6.4 THE INFLUENCE OF AEROSOL PARTICLE NUMBER AND HYGROSCOPICITY ON THE EVOLUTION OF CONVECTIVE CLOUD SYSTEMS AND THEIR PRECIPITATION: A NUMERICAL STUDY BASED ON THE COPS OBSERVATIONS ON 12 AUGUST 2007

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ABSTRACT

The 3D cloud model DESCAM-3D with bin resolved microphysics for ice, water and aerosol particles is used to study the role of particles on the evolution of summertime mid-level convective clouds and the subsequent precipitation during the COPS field campaign which occurred at mid-latitude near the French/German border in summer 2007.

Using a 3D grid resolution of 250m, DESCAM-3D, is able to simulate well the dynamical, cloud and precipitation features of the convective cloud system observed during the afternoon of the 12th August.

In order to better understand the role of aerosol particles on cloud evolution and precipitation formation, several sensitivity studies were performed by modifying aerosol number concentration as well as their physico-chemical properties.

1. INTRODUCTION

In the general debate about climate change, the impact of air pollution on clouds, in particular the aerosol indirect effect (Twomey, 1974), is in the centre of numerous studies. Indeed, the cloud feedbacks remain the largest source of uncertainty in climate sensitivity estimates (Solomon, 2007). Thus, many studies have tried to understand the role of aerosol particles in the formation and evolution of clouds (see, e.g., Levin and Cotton, 2009) and the possible consequences of anthropogenic changes in these particle populations. Here, especially numerical models treating cloud microphysics in a bin resolved way can provide important insight, as cloud droplet nucleation is governed by supersaturation.

2. THE COPS CAMPAIGN

COPS (Convective and Orographically-induced Precipitation formation Study) was an international field campaign taking place at the French/German border (in the Upper Rhine Valley, the Black Forest and the Vosges Mountains, see Fig.1) during summer 2007.

Fig.1; The COPS measuring region

The aim of COPS was to study the orographic influence on precipitation and to improve cloud representation in forecast models (Wulfmeyer et al., 2008).

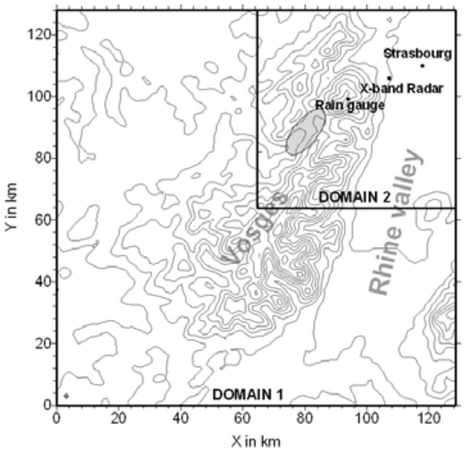

Fig.2: the model domain and the radar and rain gauge sites

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The LaMP X-band radar was deployed for 3 months within the Vosges French supersite Bischenberg, at Obernai near Strasbourg (Fig.2). It is designed to provide the precipitation field over a typical domain of a small catchment basin (about 20 km range) using a beam elevation of 5°. The major characteristic of the system is its high spatial and temporal resolution (60 m in range, 2° in azimuth and 30 s in time) (Van Baelen et al., 2009). Its observational area was also covered by the DLR polarimetric C-band Doppler radar POLDIRAD (Schroth et al., 1988) deployed 20 km north-west of Strasbourg. During COPS, POLDIRAD followed a dedicated observation strategy of which we are using the 2° elevation reflectivity scans with a resolution of 300 m and a range of 120 km repeated every 10 min.

3. THE MODEL

The 3D model with detailed (bin) microphysics used herein couples the 3D non-hydrostatic model of Clark and Hall (1991) with the Detailed Scavenging Model DESCAM (Leroy et al., 2009; Flossmann and Wobrock, 2010) for the microphysical package.

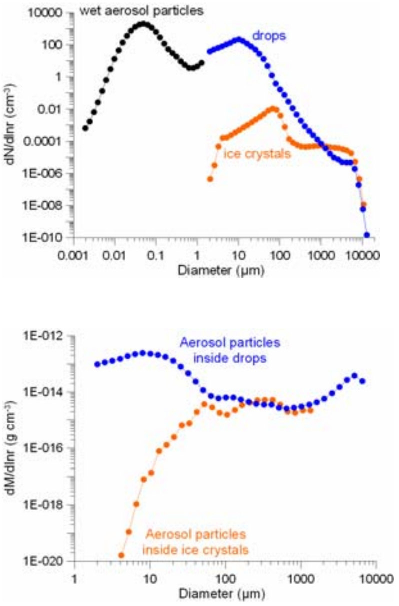

Fig.3: the grid resolution of the different distribution functions treated by DESCAM 3D.

It follows 5 density distribution function: the number distribution function for the aerosol particles $f_{AP}(m_{AP})$, for drops $f_d(m)$ and for the ice particles $f_i(m_i)$, as well as the mass density distribution function for aerosol particles in the drops $g_{AP,d}(m)$and in the ice crystals $g_{AP,i}(m_i)$.

The model considers as warm microphysical processes aerosol particle growth and activation, droplet de-activation, growth of drops by condensation and collision-coalescence. As cold microphysical processes homogeneous and heterogeneous nucleation, growth by vapor deposition, riming and melting are taken into account. A discussion of the different processes considered in the microphysics code can be found in Flossmann and Wobrock (2010), and the coupling with the 3-D code is discussed in Leroy et al (2009).

The model domain is set to 130 km x 130 km in the horizontal and 16 km in the vertical. The resolution is 1 km for the horizontal coordinates and 200 m for the vertical one. A second domain with a surface 64 x 64 km^2 and grid resolution of 250 m is nested into the area of interest. The model domains are represented in Fig.2. The sounding is given in Fig.4.

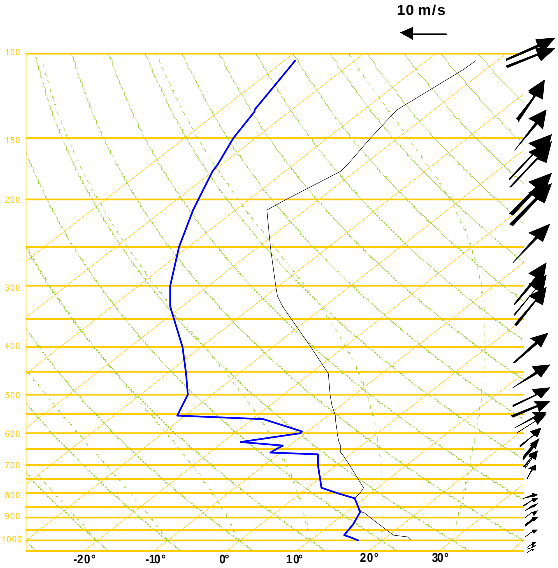

Fig.4: Temperature and dew point temperature profiles from the sounding at Nancy, 12 August 2007, 12 UTC.

The initial aerosol particles distribution follows Jaenicke (1988) for a continental air mass, which corresponds to the solid line in Fig.5. The total number of aerosol particles in the boundary layer is 1411 cm^{-3} and the particles are assumed to be ammonium sulphate, entirely soluble and with a molecular weight of 132 g mol^{-1}. The aerosol particle concentration decreased exponentially up to 3 km, above it is kept constant.

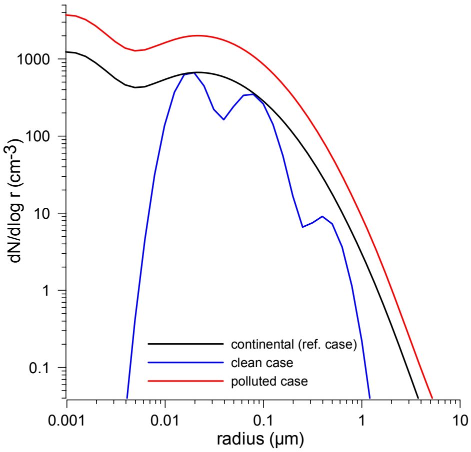

Fig.5: Initial aerosol particle spectra for the reference, clean and polluted case.

3. RESULTS

The cloud field for the COPS field campaign, is shown the figure 6.

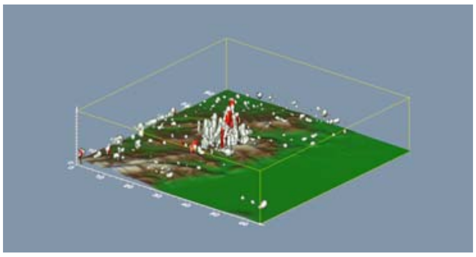

Fig. 6. Cloud field over complex terrain white: cloud water, red: rain water

Radar reflectivity is calculated from the simulation by taking into account the simulated hydrometeor size distribution. The sixth momentum of the raindrop size distribution is used to calculate the radar reflectivity factor Z for liquid hydrometeors:

$$Z\left[mm^6 m^{-3}\right] = \int N(D) D^6 dD \quad .$$

For the ice phase we use the approach of Delanoë et al. (2005):

$$Z\left[mm^6 m^{-3}\right] = \frac{|K_i|^2}{|K_w|^2}\left(\frac{\rho_w}{\rho_i}\right)^2 \int N(D) D^6 dD$$

The ice and water di-electrical constants are $|K_i|^2 = 0.176$ and $|K_w|^2 = 0.93$, respectively.

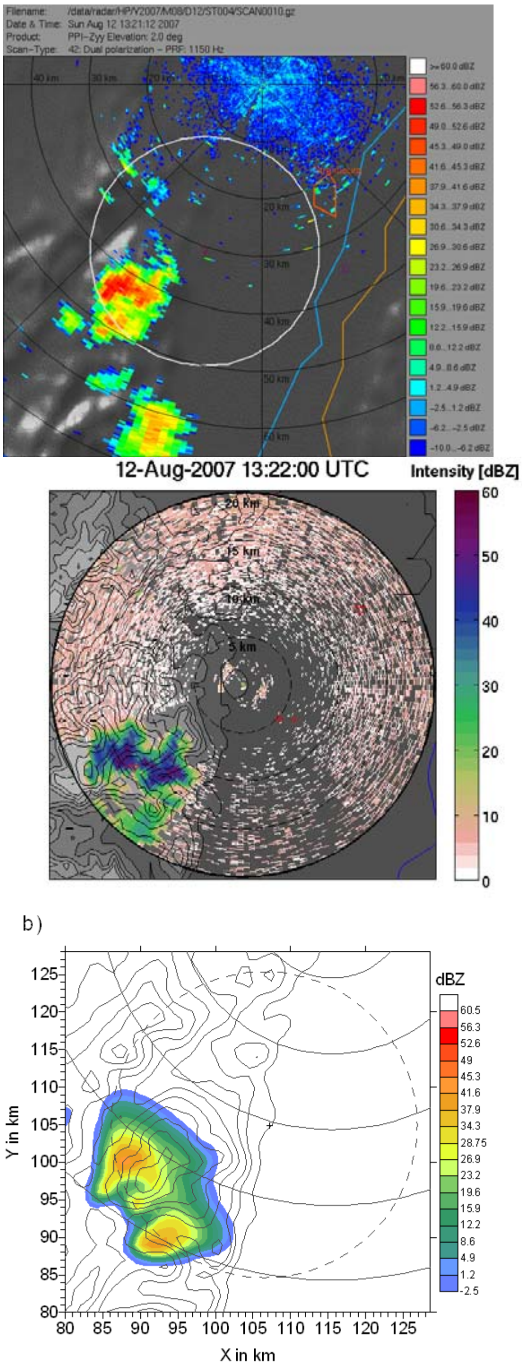

Fig. 7: after 20 min of precipitation (a) observation POLDIRAD, (b) observation X-Band, (c) modelling results.

For the comparison with the radar observations we use herein the normalized radar reflectivity Z_{dBZ}:

$$Z_{\mathrm{dBZ}} = 10\ \log\left[\frac{Z\left[mm^6 m^{-3}\right]}{Z_0}\right]$$

with $Z_0 = 1\,mm^6 m^{-3}$. However, the ice phase contribution is quite small in the case study of 12 August 2007.

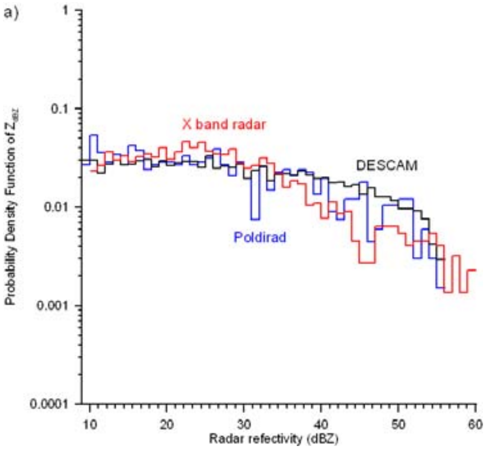

Fig. 8: Probability density function of the X band radar with 300m radial resolution (in red) and Poldirad radar observations at 13h21 UTC (in blue) and modelled (in black) with DESCAM 3D, 20 min after rain onset

For a more objective method to compare the radar measurements with the model results, we analysed the probability density function (PDF) of the radar reflectivities. We selected a resolution of 1dBZ for the probability function and restricted the analyses of modelled and observed values to the range from 10 to 60 dBZ. The range below 10 dBZ was excluded due to noisy data of the X-band radar. Fig.8 shows the PDF for measurements at 13h22 UTC and model results 20 minutes after rain onset.

As in general the features of the observed convective system were well reproduced by the model, we feel encouraged to do the planned sensitivity studies using this situation.

4. SENSITIVITY TESTS

In order to study the importance of aerosol particle concentration for cloud evolution and precipitation formation, several sensitivity studies were performed varying the number of aerosol particles as well as their solubility.

In figure 9, we have analyzed the distribution of the supersaturation in the grid points with liquid water contents exceeding 0.1 g m^{-3} in the white box of fig.9a.

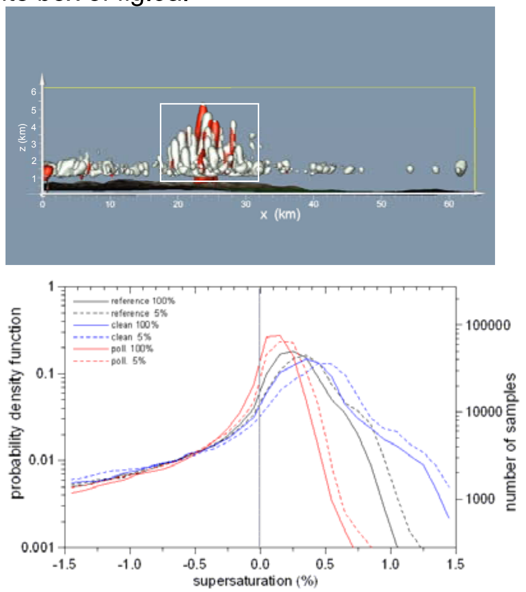

Fig. 9. Role of aerosol concentration and composition for supersaturation in the white box of the cloud field (solid line pertains to complete soluble particles; dashed curve to particles with 0.01% solubility (Planche et al, 2010).

We note that the most probable supersaturation is a function of the total number of particles available and their composition: $S_{\mathrm{mean,min}}$= 0.05% for polluted all soluble particles and $S_{\mathrm{mean,max}}$=0.5% for clean mostly insoluble particles. The influence of the initial aerosol particle spectrum propagates to the precipitation on the ground.

Furthermore, we have studied the distribution of the precipitation on the ground in the different sensitivity cases.

In a first study we just modified the number concentration of the particles. The resulting pattern for the integrated amount of rain on the ground can be found in Fig.10. For these simulations the solubility was kept the same and ε=1.

We note that, globally, decreasing pollution increases the total amount of precipitation and the total watered surface. But locally, the tendency can be opposite. Table 1 summarizes the properties for the three cases.

If the number concentration is kept constant, but only the solubility is varied, we obtain the results of fig. 11:

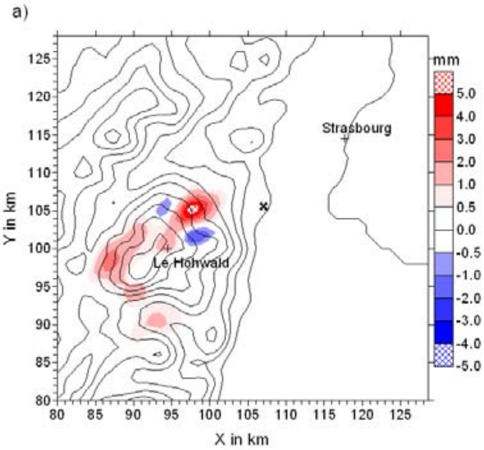

Fig. 10a. Difference continental - polluted

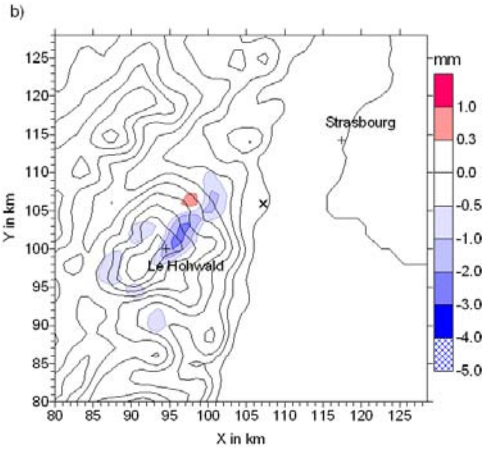

Fig.10b Difference continental - clean

	Continental ε =1	Clean ε =1	Polluted ε =1
rain max. (mm)	7.42	8.02	6.39
mean rain (mm)	1.74	2.00	1.31
rain area (km2)	337	344	307
total rain (Mt)	0.59	0.69	0.40

Table 1: varables for the cumulative rain on the ground as a function of initial aerosol particle number.

We note that globally decreasing solubility increases the total amount of precipitation and the total watered surface. However, the effect is much weaker than the influence of the number concentration. However, locally (Fig.11), we have also an inverse behaviour (Planche et al, 2010).

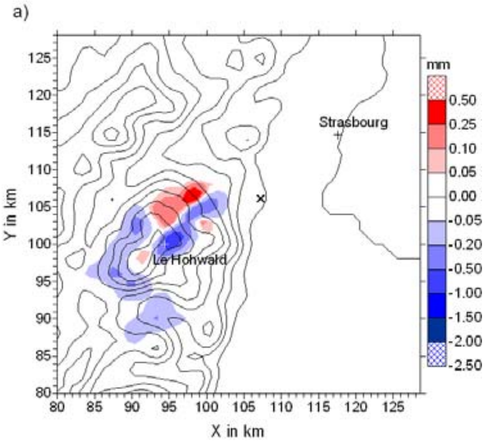

Fig.11a : Difference continental (ε=1 – ε=0.5)

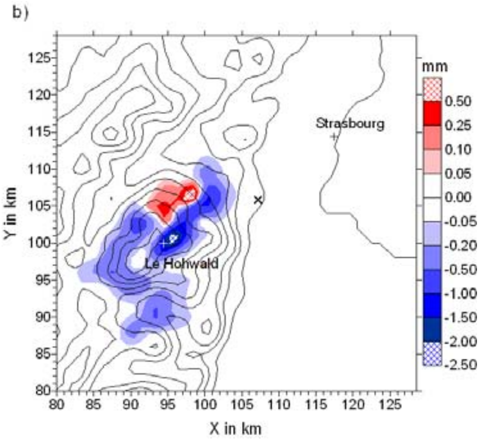

Fig.11b : Difference continental (ε=1 – ε=0.05)

Table 3 shows the overall results if only solubility is varied:

	Continental ε=1	Continental ε =0.5	Continental ε =0.05
rain max. (mm)	7.42	7.52	7.68
mean rain (mm)	1.74	1.80	1.93
rain area (km2)	337	335	339
total rain (Mt)	0.59	0.60	0.65

Table 2: variables for the cumulative rain on the ground as a function of initial aerosol particle solubility.

4. CONCLUSION

In this study for the first time a highly resolved 3D cloud model with detailed microphysics called DESCAM 3D is used to simulate a convective cloud system of medium altitude of cloud top. The simulated radar reflectivity fields and cumulative rain amount show a reasonable agreement with the high resolution X-band and Poldirad radar observations and with the rain gauge of Le Hohwald, even if the statistical behaviour of the two radars are slightly different.

We investigated the role of the initial aerosol particle concentration and solubility on the cloud evolution and precipitation formation.

From the simulations we noted that inside cloud (LWC > 0.1 gm^{-3}) about 20% of the area is subsaturated. The most probable supersaturation in clouds increases with decreasing total number concentration and decreasing solubility. Varying particle number concentration can suppress precipitation for individual clouds. For cloud fields and longer simulations times, we note a strong temporal and spatial variability. Some locations show an inverse behaviour from the overall trend at some time.

For the time being, we can state, that pollution can suppress precipitation, but more research is needed to confirm the behavior for longer periods and larger scales.

One hypothesis needs to be studied: precipitation is mainly determined by water vapor availability; pollution might only influence the spatial and temporal variation. This hypothesis needs to be validated in the future.

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