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Mixing of dust aerosols into a mesoscale convective system. Generation, filtering and possible feedbacks on ice anvils.<sup>☆</sup>

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**Abstract**

During the second Specific Observing Period (SOP) of the African Monsoon Multidisciplinary Analyses (AMMA) campaign, several intense Mesoscale Con-

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vective Systems (MCS) developed over Niger. An examination of a particular convective storm simulated with a mesoscale model near Banizoumbou, Niger, on 1 July, 2006, shows that this MCS generates a strong emission of dust particles at the leading edge of its density current. A fraction of these dust aerosols are uplifted by the convective core of the system and redistributed by aqueous processes. Aerosol impaction scavenging is the main process by which particles are deposited within the mesoscale convective system. However, small particles (smaller than  $1\mu m$ ) that are not efficiently scavenged, are able to reach the upper troposphere at a concentration of 6 particles per  $cm^3$ . This suggests that deep convection over semi-arid regions is able to create its own ice nuclei in high concentrations. This leads to the question: can deep convection over semi-arid regions affect particular ice properties such as ice anvil extension or induce possible feedbacks of dust on precipitation through ice sedimentation?

*Key words:* dust, aerosol scavenging, ice nuclei, gust front, Mesoscale Convective Systems

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## 1. Introduction

Mineral dust contributes significantly to the global radiative budget calculations through absorption and scattering of longwave and shortwave radiation (Houghton et al., 2001), and its indirect effect on cloud microphysics (IPCC, 2007), (Twomey, 1959), (Albrech, 1989), (Sandu et al., 2008)). At a local level, high dust concentrations are shown to impact the vertical structures of storms (Lohmann and Diehl, 2006) and local energy budgets (Grini et al., 2006). The dust diameters have a large range in size, from below

9 0.2 to 40  $\mu m$ , implying that different sink mechanisms need to be correctly  
10 modeled, such as sedimentation, wet and dry deposition. Dry deposition  
11 includes turbulent transfer to the surface and gravitational settling. Wet  
12 deposition includes nucleation scavenging and impaction scavenging, both  
13 involving complex microphysical interactions between aerosols and hydrometeors.  
14 Nucleation scavenging describes the activation of aerosols into cloud  
15 droplets and ice crystals and subsequent growth to precipitating hydrometeors.  
16 Impaction scavenging is the collection of the aerosols by cloud droplets,  
17 ice crystals and precipitating hydrometeors through Brownian motion, in-  
18 terception, inertial impaction, thermophoresis, diffusiophoresis, airflow  
19 turbulence and electrostatic attraction in and below the cloud layer (An-  
20 dronache, 2003). Usually, impaction scavenging by iced hydrometeors are  
21 not considered in numerical models because of its low effect compared to  
22 collection by liquid particles. Theoretically, impaction scavenging is usu-  
23 ally split into two processes: in-cloud impaction scavenging that treats the  
24 interactions between cloud droplets and raindrops with interstitial aerosol  
25 particles, and below cloud scavenging that concerns the collection of aerosol  
26 particles by falling raindrops below the cloud base. The relative importance  
27 of in-cloud scavenging processes (nucleation and impaction scavenging), also  
28 called washout, and below cloud scavenging, also called rainout, depends  
29 on meteorological conditions and on the properties of aerosol particles (size  
30 distribution and chemical composition) as well as on the stage of cloud de-  
31 velopment. Global models have estimated that wet deposition only accounts  
32 for 10% of the total dust loss globally (Tegen and Lacis, 1996; Ginoux et al.,  
33 2001). Locally, wet deposition can be expected to play a larger role in dust

34 removal processes, especially in areas of intense convection, as observed dur-  
35 ing the monsoon season in West Africa.

36

37 The data acquired in the framework of the 2006 African Monsoon Mul-  
38 tidisciplinary Analyses (AMMA) field campaign (Redelsperger et al., 2006)  
39 provides an excellent opportunity to analyze the processes of dust gener-  
40 ation and deposition associated with convection. Within semi-arid regions,  
41 the strong winds associated with gust fronts of Mesoscale Convective Systems  
42 (MCSs) generate dust and aerosol lofting. Some recent studies have detailed  
43 the dust formation in the leading edge of the density current observed in the  
44 intertropical discontinuity region (Flamant et al., 2007, 2009; Marsham et al.,  
45 2008). The dust generation associated with the 1 July, 2006 MCS (SOP 2  
46 of the AMMA campaign) was modeled near the Niamey and Banizoumbu  
47 region (Niger) and compared to observations by Crumeyrolle et al. (2008).  
48 The present work is an extension of the Crumeyrolle et al. (2008) study with  
49 a special focus on the generation and the vertical transport of dust in the  
50 convective core of a MCS in relation to precipitation. In particular, several  
51 aspects of a MCS over a semi-arid region that can cause feedbacks of dust  
52 on ice in the upper troposphere are emphasized.

## 53 **2. Model Description**

### 54 *2.1. General parameterizations*

55 The MesoNH model has been jointly developed by the CNRM (Météo-  
56 France and Centre National de la Recherche Scientifique) and Laboratoire  
57 d’Aérodologie (Centre National de la Recherche Scientifique and Université de

58 Toulouse) (Lafore et al., 1998). MesoNH simulates small scale atmospheric  
59 circulation (horizontal resolution of a few meters) to synoptic scale (hori-  
60 zontal resolution of several tens of kilometers) and can be run in a two-way  
61 nested mode involving up to 8 nesting stages. Parameterizations have been  
62 introduced for convection (Bechtold et al., 2001), cloud microphysics (Pinty  
63 and Jabouille, 1998; Cohard and Pinty, 2000), turbulence (Bougeault and  
64 Lacarrere, 1989), biosphere-atmosphere thermodynamic exchanges (ISBA)  
65 (Noilhan and Mahfouf, 1996), urban-atmosphere interactions (Masson, 2000),  
66 lightning processes (Barthe et al., 2005), gaseous chemistry (Suhre et al.,  
67 1998; Tulet et al., 2003) and aerosols chemistry (Tulet et al., 2006).

## 68 *2.2. Mineral dust parameterization*

69 Mineral dust emissions are parameterized following Grini et al. (2006). In  
70 this parameterization, the three lognormal modes are generated and trans-  
71 ported by the ORILAM lognormal aerosol scheme (Tulet et al., 2005). Re-  
72 garding emission processes, dust aerosols are mobilized using the Dust En-  
73 trainment and Deposition model (DEAD) (Zender et al., 2003) which calcu-  
74 lates dust fluxes from wind friction speeds. The physical basis of the model  
75 is taken from Marticorena and Bergametti (1995) in which dust fluxes are  
76 calculated as a function of saltation and sandblasting processes. Here, the  
77 emission of dust aerosols is calculated directly from ISBA surface parameters,  
78 and then sent to the atmosphere consistently with the fluxes of momentum,  
79 energy and humidity. The initial dust size distribution contains three modes  
80 with median radii of 0.32, 1.73 and 4.33  $\mu\text{m}$  and standard deviations of 1.7,  
81 1.6 and 1.5, respectively as defined by Alfaro and Gomes (2001).

82 *2.3. The ICE3 cloud microphysics scheme*

83 The study uses the ICE3 cloud microphysics scheme described by Pinty  
84 and Jabouille (1998). This scheme follows the approach of Lin et al. (1983)  
85 in that a three-class ice parameterization is coupled to a Kessler’s scheme  
86 for warm processes. It is a bulk single moment scheme that predicts the  
87 evolution of the mixing ratios of six water species (vapor, cloud droplets  
88 and raindrops, pristine ice, snow and graupel). Pristine ice crystal are here  
89 assumed to be plates. The precipitation of water drops and ice crystals is  
90 parameterized according to Caniaux et al. (1994). The size distribution of the  
91 hydrometeors is assumed to follow a generalized  $\gamma$ -law function in normalized  
92 form. Due to the use of a single moment bulk scheme, this study does not  
93 consider the cloud condensation nuclei and ice nuclei activation. There is  
94 no interaction between the dust particles and the microphysics within the  
95 model.

96 *2.4. Impaction scavenging and aerosol-cloud interactions*

97 The impaction scavenging of aerosols is calculated in MesoNH based upon  
98 first order principals. In- and below-cloud impaction scavenging by cloud  
99 droplets and raindrops uses a kinetic approach to calculate the aerosol mass  
100 transfer as:

$$\frac{dM_p}{dt} = -\Lambda_M M_p \quad (1)$$

101 where  $dM_p/dt$  represents the aerosol dry mass transfer in the aqueous phase,  
102  $M_p$  the aerosol dry mass and  $\Lambda_M$  the path normalized scavenging coefficient in  
103  $s^{-1}$ . For impaction scavenging by cloud droplets, the main process to consider  
104 is the Brownian motion of dry aerosols and cloud droplets (Pruppacher and

105 Klett, 2000) leading to the normalized scavenging coefficient determined by  
 106 the semi-empirical formulation as:

$$\Lambda_M M_p = \frac{1.35 LWC D_p}{r_{cloud}^2} \quad (2)$$

107 where  $LWC$  is the cloud liquid water content in  $g.cm^{-3}$ ,  $D_p$  is the diffusivity  
 108 of the particle in  $m^2.s^{-1}$  and  $r_{cloud}$  the cloud droplet radius in  $m$ .

109 The impaction scavenging by raindrops depends mainly on Brownian motion,  
 110 interception, and inertial impaction following a formula originally described  
 111 by Slinn (1983):

$$\Lambda_M M_p = \frac{3}{2} \frac{E}{r_{rain}} \cdot F_{rain} \quad (3)$$

112 where  $E$  is the collection efficiency fully described in Seinfeld and Pandis  
 113 (1997); Tost et al. (2006),  $r_{rain}$  the radius of the rain droplets in  $mm$  and  
 114  $F_{rain}$  the effective precipitation flux in  $kg.m^{-2}.s^{-1}$ .

115 Within this impaction scavenging scheme, the efficiency is calculated for  
 116 three types of collection. Small particles are collected efficiently by raindrops  
 117 and cloud droplets through Brownian diffusion, but the collection efficiency  
 118 decreases with increasing particle size. Inertial impaction by raindrops is  
 119 important for large particles, with collection efficiencies approaching one for  
 120 particles with diameter greater than  $20 \mu m$ . Inside the cloud, impaction scav-  
 121 enging by cloud droplets is less efficient for particles with diameters from  $0.2$   
 122 to  $2.0 \mu m$ . Indeed interception by raindrops is difficult since particles follow  
 123 the streamlines of air around the falling droplets. The in-cloud mass aerosol  
 124 transfer into rain droplets by autoconversion and accretion processes have  
 125 been introduced as described by Pinty and Jabouille (1998). The sedimen-  
 126 tation of aerosol mass included in raindrops is solved using a time splitting

127 technique with an upstream differencing scheme of the vertical flux as:

$$P_{asr} = \frac{m_{aero}}{m_{rain}\rho} \frac{d}{dz} (V_r \cdot \rho \cdot r_{rain}) \quad (4)$$

128 where  $P_{asr}$  is the raindrops aerosol mass sedimentation rate,  $m_{aero}$  the aerosol  
129 mass included in raindrops in  $kg.kg^{-1}$ ,  $m_{rain}$  the rain water mass in  $kg.kg^{-1}$ ,  
130  $\rho$  the air density and  $V_r$  the raindrop sedimentation velocity in  $m.s^{-1}$ .

131 As for the microphysical scheme, only mass transfer between aerosols and  
132 warm cloud processes have been considered here. As a consequence the  
133 model follows the aerosol mass of each lognormal mode during its evolution  
134 through the warm ICE3 processes. Indeed the aerosol scheme is limited to one  
135 moment (mass) during its warm processes exchange. This limitation involves  
136 two assumptions: (1) Mass transfer does not change the size distribution of  
137 the aerosol modes, and (2) the rerelease of aerosols into the air due to rain  
138 evaporation is proportional to the mass of water evaporated (Chin et al.,  
139 2000). As a consequence, the mean radii and standard deviation of the  
140 raindrops aerosol modes are identical to the dry aerosol size distribution,  
141 implying that neither coagulation nor chemical transformation occurred in  
142 the cloud droplets are considered in the model. The second assumption  
143 however, is likely to overestimate the release of aerosols due to evaporation  
144 as some evaporation of the rain results in smaller raindrops that still contain  
145 the aerosols.

#### 146 *2.5. Model configuration*

147 The simulation begins at 00 UTC on June 29, 2006, and ends at 00  
148 UTC on July 2, 2006. Three two-way nested domains are used. The large  
149 domain at 36 km resolution (3.1°S - 31.7°N; 25.64°W - 35.64°E) gives a

150 large scale synoptic view of west Africa. The first embedded domain (12 km  
151 resolution) is centered over northwest Nigeria and covers a large part of the  
152 AMMA campaign area (4.3°N - 17.6°N; 4.19°W - 16.24°E). The embedded  
153 smallest domain at 3 km resolution (11.80°S - 16.82°N; 0.10°W - 8.31°E)  
154 gives a fine scale view of Niger. For the two larger domains, the Bechtold  
155 et al. (2001) convection scheme is used, whereas deep convection is assumed  
156 to be explicitly resolved at 3 km resolution. Only the smallest domain is  
157 examined in this study. The vertical grid is composed of 60 stretched vertical  
158 levels reaching the altitude of 34,000 meters above ground level (m agl); 30  
159 levels are located in the boundary layer between the surface and 2,000 m agl.  
160 Initialization and lateral boundary conditions of the large domain are taken  
161 from the ECMWF analysis.

162 Two types of simulations have been performed. The complete simulation  
163 (SCAV) uses the dust scavenging scheme whereas the NOSCAV simulation  
164 do not include wet deposition for dust. Because there is no connection of  
165 dusts with dynamics and microphysics, these two simulations give identical  
166 MCSs. Actually, the difference between these two simulations shows the  
167 impact of precipitation on the dust distribution.

### 168 **3. Mesoscale convection and dust generation**

169 This section focus on the vertical structure of a MCS observed and simu-  
170 lated east of Banizoumbou the 1<sup>st</sup> of July at midnight. Figures 1.a, 1.c and  
171 1.e display the MSG brightness temperature (at 10.8  $\mu m$ ) on 1 July 2006 at  
172 18 UTC, 2 July at 00 UTC and 2 July at 02 UTC, respectively (Chaboureau  
173 and Pinty, 2006). The MSG satellite images show the passage of several con-

174 vective systems over the south-east of Niger: most of the MCSs are triggered  
175 at the eastern part of the simulated domain (Fig. 1.a) and propagate to the  
176 west over the Niamey and Banizoumbou region. The two most intense sys-  
177 tems are observed north of and over Banizoumbou between 00 UTC and 02  
178 UTC (Fig. 1.c and 1.e). The northern MCS is triggered  $3^\circ$  north-east of Ban-  
179 izoumbou (Fig. 1.a) and propagates to Niamey and Burkina Faso (Fig. 1.c  
180 and 1.e). The second system is triggered at  $2.5^\circ$  east of Banizoumbou (Fig.  
181 1.a). It moves slowly (Fig. 1.c) and disappears over Banizoumbou (Fig. 1.e).

182

183 Figures 1.b, 1.d and 1.f give the cloud top height (dashed) and the in-  
184 stantaneous precipitation (isolines) simulated by MesoNH on 1 July 2006 at  
185 16 UTC, 18 UTC and 20 UTC, respectively. As observed previously on the  
186 satellite images, some convective systems are simulated over the south-east  
187 of Niger where the two main systems have reached the Niamey region. The  
188 system north is triggered on 1 July 2006 at 16 UTC, at  $1^\circ$  north-east of  
189 Banizoumbou (Fig. 1.b). This difference in the initiation location will cause  
190 a 5 hours time lag in the passage of the systems over the Niamey and Ban-  
191 izoumbou region. However it evolves similarly to observations, except for  
192 passing a bit too south of Niamey (Fig. 1.c, 1.d, 1.e and 1.f). The southern  
193 system begins  $2.5^\circ$  east of Banizoumbou at the same location as observed  
194 on the MSG images, but one hour earlier than observed (Fig. 1.a and 1.b).  
195 Furthermore, its propagation is quite different from the observed MCS. At  
196 20 UTC, the system is simulated  $1^\circ$  southeast of Banizoumbou (Fig.1.d) and  
197 instead of disappearing on Banizoumbou (Fig. 1.e), the system continues its  
198 propagating to the border of Burkina-Fasso (Fig. 1.f).

199 The overshoot of the modeled MCS reaches 19,000 m agl in the lower strato-  
200 sphere and the top of the anvil is simulated at 16,000 m agl at the altitude of  
201 the tropopause. The area of detrainment is spread over a large area located  
202 at the tropopause. By comparison between the MSG images and the simula-  
203 tion, one can observe that differences exists in the location of triggering and  
204 in the stage of the cycle life. However, these differences should not strongly  
205 affect the focus of this article; the major features of these MCS are realistic  
206 enough and characteristic of convection systems over west Africa.

207 The instantaneous simulated precipitations (isolines) have been superim-  
208 posed to the altitude of the cloud top (isolines on Fig. 1.b, 1.d and 1.f).  
209 The maximum of observed precipitation gives the cloud zone where the con-  
210 vection is strong. The cell simulated at the south-east of Banizoumbou at 20  
211 UTC (Fig 1.d) is particularly active with the precipitation rate reaching 50  
212  $mm.h^{-1}$  at the surface.

213 MCS downdrafts create a gust front where surface winds exceed  $15 m.s^{-1}$ .  
214 This gust front can be seen on Figure 2 by surface winds vector divergence  
215 in the front of each MCS. The associated surface winds are greater than the  
216 wind speed threshold of  $6.5 m.s^{-1}$  for soil erosion determined by Chomette  
217 et al. (1999) over Sahelian-Saharan regions. However, around the MCSs, the  
218 surface winds are low (less that  $5 m.s^{-1}$ ) and they cannot produce dust emis-  
219 sion (Fig. 2). As a consequence, only in the MCS gust front, winds are for  
220 the most part strong enough to move soil particles by saltation and generate  
221 high dust concentration at the surface as simulated by the model ahead and  
222 west of the MCS (Fig.2). More than  $10,000 \mu g.m^{-3}$  of dust are simulated  
223 at the surface east of Banizoumbou and  $5,000 \mu g.m^{-3}$  over Niamey. Three

224 others gust fronts with high dust concentration are simulated in the northern  
225 part of the domain. They are created by the small convective cells simulated  
226 at 19 UTC (not shown) and 20 UTC (Fig. 1.d). Note that the intense dust  
227 plume formed in this northern area, is mainly due to the soil characteristics  
228 and to the absence of any vegetation.

229 Simulation validation through observation data was detailed by Crumeyrolle  
230 et al. (2008): it was shown that the increase of the surface wind speed, the  
231 decrease of surface pressure and the measured precipitation at the arrival of  
232 the MCS are correctly reproduced over Niamey. In addition, the dust mass  
233 concentration profile in the lower troposphere simulated over the Niamey  
234 region before and after the MCS passage are closed to the ATR42 aircraft  
235 observations. The next section of the paper focuses on a particular cell which  
236 was simulated at the east of Banizoumbou (Fig.2). This cell is significant  
237 in the high concentration of dust contained in the gust front (about 3,000  
238  $\mu g.m^{-3}$ ) and the associated precipitation rate is important ( $50 mm.h^{-1}$ ).  
239 Furthermore, this cell generates considerable levels of detrainment in the  
240 upper troposphere as shown in Figure 1.d.

#### 241 **4. Dust formation and redistribution in the convective cloud**

242 Figure 3 gives the vertical cross section of the total condensed water by  
243 the solid line delineated on the Figure 2 (the left corner of the cross section  
244 corresponds to the south west point of the solid line). The simulated verti-  
245 cal velocity reached  $20 m.s^{-1}$  between 9,000 m agl and 15,000 m agl, (black  
246 isolines of Fig. 3), that represents the convective core of the system. More  
247 than  $2.5 g.kg^{-1}$  of condensed water is simulated within this convective core.

248 Detrainment of the system is modeled at the tropopause located at 16,000  
249 m agl. The convective overshoot reaches 19,000 m agl and transports to the  
250 lower stratosphere more than  $1 \text{ g.kg}^{-1}$  of total condensed water. Above 6,000  
251 m agl, rainfall appears with a maximum of  $4 \text{ g.kg}^{-1}$  at 2,000 m agl under  
252 the convective core.

253 This intense precipitation (evaporation and drag of hydrometeors) plays an  
254 important role in the downdrafts and the cold pool formation at the sur-  
255 face. This cold pool is characterized by low potential temperature (298 K),  
256 7 degrees lower than its environment and a negative buoyant air (Fig. 4.a).  
257 However, at the leading edge of the gust front identified by the surface winds  
258 convergence (streamlines in Figure 4.a), the vertical gradient of potential  
259 temperature is null between the surface and 2,000 m agl, showing that the  
260 air of this frontal region is mixed in this layer. Within the cold pool and the  
261 frontal zone, the turbulent kinetic energy modeled is greater than  $2 \text{ m}^2.\text{s}^{-2}$   
262 (shaded area on Figure 4.a). Indeed the air from the gust front is mixed by  
263 turbulence in the frontal zone and the winds convergence has forced the air  
264 to rise dynamically above the gust front.

265 Figure 4.b gives the vertical cross section of the dust concentration (SCAV  
266 simulation). On the right side of this Figure, the Sahelian boundary mixing  
267 layer appears with about  $200 \mu\text{g.m}^{-3}$  of dust. High surface winds associated  
268 with the gust front (more than  $12 \text{ m.s}^{-1}$  at 10 m agl) generate sandblasting  
269 and a saltation flux of dust particles at the surface (Figure 4.b). Within  
270 the cold pool,  $1,500 \mu\text{g.m}^{-3}$  of dust is simulated in the complete simula-  
271 tion (SCAV) of Figure 4.b. At the leading edge of the gust front, dust is  
272 transported upward in the ascending current and entrained into the cloud.

273 Knippertz et al. (2009) have also found similar results where strong emission  
274 of dust have been engendered by density current associated to moist convec-  
275 tion over the Atlas Mountains. This entrainment of dusty air in the cloud is  
276 also visible by the curvature of the streamlines of Figure 4.a. Precipitations  
277 (marked by the shaded area of Figure 4.b) scavenges most of the dust mass  
278 concentration entrained into the cloud. They also contributes to decreases  
279 significantly the dust concentration simulated in the cold pool: the maximum  
280 of rain mixing ratio at the surface ( $2 \text{ g.kg}^{-1}$ ) corresponds to a local minimum  
281 of dust concentration. Moreover, beyond this maximum, a second band of  
282 precipitation ( $1 \text{ g.kg}^{-1}$ ) corresponding to a cumulus simulated in front of the  
283 MCS, also contribute to limit the dust concentration at the east part of the  
284 gust front. Despite of these precipitations, in the SCAV simulation, 500 and  
285  $100 \mu\text{g.m}^{-3}$  of these fresh particles produced in the gust front remains at  
286 1,000 m agl and 3,000 m agl, respectively. Dust concentrations decreases to  
287  $30 \mu\text{g.m}^{-3}$  at 6,000 m agl (not shown). Flamant et al. (2007) and BouKaram  
288 et al. (2008) showed that the frontal zones like the inter-tropical discontinu-  
289 ity are favorable to dust lifting through turbulence. At a more local scale  
290 and over desert, gust fronts associated to single convection cell, also generate  
291 dust particles and transport aerosols above the boundary layer.  
292 Without any dust scavenging (NOSCAV simulation), the modeled dust con-  
293 centration is much larger, reaching  $3000 \mu\text{g.m}^{-3}$  in the gust front (Fig. 5).  
294 Convection lifts more than  $300 \mu\text{g.m}^{-3}$ ,  $100 \mu\text{g.m}^{-3}$  and  $50 \mu\text{g.m}^{-3}$  of dust  
295 formed in the gust front to 5,000 m agl, 10,000 m agl, and 16,000 m agl,  
296 respectively. At the tropopause, these particles are detrained far from the  
297 convective core in the anvil.

298 Figure 6 shows the differences between the dust mass concentration of the  
299 two simulations NOSCAV and SCAV. In both simulations, the dynamics and  
300 cloud microphysics are the same. In-cloud, the differences of dust mass con-  
301 centration between the two simulations (NOSCAV-SCAV) are on the same  
302 order of magnitude as the dust concentration simulated in the NOSCAV  
303 simulation, showing that the dust mass is close to zero in the cloud of the  
304 simulation SCAV. This implies that most of the dust mass concentration has  
305 been scavenged by raindrops. Indeed, with dust scavenging, less than 1500  
306  $\mu g.m^{-3}$  of dust is modeled in the cold pool and as little as 15  $\mu g.m^{-3}$  is able  
307 to reach 8,000 m agl. Furthermore, some notable differences are modeled in  
308 the rain and cloud evaporation zones at the rear part of the MCS. Some of the  
309 dust particles collected by cloud droplets and raindrops are re-released in the  
310 SCAV simulation in the evaporative zones where the precipitation does not  
311 reach the surface. It results in more than 200  $\mu g.m^{-3}$  of dust mass near the  
312 surface due to precipitation evaporation. This can be observed behind the  
313 MCS and during the dissipation stage (Figure 6). Note that these released  
314 dust particles have likely enhanced hygroscopic properties caused by soluble  
315 materials within the monsoon flux coating the mineral dust (Levin et al.,  
316 1996; Crumeyrolle et al., 2008). Indeed, precipitations serve contradicting  
317 purposes: it washes out most of the dust mass in the gust front, but it play  
318 also a major role in the production of fresh dust.

## 319 **5. Dust vertical distribution and aerosol size filtering**

320 The profile of mass and number concentration for the three dust modes  
321 are plotted in Figure 7 along the cross section of Figure 6 indicated by the

322 vertical dashed line. Note that this profile corresponds to the convective  
 323 region of the MCS except between 3,000 m agl and 8,000 m agl where the  
 324 profile is at the boundary of the cloud which is indicated by a decrease in  
 325 the plotted dust concentration. The vertical profile of the smaller mode with  
 326 the median radius of  $0.32 \mu m$  is the same for both SCAV and NOSCAV  
 327 simulations (green solid and dashed lines are superimposed on Figure 7). It  
 328 indicates that this mode is not affected by impaction scavenging since these  
 329 particles are too large to be collected by Brownian motion and too fine to  
 330 have significant inertial velocities (i.e. collection by inertial impaction). The  
 331 collection efficiency factor is less than 0.5 % for the smallest particles mode  
 332 whereas for the two larger ones (with median radii of  $1.73 \mu m$  and  $4.33 \mu m$ )  
 333 the efficiency factors are 30 and 99 % respectively. As a consequence, the  
 334 majority of the dust concentration of the two larger modes has been scav-  
 335 enged and 99 % of the mass of the smallest mode is preserved and is able  
 336 to be transported upward in the convective core. The vertical profile of Fig-  
 337 ure 7.a shows that the precipitation decreases the dust mass concentration  
 338 above 2,000 m agl from  $150 \mu g.m^{-3}$  ( $R_g = 1.73 \mu m$  and  $R_g = 4.33 \mu m$ ) to  
 339  $30 \mu g.m^{-3}$  ( $R_g = 1.73 \mu m$ ) and  $8 \mu g.m^{-3}$  ( $R_g = 4.33 \mu m$ ). Between 3,000 m  
 340 agl and 9,000 m agl, a local maximum of the two larger modes is observed in  
 341 the SCAV simulation and corresponds to a local minimum of the NOSCAV  
 342 simulation. As explained before, this area is at the boundary of the cloud and  
 343 this local maximum corresponds to evaporated rain and cloud droplets that  
 344 re-release some new dusts. In the upper troposphere above 15,000 m agl, the  
 345 mass concentration from the SCAV simulation reaches respectively  $6 \mu g.m^{-3}$   
 346 ( $R_g = 0.33 \mu m$ ),  $0.8 \mu g.m^{-3}$  ( $R_g = 1.73 \mu m$ ) and  $0.4 \mu g.m^{-3}$  ( $R_g = 4.33$

347  $\mu m$ ) whereas the total mass concentration modeled in the NOSCAV simula-  
348 tion is  $50 \mu g.m^{-3}$ . Indeed, within the SCAV simulation the majority of the  
349 mass transported near the tropopause is due to the first mode, representing  
350 80 % of the total mass, whereas this mode represents only 10 % of the total  
351 emitted mass.

352 Even though most of the aerosol mass has been scavenged, the number of  
353 small particles reaching higher altitudes still are significant within the SCAV  
354 simulation (Figure 7.b). The SCAV simulation shows that 20 particles per  
355  $cm^3$  have been transported to 7,000 m agl and more than 6 particles per  $cm^3$   
356 reach the tropopause (16,000 m agl). This represents 1 to 2 % of the number  
357 concentration modeled at the surface in the gust front whereas only 0.4 % of  
358 the mass concentration in the gust front reach the upper troposphere.

359 These results show that the MCS precipitations are two-fold. Firstly, rain-  
360 fall (drag of hydrometeors and evaporation) is one of the main mechanism  
361 that generates downdrafts and causes the formation of a gust front. When  
362 convection is located over a dust source region, dust particles are formed in  
363 the cold pool, ahead of the precipitation zone, where the soil is dry enough.  
364 The size distribution of the emitted dust depends on the wind friction at the  
365 surface and thus is also influenced by surface conditions and by the intensity  
366 of downdrafts. That is as the emitted dust modes are strongly dependent on  
367 the energy of impaction during the sandblasting (Alfaro and Gomes, 2001):  
368 the stronger the surface wind, the greater the respective concentration of the  
369 smallest particles. A part of these fresh aerosols, which are located at the  
370 leading edge of the cold pool, can be entrained in the cloud updraft. Sec-  
371 ondly, rainfall filters the size distribution of particles by preserving the first

372 mode and scavenging the two larger modes. Our model shows that a large  
373 number of small particles can be transported by convection. In this case  
374 study, more than 6 particles per  $cm^3$  reached the tropopause.

## 375 **6. Dust transport in the UTLS**

376 On July, 2 at 00 UTC (Figure 1), the simulation shows that the upper  
377 level of the MCS detrainment is modeled at 16,000 m agl at the tropopause.  
378 At this altitude, more than 6 dust particles per  $cm^3$  have been transported  
379 up to the convective core of each MCS (Fig. 8.a). High dust concentration  
380 up to  $0.2\ cm^{-3}$  are spreading over the main part of the south-west of Niger.  
381 The previously studied MCS gets the most intense plume dust concentra-  
382 tion: more than  $2\ cm^{-3}$  of dust particles have been modeled on a surface  
383 exceeding  $22,500\ km^2$ . Figure 8.b shows that in the lower stratosphere (at  
384 20,000 m agl) the overshoot of the most intense MCS can transport high  
385 concentration of dust particles (with maxima reaching over  $0.1\ cm^{-3}$ ). As  
386 for water vapor, it is known that convection is an important source of aerosol  
387 in the lower stratosphere. For the first time, IN concentration was measured  
388 during the CRYSTAL-FACE campaign in Florida (Prezzi et al., 2007) using  
389 a continuous flow diffusion chamber sampling residual particles remaining af-  
390 ter evaporation of cloud particles initially collected by a counterflow virtual  
391 impactor. These measurements were made onboard the citation aircraft in  
392 anvil of convective clouds at altitudes between 8 and 11 km corresponding  
393 to the upper troposphere in Florida. In this study, the range of measured  
394 IN concentration is 0.001 to  $1\ cm^{-3}$  whereas it is 0.2 to  $6\ cm^{-3}$  in the sim-  
395 ulated MCS in the upper troposphere over Africa. This suggests that the

396 order of magnitude of aerosol produced and transported by the MCS over  
397 African semi-arid regions are high and greater than over Florida (DeMott  
398 et al., 2003).

## 399 **7. Conclusion**

400 This study emphasizes several important aspects of dust emissions which  
401 can be formed in the gust front of a MCS over semi-arid region. In the sim-  
402 ulations, this process models dust concentrations greater than  $3000 \mu\text{g}\cdot\text{m}^{-3}$   
403 at the surface. Without wet dust scavenging (NOSCAV), about  $50 \mu\text{g}\cdot\text{m}^{-3}$   
404 of dust is transported in the convective core and reaches the tropopause,  
405 whereas less than  $1 \mu\text{g}\cdot\text{m}^{-3}$  is modeled in the complete simulation (SCAV).  
406 Indeed, precipitations filter the main part of the super-micronic dust mode  
407 and the sub-micronic mode (here at a median radius of  $0.32 \mu\text{m}$ ) is preserved.  
408 This mode represents 10 % of the emitted mass but 98 % of the aerosol  
409 number. A large number of dust particles reaches the tropopause, with a  
410 maximum of 6 particles per  $\text{cm}^3$  at 16,000 m agl. This number concentration  
411 is several orders of magnitude greater than ice nuclei observed over Florida  
412 during the CRYSTAL-FACE campaign where there is no local dust emissions  
413 comparable from a semi-arid region. Considering that mineral dust can serve  
414 as good ice nuclei (Kanji and Abbatt, 2006; Richardson et al., 2007), the dust  
415 generated by convection may influence the ice number concentration near the  
416 tropopause. It is assumed that if high ice nuclei concentrations are able to  
417 significantly decrease the supersaturation in the section of cloud containing  
418 ice, we can expect that dust can play an important role in the type of ice  
419 crystals (Nelson, 2001). It is also possible that the modification of the size

420 and type of ice crystals can also affect the lifetime of anvils and impact the  
421 convective precipitation (Gilmore et al., 2004b) through the ice sedimenta-  
422 tion velocity in the upper troposphere. Another study from Gilmore et al.  
423 (2004a) compared cloud resolving model simulations results using liquid-only  
424 and liquid and ice phase microphysics. They underlined the role of the ice  
425 phase in increasing precipitation production aloft and producing stronger  
426 downdrafts and greater low-level downward precipitation fluxes (and ground  
427 accumulations) compared to liquid-only simulations. In addition, the varia-  
428 tion of ice-forming nuclei concentration leads to changes in hail sizes, which  
429 are known to have a great impact on the dynamic, thermodynamic and pre-  
430 cipitation characteristics of the resulting storm (van den Heever and Cotton,  
431 2004).

432 Some studies on the effects of dust on the masses of various ice species within  
433 the anvils of convective storms during CRYSTAL-FACE dedicated to the  
434 Florida convection have been performed by van den Heever et al. (2006) and  
435 Carrio et al. (2007). These studies used the mesoscale model RAMS, which  
436 includes simple a parameterization for connecting the concentration of CCN  
437 and IFN and the nucleated droplets and ice crystals. These studies have  
438 shown that IFN concentrations have a greater impact on updraft strength  
439 during the mature and dissipating storm stages.

440 Moreover, a recent study based on satellite and direct aircraft measurement  
441 during NAMMA (a part of AMMA operated by NASA) supports the hypoth-  
442 esis that Saharan dust may have led to invigoration of rain bands associated  
443 with tropical cyclogenesis near West African coastline (Jenkins et al., 2008).  
444 Another recent study (Min et al., 2009), that uses some multi-platform and

445 multi-sensor observations, shows that the consequences of dust on a particu-  
446 lar MCS observed over the Gulf of Guinea were to shift the precipitation size  
447 spectrum from heavy rain to light rain. At this stage several questions arise  
448 on the particular properties of MCS formed over semi-arid regions: (1) Are  
449 anvil lifetimes and spatial dimensions affected by the dust aerosols generated  
450 by downdrafts? (2) Does a feedback exist between dust and precipitation  
451 by limiting the ice sedimentation in the upper troposphere? On the one  
452 hand observations alone cannot answer these questions due to the difficul-  
453 ties of de-aliasing the meteorological particularities of each MCS. On the  
454 other hand, mesoscale models need to improve their parameterization of ice  
455 microphysics processes to include a realistic scheme of ice activation before  
456 sensitivity studies on ice nuclei impacts can be performed. One goal of the  
457 AMMA field campaign was to improve the knowledge of the properties of  
458 the West African MCS. The goal of this study was to show that a MCS  
459 over a semi-arid region can create high concentration of ice nuclei in the up-  
460 per troposphere. A joint investigation using both observations and models  
461 can serve to further understanding how aerosols can impact significantly the  
462 microphysics and the dynamics of an MCS in semi-arid region.

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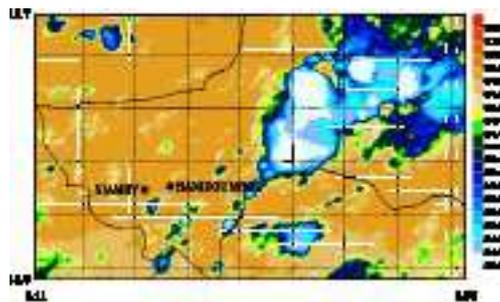
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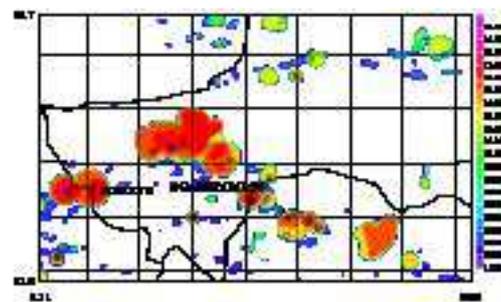
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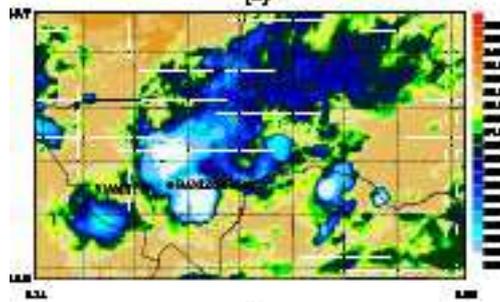
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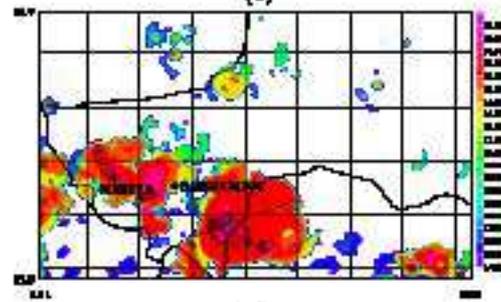
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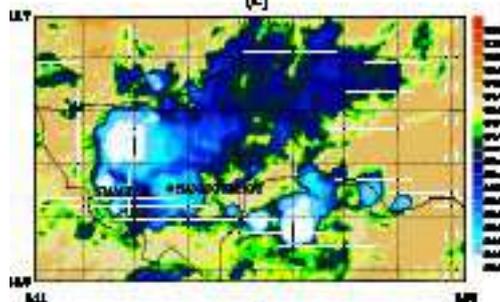
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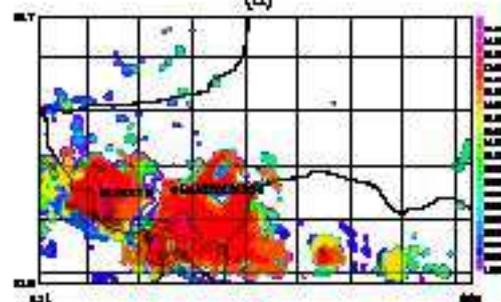
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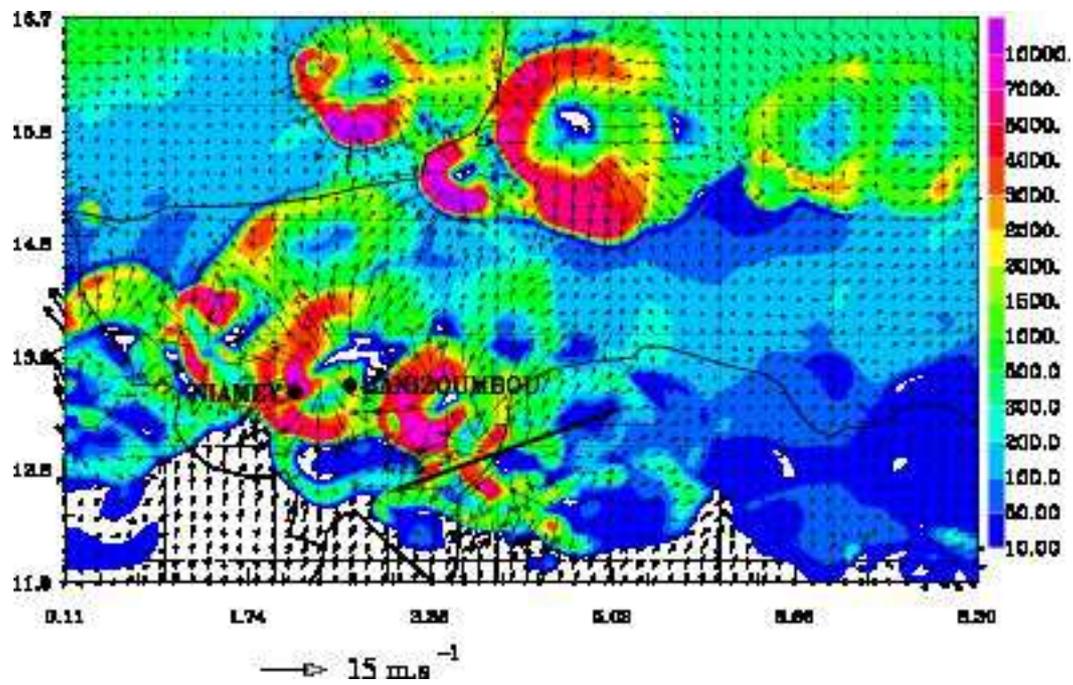
(d)

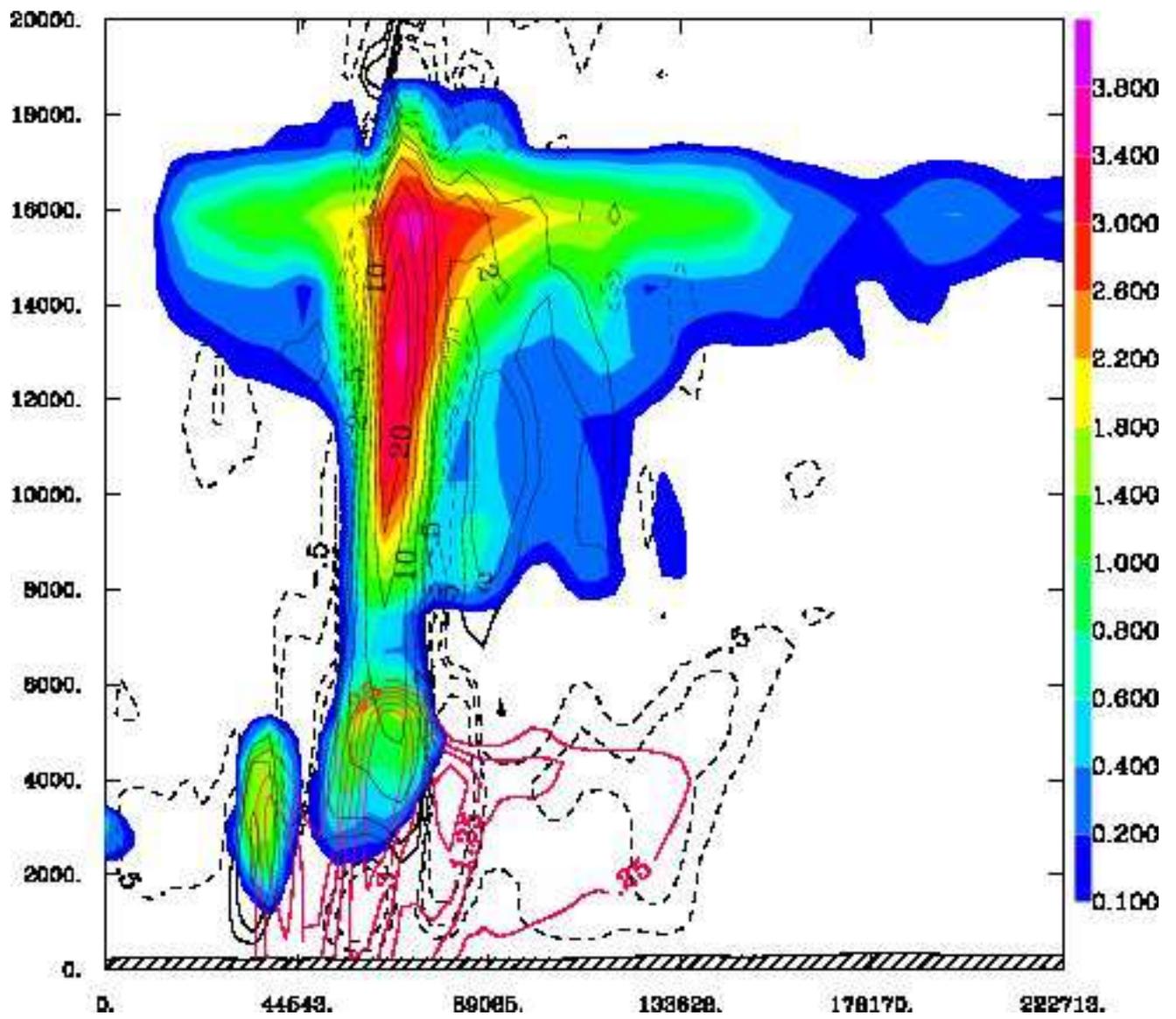


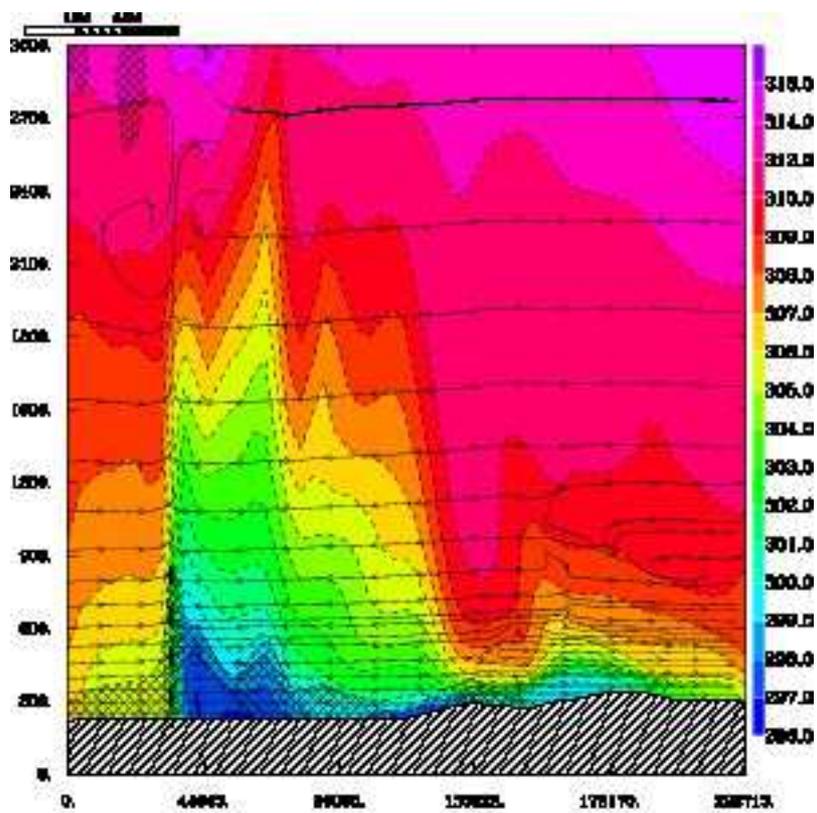
(e)



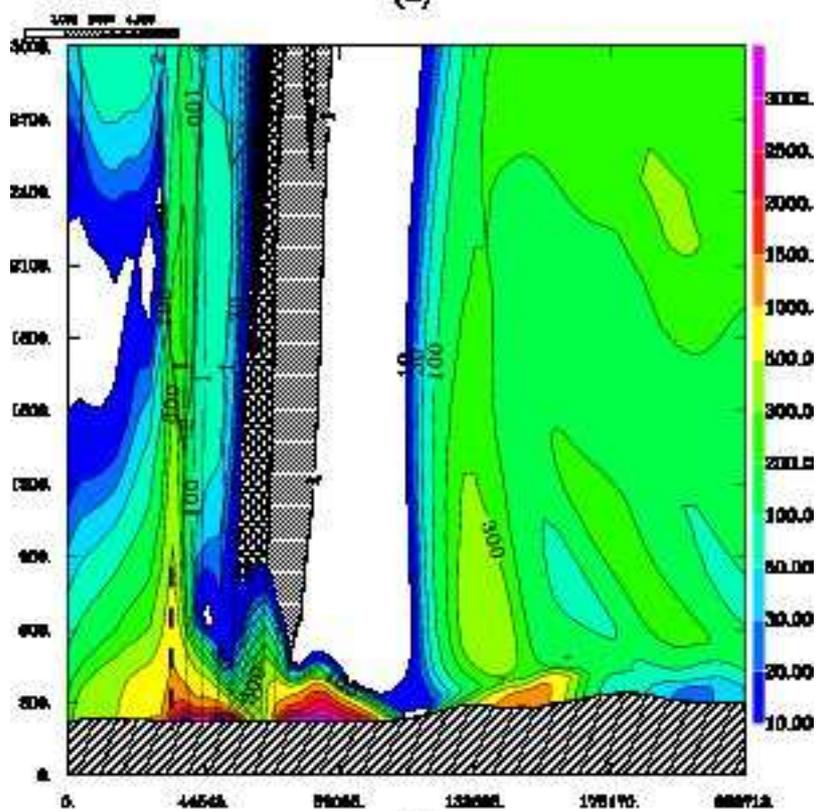
(f)



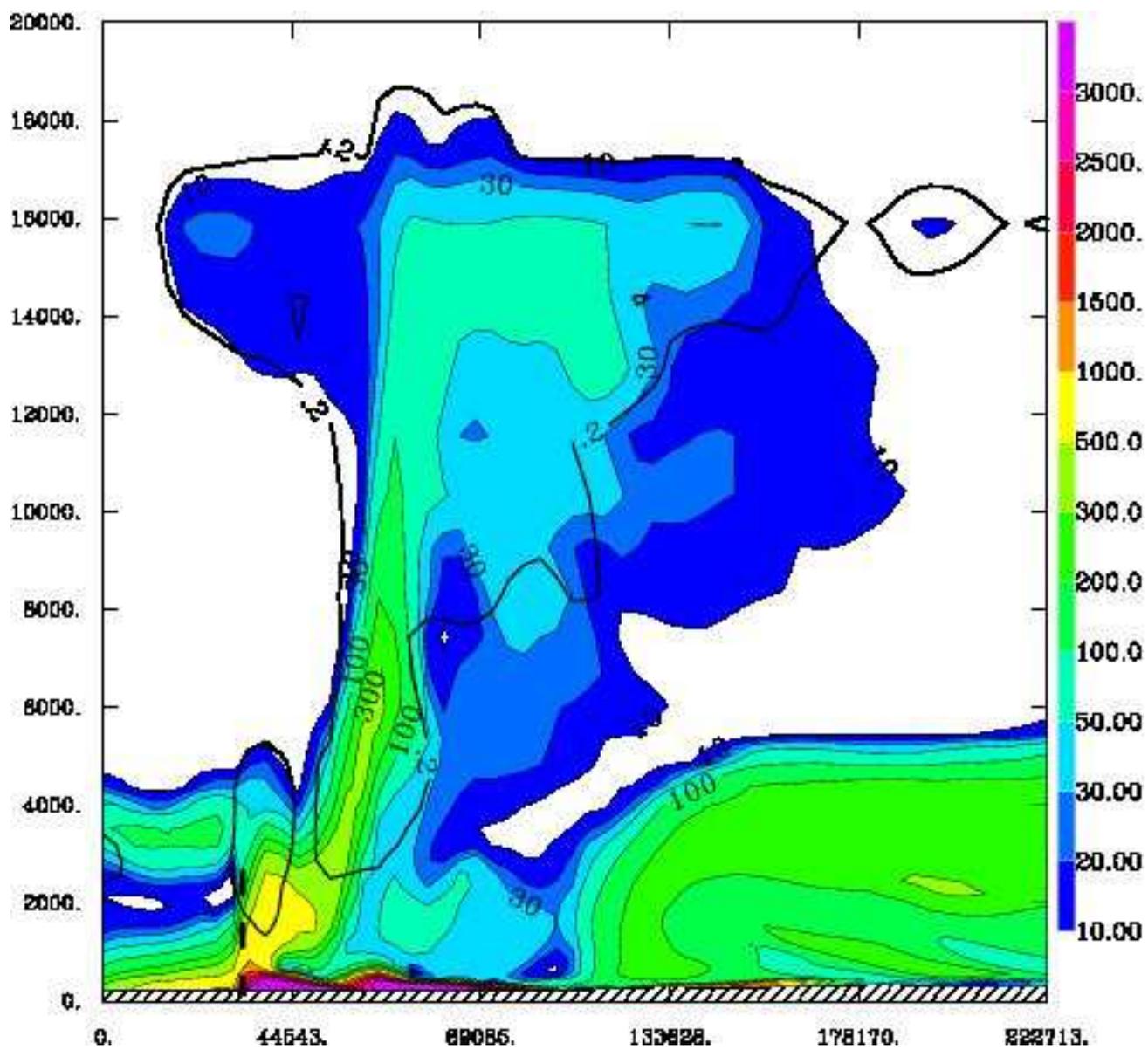


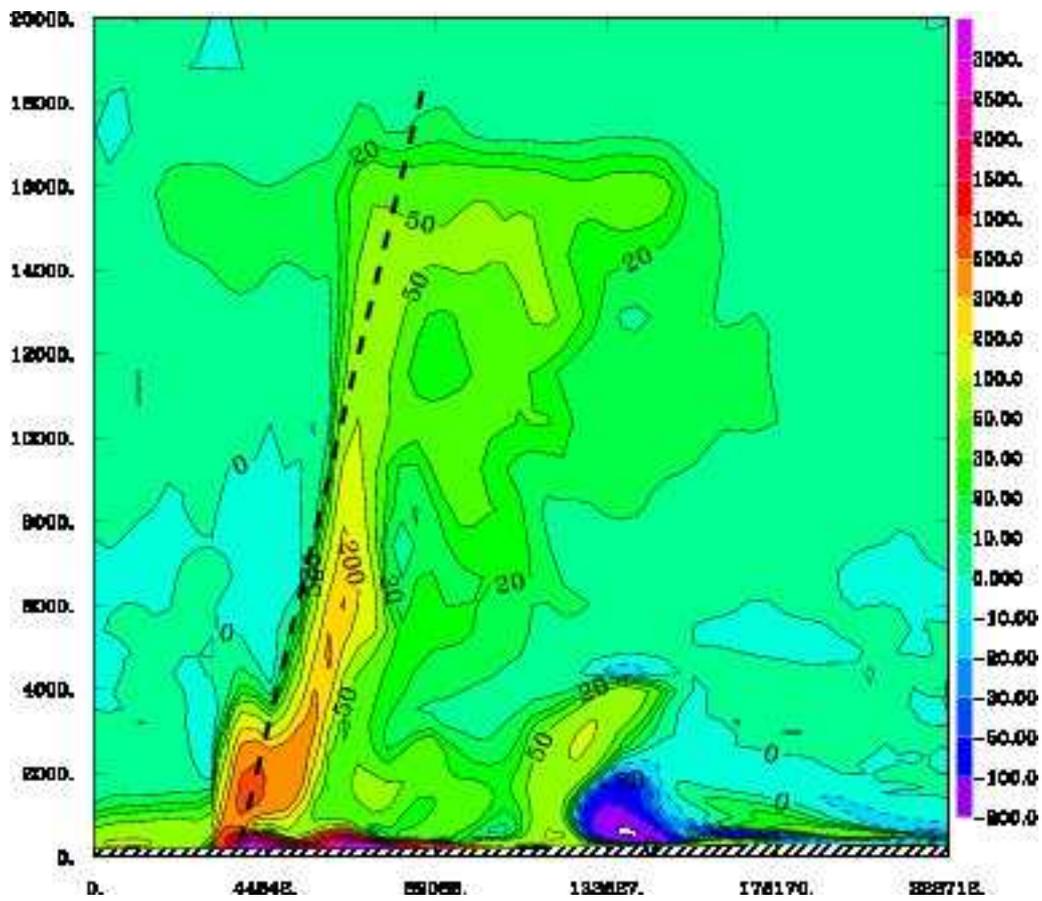


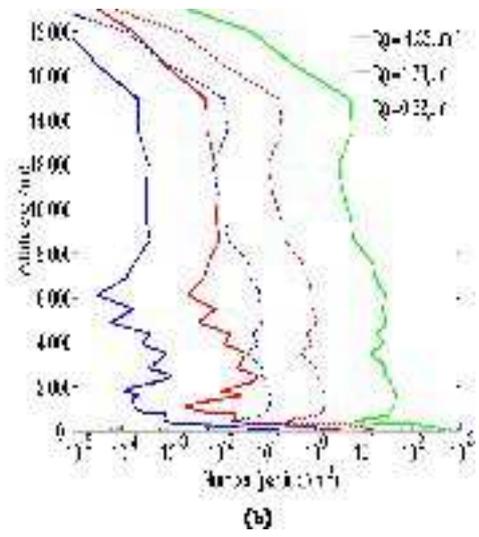
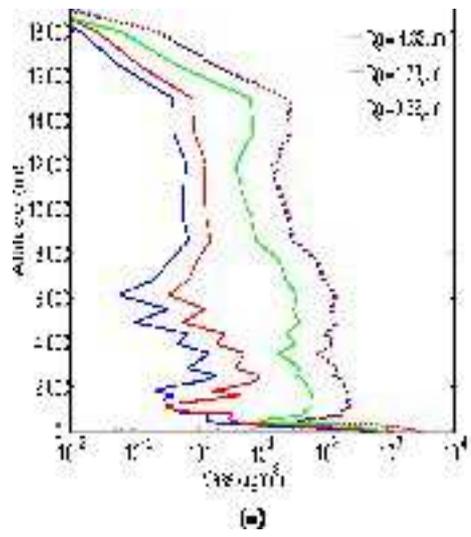
(a)

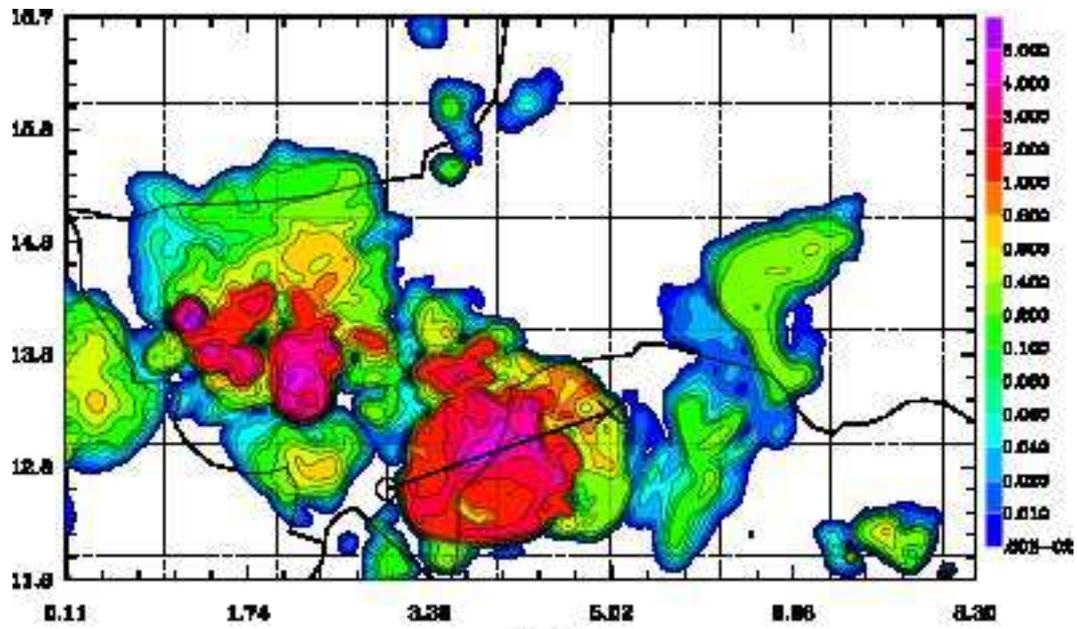


(b)

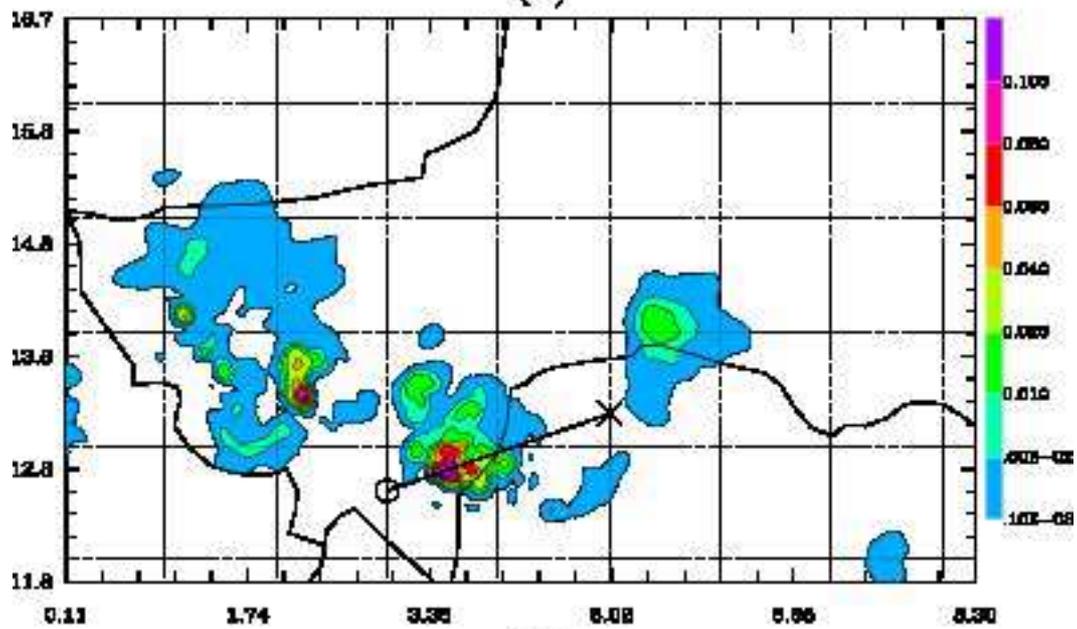








(a)



(b)