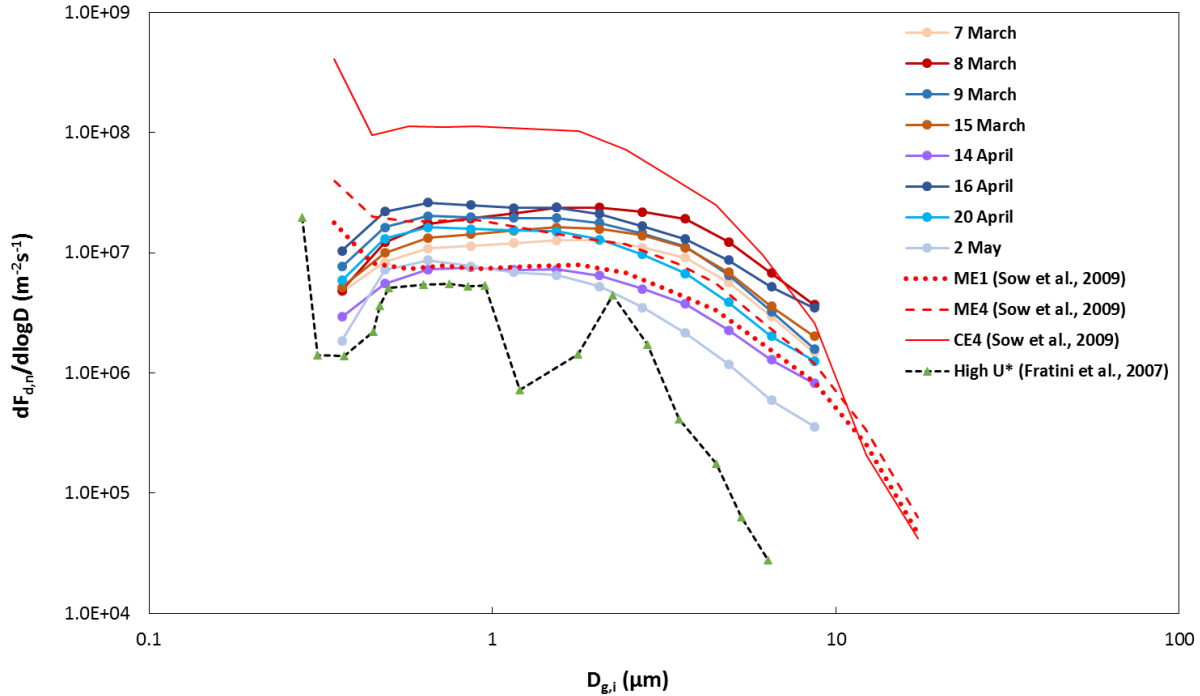


457
 458 **Figure 5.** Average size-resolved number vertical dust flux ($F_{d,n}$ in particles $\text{m}^{-2} \text{s}^{-1}$) for the 8 dust
 459 emission events of this study. Vertical dotted bars represent the standard error of the mean.

460
 461 Figure 6 presents a comparison of the size distributions of the vertical number dust flux
 462 measured during the 8 dust events of the WIND-O-V's 2017 campaign with those measured by
 463 Fratini et al. (2007) in northern China (Gobi desert) and by Sow et al. (2009) in Niger (Sahel)
 464 during the Special Observing Periods of the AMMA international program. It can be seen that
 465 the magnitude of the dust flux measured during the WIND-O-V's 2017 campaign compares to
 466 those measured by Fratini et al. (2007) and by Sow et al. (2009) during Monsoon Events (ME1
 467 and ME4), but is from one order of magnitude lower than those measured during Convective
 468 Events (CE4). Part of these differences can be explained by the range of variation of u_* : the high

469 friction velocity class defined by Fratini et al. (2007) refers to the peak of the dust event, i.e., to
 470 $0.45 < u_* < 0.60 \text{ m s}^{-1}$, and during the monsoon events ME1 and ME4 u_* ranged from 0.40 to
 471 0.50 m s^{-1} . In both cases, it corresponds to the range of variation of u_* measured during the 8
 472 dust events of the WIND-O-V's 2017 campaign (Tab. 1). Conversely, u_* reached values as high
 473 as 0.8 m s^{-1} during the CE4 event reported by Sow et al. (2009).

474

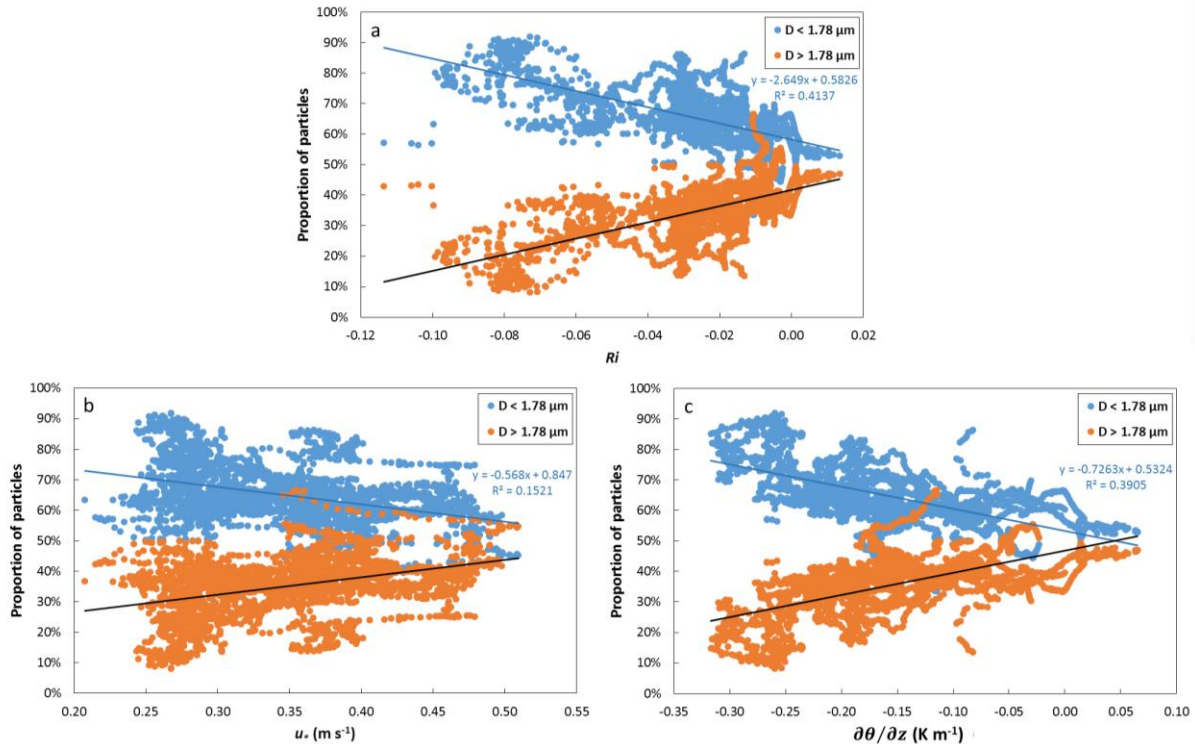


475

476 **Figure 6.** Average size-resolved number vertical dust flux ($F_{d,n}$ in particles $\text{m}^{-2} \text{s}^{-1}$) for the 8 dust
 477 emission events of this study (7 March to 2 May), for the ME1, ME4 and CE4 events from Sow
 478 et al. (2009), and for the high u_* event from Fratini et al. (2007).

479

480 As a first clue for understanding the variability of the size-distribution, it can be noted
 481 that the events whose average vertical fluxes are the richest in fine particles (16 April, 2 May...)
 482 are also globally the most unstable (see their range of R_i and $\partial\theta/\partial z$ in Fig. 3). Therefore, it
 483 seems necessary to study in more detail not only the influence of R_i on the size distribution of
 484 $F_{d,n}$, but also those of its u_* and $\partial\theta/\partial z$ components. For doing so, we have calculated the
 485 cumulative proportions of ‘fine’ and ‘coarse’ (with a separation set arbitrarily at $D = 1.78 \mu\text{m}$
 486 after visual observation of Fig. 5) particles in the vertical flux and plotted them as a function of
 487 R_i (Fig. 7a), u_* (Fig. 7b) and $\partial\theta/\partial z$ (Fig. 7c).



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Figure 7. Evolution of the proportions of particles with diameter $< 1.78 \mu\text{m}$ (blue dots – $N = 3135$) and $> 1.78 \mu\text{m}$ (orange dots – $N = 3135$) as a function of the Richardson number Ri (a), the wind friction velocity u_* (b), and the thermal gradient $\partial\theta/\partial z$ (c). All the measurements performed during the 8 dust emission events were considered. The linear regression line and the quality of the regression (R^2) are also reported.

The relative proportion of fine particles and that of the coarse particles (which is simply its complementary) clearly evolve linearly ($R^2=0.41$) with Ri (Fig. 7a). Though significant, the correlation with u_* is weaker ($R^2=0.16$) (Fig. 7b). Thus, it can be concluded that rather than u_* this is mostly the thermal component of Ri (namely, $\partial\theta/\partial z$) that controls the proportions of fine and coarse particles in the vertical flux (Fig. 7c). This influence of the thermal instability on the size-distribution can be further evidenced (Fig. 8) by sorting all the available data in categories of increasing instability on the basis of their $\partial\theta/\partial z$ values. The limits of the $\partial\theta/\partial z$ classes were chosen in order to 1) cover the full range of variability of the thermal gradient, and 2) ensure that a representative number of cases (N , reported in Table 2) were present in each of these categories. It clearly appears (Fig. 8) that the proportion of particles smaller (larger) than $1.78 \mu\text{m}$ in the vertical flux increases (decreases) with the thermal instability.

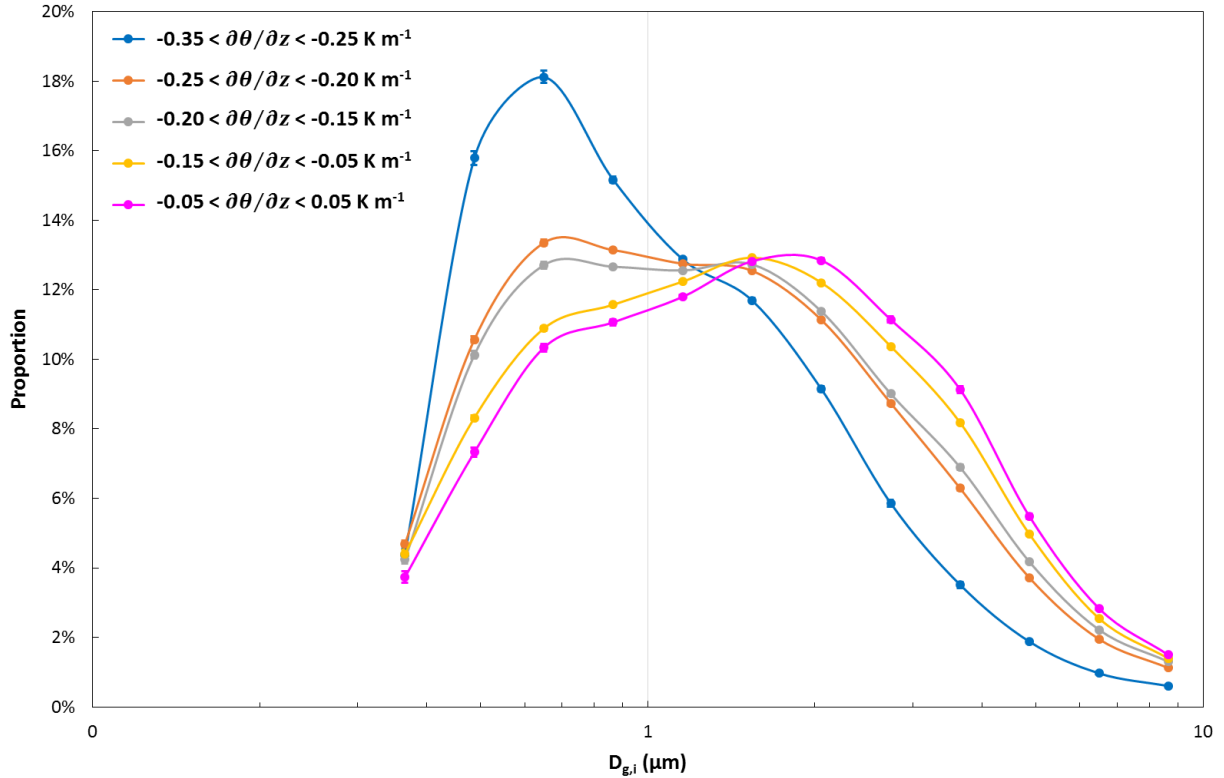


Figure 8. Relative size-resolved number emission flux for different classes of thermal stability. Vertical dotted bars represent the standard error of the mean.

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3.2.3. Modes of the vertical dust flux size distribution

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In order to characterize more precisely the evolution of the size-distribution, we have determined the geometrical mean diameters (gmd), geometrical standard deviations (gsd), and proportions of the two lognormally distributed populations of particles present in the 5 normalized vertical fluxes of Fig. 8. The quality of the deconvolution is excellent ($R^2 > 0.99$, slope $= 1.00 \pm 0.02$, vertical intercept $< 0.05\%$), and the gmd, gsd and proportions of the two particles populations are reported in Table 2. The size characteristics of the fine mode (gmd = $0.58 \pm 0.02 \mu\text{m}$ and gsd = 1.37 ± 0.10) are independent of the thermal stability. Its proportion is maximal (26.2%) in the most unstable category ($-0.35 < \partial\theta/\partial z < -0.25$), but the evolution of the size-distribution is in large part explained by changes of the coarse mode. Indeed, its breadth remains rather constant (gsd = 2.12 ± 0.13) but its gmd decreases continuously from $1.93 \mu\text{m}$ in quasi-neutral conditions to $1.20 \mu\text{m}$ in the most unstable ones. This progressive shift of the gmd of the coarse mode towards finer diameters is responsible for the increase of the relative proportion of submicron particles in the vertical flux.

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After conversion of the number distributions into volume ones, the proportion of the total volume occupied by the submicron mode is found to be less than 2% (Table 2) and therefore negligible. Except in the most unstable case when it is significantly smaller ($4.62 \mu\text{m}$), the gmd of the coarse mode is close to the upper limit of the measuring capacity of the OPC ($10 \pm 1 \mu\text{m}$).

529

530 **Table 2.** Results of the deconvolution of the number size-resolved dust emission fluxes for different classes of thermal stability. In
 531 each class, the dust emission flux can be considered as a mixture of 2 lognormally distributed populations whose geometric standard
 532 deviation (gsd), geometric mean diameters (gmd, in μm), and proportions (%) are reported. The number of points N in each class is
 533 reported in parenthesis. The gmd and proportions obtained after number to mass conversion are also indicated.

		$-0.35 < \frac{\partial\theta}{\partial z} < -0.25$		$-0.25 < \frac{\partial\theta}{\partial z} < -0.20$		$-0.20 < \frac{\partial\theta}{\partial z} < -0.15$		$-0.15 < \frac{\partial\theta}{\partial z} < -0.05$		$-0.05 < \frac{\partial\theta}{\partial z} < 0.05$	
		$(N = 415)$		$(N = 600)$		$(N = 827)$		$(N = 992)$		$(N = 301)$	
		Pop. 1	Pop. 2	Pop. 1	Pop. 2	Pop. 1	Pop. 2	Pop. 1	Pop. 2	Pop. 1	Pop. 2
gsd		1.3	2.0	1.3	2.2	1.3	2.3	1.5	2.1	1.5	2.0
gmd (μm)	in number	0.57	1.20	0.58	1.39	0.58	1.43	0.57	1.71	0.61	1.93
	in mass	0.67	4.62	0.74	9.39	0.71	10.49	0.88	9.71	0.99	8.70
%	in number	26.2	75.0	15.5	88.1	13.2	90.7	18.3	84.8	22.5	79.9
	in mass	0.6	99.4	0.1	99.9	1.8	98.2	0.1	99.9	0.2	99.8

534

535 **4 Discussion**

536 Because of the constant flux layer assumption underlying the gradient method, the
 537 diffusive flux F_d is supposed to represent the emission flux (F_{emiss}), which is to say the vertical
 538 upward flux of particles leaving the surface from which they were released. Our results show that
 539 both the intensity and size distribution of F_d measured between 2.04 and 4.10 m above the
 540 ground depend on the thermal structure of the SBL.

541 It is physically difficult to conceive that the magnitude and size-distribution of the flux at
 542 the immediate interface between the soil and the atmosphere might be affected by the thermal
 543 stratification of the air layer above it. Thus, our results strongly suggest that the vertical
 544 temperature gradient alters significantly the efficiency of the transfer of particles from the
 545 surface to higher altitudes and that this effect is size-dependent.

546 This contradicts two fundamental assumptions of the gradient method, namely that 1) the
 547 diffusion coefficient (K_p) of Equation (2) is independent of D_p , which is to say that all particles
 548 with diameters smaller than approximately 10 μm have the same ability to follow the movements
 549 of the air-masses in which they are embedded (Gillette et al., 1972), and 2) it is possible to
 550 account for the effect of thermal stratification on the vertical transport of these fine particles by
 551 using the momentum stability functions.

552 **4.1 Dependence of K_p on particle size in neutral conditions**

553 K_p is expected to depend on particle size even in neutral conditions because of the
 554 trajectory crossing effect. This theory proposed by Csanady (1963) states that because of their
 555 inertia large particles respond more slowly to turbulent fluctuations than smaller ones, are less
 556 apt to follow the trajectories of the air parcels in which they were initially embedded, and are
 557 thus more prone to shift from one turbulent eddy to another. The author suggested that the effect
 558 should be significant for particles larger than about 20 μm in highly turbulent conditions but also
 559 for smaller particles in less turbulent conditions. Later on, Wang & Stock (1993) showed that for
 560 coarse particles K_p should not only decrease with particle size because of the trajectory crossing
 561 effect but also depend on the intensity of the turbulence (the lower this intensity, the larger the
 562 deviation of K_p from the eddy-diffusivity of the air-parcels). In summary, K_p should depend not
 563 only on particle size but also on the structure of the turbulence. Unfortunately, for lack of
 564 experimental data it has not been possible so far to test the applicability of these results to
 565 particles smaller than 20 μm .

566 **4.2 Dependence of K_p on particle size in thermally unstable conditions.**

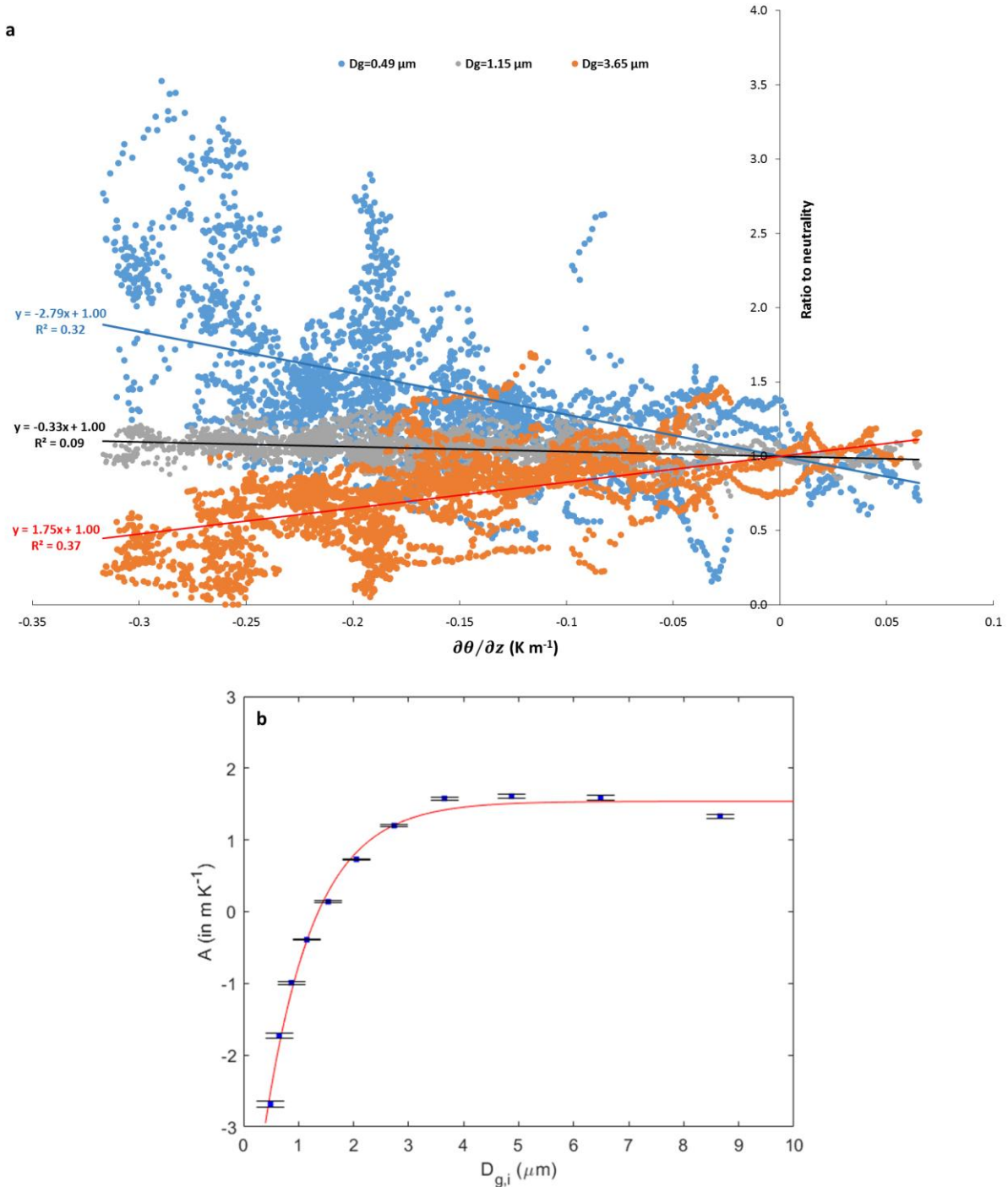
567 Our results showed that, though not exclusively, the influence of atmospheric instability
 568 on the characteristics of the vertical flux was more strongly felt during the 14 April and 2 May
 569 events, which is to say during events characterized by the lowest u_* and the strongest thermal
 570 gradients. These conditions correspond to the so-called ‘Unstable Very Close to Neutral’
 571 (UVCN) regime studied by Smedman et al. (2007). These authors showed that in such situations,
 572 the eddy structure in the surface layer was considerably altered by the weak buoyancy, which in
 573 turn enhanced the efficiency of the heat and water vapor upward transfers. As seen above, the
 574 finest particles are more apt than the coarsest ones to follow the air-parcels movements and could
 575 thus be more liable to benefit from the apparition of new turbulent structures favoring their
 576 vertical transport from the surface to higher atmospheric levels. Note that this interpretation is

577 also consistent with the theory of Wang & Stock (1993) extrapolated to diameters $< 10 \mu\text{m}$.
 578 Finally, our experimental results are also in good qualitative agreement with the theory proposed
 579 by Freire et al. (2016) for establishing a link between the emission flux at the surface and the
 580 vertical profile of particles concentration in the SBL. These authors hypothesized that the effect
 581 of atmospheric stability should be incorporated in the parameterization of the eddy diffusivity of
 582 the particles. In their theory, the dependence to size is taken into account by the means of the
 583 trajectory-crossing effect, which is tantamount to considering that K_p decreases slowly with size,
 584 and they thus conclude from their idealized LES simulations that the effect of gravitational
 585 settling should be only secondary to that of stability. Our own results confirm the importance of
 586 atmospheric stability but also suggest that in natural conditions the dependence of this effect on
 587 particle size could be larger than considered in the current theories.

588 In order to quantify the effect of thermal instability on the size-distribution of the vertical
 589 flux, we examined the evolution with $\partial\theta/\partial z$ of its deviations from the neutral case. Practically,
 590 we calculated the ratio of the relative contributions of each size class in non-neutral and near-
 591 neutral ($-0.05 < \partial\theta/\partial z < 0.05^\circ\text{C m}^{-1}$) conditions and plotted this ratio as a function of $\partial\theta/\partial z$.
 592 Figure 9 illustrates the results for a fine ($0.49 \mu\text{m}$), an intermediate ($1.15 \mu\text{m}$), and a coarse
 593 ($3.65 \mu\text{m}$) size class. In spite of some scatter, a linear behavior is observed in the three cases. By
 594 definition of the ratio, the vertical intercept that represents its value in neutral conditions is equal
 595 to 1. The slope (hereinafter noted A) denotes the sensitivity of the proportion of one given size-
 596 class to thermal stratification. Consistent with the enrichment of the dust flux in fine particles in
 597 thermally unstable conditions, A is found to be negative for the fine class and positive for the
 598 coarse one. For particles with $D_{g,i}$ between 1.15 and $1.54 \mu\text{m}$, A is almost nil, which means that,
 599 in the vertical dust flux, the proportion of particles in this size range is independent of $\partial\theta/\partial z$.
 600 The increase of A with the diameter of the particles (Fig. 9b) can be quite well (slope=0.99,
 601 $b=0.002$, $R^2=0.99$, $n=11$) represented by the following expression:

$$602 \quad A = A_{lim}(1 - B \exp(-k D_p)) \quad (9)$$

603 The values of B and k are 4.56 (unit-less) and $1.11 (\mu\text{m}^{-1})$, respectively. A_{lim} is the limit
 604 towards which A tends above a diameter of approximately $4 \mu\text{m}$. Its numerical value is 1.54 .
 605 Note that the fact that the dependence of A to size is notable only below $4 \mu\text{m}$ and larger for the
 606 smallest size-classes, confirms again that the very fine particles are the most sensitive to the
 607 modification of the turbulence structure resulting from a moderately unstable thermal
 608 stratification.



609

610 **Figure 9.** (a) Influence of thermal instability ($\partial\theta/\partial z$) on the proportions of particles of three
 611 different size-classes (0.49, 1.15, and 3.65 μm) in the vertical flux (the reference is the neutral
 612 case), and (b) evolution with particle size of the slopes of the linear trends obtained in each size-
 613 class (blue squares and the associated standard deviation in black). In the latter case, the solid red
 614 line is the mathematical adjustment (Eq. 9) to the measurements.

4.3 Implications for the modeling of the vertical flux.

Independently of the physical scheme used to estimate them, dust emission models usually assume that the flux of particles released from the surface (F_{emiss}) can be characterized by its intensity and relative size-distribution and that the latter is either constant (Kok, 2011b) or varying with wind speed (Alfaro & Gomes, 2001; Shao, 2001) but the size distribution of the vertical flux is simply derived from that of F_{emiss} . The results of our study emphasizes the importance of taking into account the thermal stratification of the surface layer in this derivation. Practically, this could be achieved by using transfer functions similar to that of Equation (9) allowing to express the size-distribution of the vertical flux in unstable conditions to that of neutral conditions.

5 Summary and conclusions

The vertical dust flux was measured during 8 wind erosion events over a flat sandy field in southern Tunisia. This allowed documenting the impact of a large variety of dynamical conditions on the magnitude of the number vertical flux and its size-distribution. During the first events of the WIND-O-V's 2017 experimental campaign, the wind strength tended to be globally larger and the thermal stratification of the SBL closer to thermal neutrality than later on. These differences had a direct impact on the magnitude and number size-distribution of the vertical dust flux measured by the gradient method between 2.04 and 4.10 m above the surface. The size-distribution was relatively richer in supermicron particles during the thermally neutral periods of the events, whereas the proportion of submicron ones increased in unstable conditions. Note that the discussion provided as supplementary material to this paper shows that the relatively short fetch of our experimental plot can be ruled out as a primary explanation for the enrichment in fine particles. Therefore, our results are in contradiction with two of the assumptions of the gradient method, namely 1) that the turbulent eddy coefficient of the particles smaller than 10 μm should be the same as that of the transport of the air-mass momentum, and therefore independent of the particles size, and 2) that the stability functions commonly used to account for the thermal stratification of the surface layer should be directly applicable to the vertical transport of dust particles.

Fundamentally, this means that the size-distributions determined from in-situ measurements performed at a height of a few meters do not necessarily reflect directly the actual distribution of the particles produced by sandblasting at the surface. According to the results of this study, only the size-distributions measured in, or very close to, neutral conditions are expected to approach this size-distribution. This could also explain that the size-distributions measured in this study seem to be controlled more by thermal instability than by u_* and that the effect of the latter parameter, though evidenced in wind-tunnel experiments (Alfaro et al., 1997, 1998), appears as one of second order magnitude in natural conditions.

Finally, our results stress the importance of documenting carefully not only the evolution of u_* but also those of the vertical thermal gradient when interpreting the variations of the size-distribution of the vertical flux measured over an eroding field. In our case, it has been possible to find an empirical correlation between this size-distribution and the thermal structure of the SBL but it is not certain whether similar corrections could be applied to any type of eroding surface. More experimental studies performed in other source areas are clearly necessary to

658 address this question and over plots with a longer fetch would clearly be necessary to address
 659 this question. New theoretical developments would also be urgently needed to explain the larger
 660 than expected dependence to size of the vertical entrainment of fine dust particles in unstable
 661 SBL.

662

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