EFFECTS OF AEROSOLS ON LIGHTNING AND INTENSITY OF HURRICANES

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ABSTRACT

According to the observations of hurricanes located at comparatively low distances from the land, intense and persistent lightning takes place within the 250-300 km-radius ring around the hurricane center, while the lightning activity in the eye wall takes place only during comparatively short periods usually attributed to the eyewall replacement. The mechanism responsible for the formation of the maximum flash density at the TC periphery is not well understood as yet. In this study we hypothesize that lightning at the TC periphery arises under the influence of small continental aerosol particles (AP), which affect the microphysics and the dynamics of clouds at the TC periphery. To show that aerosols change the cloud microstructure and the dynamics in the way to foster lightning formation we use a 2-D mixed phase cloud model with spectral microphysics. It is shown that aerosols which penetrate cloud base of maritime clouds, dramatically increase the amount of supercooled water, as well as the ice contents and vertical velocities. As a result, in clouds developing in the air with high AP concentration ice crystals, graupel, frozen drop/hail and supercooled water can coexist within one and the same cloud zone which allows collisions and charge separation. The simulation of possible aerosol effects on the landfalling tropical cyclone has been carried out using a 3-km resolution WRF mesoscale model. It is shown that aerosols change the cloud microstructure in a way that allows us to attribute the observed lightning structure to the effects of continental aerosols. It is also shown that aerosols, invigorating clouds at 250-300 km from the TC center decrease the convection intensity in the TC center leading to some TC weakening. The results suggest that aerosols change the intensity and the spatial distribution of precipitation in landfalling TCs and can possibly contribute to the weekly cycle of the intensity and precipitation of landfalling TCs.

1. INTRODUCTION

Tropical cyclones (TCs) are known for their destructive power, particularly as they make landfall. TCs are often accompanied by extreme winds, storm surges and torrential rainfall. The TC wind fields, the area of heavy rain, and the rain rate are determined by cloud microphysical processes. At the same time, both experimental and numerical investigations of cloud microphysics in TCs are quite limited. Microphysical observations are usually limited by the zones of the melting level (McFarquhar and Black 2004). Lightning is one of the factors, which can shed light on the microphysical cloud structure and the TC evolution. For instance, increasing lightning rate indicates invigoration of convection, accompanying increasingly larger volumes of graupel or small hail/frozen drop aloft, strengthening updrafts and increasing the probability of heavier rainfall (e.g. Lhermitte and Krehbiel, 1979; Wiens et al, 2005; Fierro et al 2007). The presence of lightning activity in storms crossing the West African coast can be used as a precursor to TC formation (Chronis et al 2007). The appearance and intensification of lightning in the eyewall can be a predictor of TC intensification (Orville and Coyne 1999, Molinari et al 1999; Shao et al 2005; Demetriades and Holle, 2006; Fierro et al 2007). Rodgers et al (2000) found that the closer the lightning is to the storm center, the more likely the TC is to intensify. The latter makes lightning and its distribution in TCs to be an important characteristic that can serve as the predictor of TCs intensity and precipitation changes. Molinari et al (1994, 1999) analyzed the radial distribution of lightning in hurricanes approaching the US coast using the National Lightning Detection Network data. This system detects only the cloud-to-ground (CG) flashes that occur over the

water. They found three zones of distinct electrical characteristics: a) the inner core, which contains a weak maximum of flash density; b) the region with a well –defined minimum in flash density extending 80-100 km outside this maximum; and c) the outer band region, which contains a strong maximum of the flash density within the 200-300 km-radius ring around the hurricane center. Further investigations of lightning in tropical cyclones over the Atlantic and Eastern Pacific Oceans (e.g., Cecil at al, 2002a,b; Demetriades and Holle, 2006) reinforced many findings of Molinari et al (1999) as regards the radial distribution of the flash density. An example of the lightning activity in hurricane Katrina (2005) at several successive time instances is presented in **Figure 1** adopted from the study by Shao et al (2005). Shao et al (2005) applied the LASA data that detect lightning using "very low" and "low" radio frequency (VLF/LF) signals and used time-of-arrival techniques to locate the lightning sources.
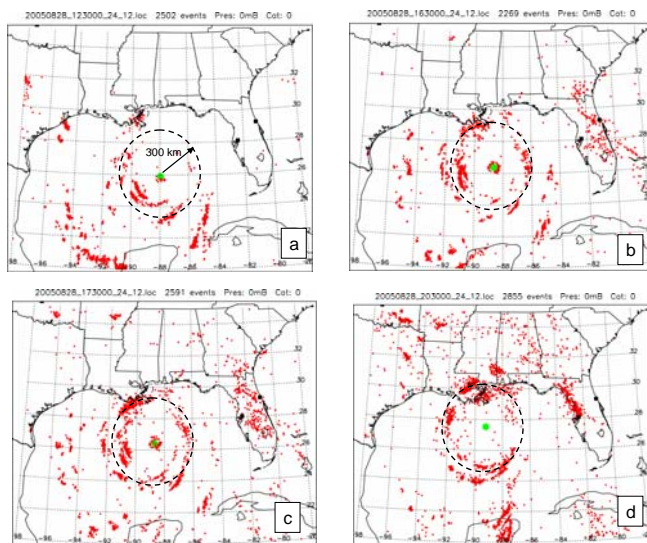

Figure 1. *Lightning in Katrina (2005) at different time instances (after Shao et al, 2005). Zones of lightning are marked by red dots; the TC eye is marked green. Dashed circles have radius of about 300 km around the TC center. One can see the high lightning intensity (which even exceeds that over the land) within the 250-300 km radii rings.*

According to the current concept the charging of hydrometeors in clouds takes place in a mixed-phase region at temperatures less than about -13°C, when collisions of ice crystals and graupel take place in the presence of supercooled droplets [e.g., Takahashi, 1978; Saunders 1993, Cecil et al 2002b; Sherwood et al (2006)]. The latter means that a significant mass of supercooled drops must ascend in clouds above ~5-6 km to trigger flash formation. Actually, the lightning onset indicates significant changes in the cloud dynamics and in the microphysics of maritime clouds: an increase in updrafts and in ice content and the appearance of supercooled water at high levels. Numerical investigations of lightning formation in clouds can be found, for instance, in studies by Solomon and Baker (1994, 1998), (Solomon et al, 2002), Mansell et al (2002) and Mitzeva et al (2006).

Molinari et al (1999) see the reasons causing the rarity of ground flashes in the vicinity of eyewall in the absence of supercooled water in the eyewall (Black and Hallet 1986), which makes a significant charge separation unlikely according to the current theories of noninductive charge transfer between ice particles in the presence of supercooled liquid water (Saunders 1993). The lack of supercooled water in the eyewall can be attributed to a) low updrafts hardly exceeding 5 ms^{-1} (Black et al 1996; Szoke et al 1996; Jorgensen et al 1985); and b) to low aerosol concentrations making the eyewall clouds microphysically extremely maritime. Both reasons lead to the formation of raindrops below the freezing level, which fall efficiently collecting remaining droplets. Low updrafts also do not allow the suspension of graupel, either, which is the key ingredient for electrification. Black and Hallet (1986) also suppose that supercooled water can be eliminated by the efficient seeding of eyewall clouds by ice in the rapidly rotating storm core. The eyewall lightning outbreaks (which tend to occur within relatively small time periods of several hours) are attributed by Molinari et al (1999) to the replacement of the inner eyewall by a new one, which usually precedes TC intensification. Such eyewall lightning outbreaks were also

observed by Demetriades and Holle (2006) in many TC over the Northern Atlantic and the Pacific using the Long-Range Lightning Detection Network. In more details the eyewall replacement has been recently investigated in studies by Houze et al (2006, 2007).

Molinari et al (1999) attribute the well-defined minimum in the lightning flash density extending 80 to 100 km outside the eyewall to lower-tropospheric downdrafts and suppressed convection.

Finally, Molinari et al (1999) attribute the well-defined maximum of the flash density at the TC periphery to the higher instability of environmental air flows around the storm as compared to that in the eyewall. They suppose that the environmental air is convectively unstable, as it follows from the Jordan's (1958) mean West Indies sounding. Williams (1995), Williams and Satori (2004) and Williams et al. (2004, 2005) stressed the dominating role of the atmospheric instability as regards the land-ocean lightning flash density difference. The question arises: Is Jordan's (1958) sounding unstable enough to be fully responsible for the high flash density at the TCs periphery? Emanuel (1994) argued, for instance, that the buoyancy of oceanic convection dramatically decreases by the condensate loading, so that in fact Jordan's (1958) sounding turned out to be close to the neutral one. Molinari et al (1999) also mentioned that the difference in the atmospheric instability between the eyewall and the TC periphery "*cannot be a complete explanation. Tropical oceanic convection often has convective available potential energy values as larger as over the land (Zipser and LeMone 1980), but it does not realize as high a fraction of the pseudoadiabatic ascent rate (Jorgensen and LeMone 1989) and contains much smaller ground flash rates than over the land (see, e.g., Lucas and Orville 1996). The reason for larger flash density in outer bands remains somewhat uncertain*". Note in this connection that simulations of evolution of an idealized hurricane by Fierro et al (2007) using a mesoscale 2-km resolution model with a bulk parameterization scheme describing 12 distinct hydrometeor habits (Straka and Mansell, 2005) and a lightning scheme

(Mansell et al 2002) showed much more intense convection and lightning within a TC ~ 50-km radius central convective zone rather than within outer rain bands. This result was obtained under the environmental conditions typical of the TC formation.

In case the atmospheric instability would be the only mechanism affecting lightning formation, one could expect an increase in the lightning frequency in the zones of high sea surface temperature (SST). To check whether lightning at the TC periphery is related to the SST we compared the SST fields in the area of hurricane Katrina calculated using the TC-ocean coupled model of Graduate School of Oceanography, the University of Rhode Island (I. Ginis, personal communication, 2007) and the lightning intensity. The comparison of the SST fields and the lightning frequency indicates no obvious correlation between the SST and the lightning rate. For instance, according to Fig.1 (panel a) at 12:30 UTC intense lightning takes place to the south from the Katrina center, where the SST is relatively low. At the same time the lightning frequency is very weak to the north of the center, where the SST is maximum. During the 20:30 UTC -24 UTC period intense lightning takes place both to the south and to the north of the TC center, while the SST is significantly higher to the north of the TC center. Besides, there is no clear dependence of the lightning rate within the lightning ring in the south-north direction (see, e.g, Fig 1c), while the SST increases northward. Note that the characteristic scales of the SST changes are much larger than the depth of the lightning ring. Thus, the comparison does not reveal any relationship between the variations of the SST and those of lightning activity at the TC periphery.

Therefore, the first question we address in the paper is: *under what microphysical conditions the instability corresponding to Jordan's (1958) sounding (typical of TC development, Molinari et al, 1999) is able to provide the conditions favorable for negative ground flashes observed at the TC periphery?* In a wider context, the first question can be formulated as follows: " *Can the high instability of the atmosphere, which can take place, say,*

over the zones of especially high sea surface temperature (SST) be considered and necessary and sufficient condition for the intense lightning formation?".

Getting ahead, numerical simulations with a high resolution cloud model with spectral bin microphysics indicate that the instability is necessary, but, supposedly, not sufficient for lightning formation.

In this study we check the hypothesis to the effect that continental aerosols penetrating TC clouds at the TC periphery (taken together with a higher instability at the TC periphery) create conditions favorable for lightning formation. Some support of the idea concerning the possible aerosol effects can be found in **Figures 2a,b.**
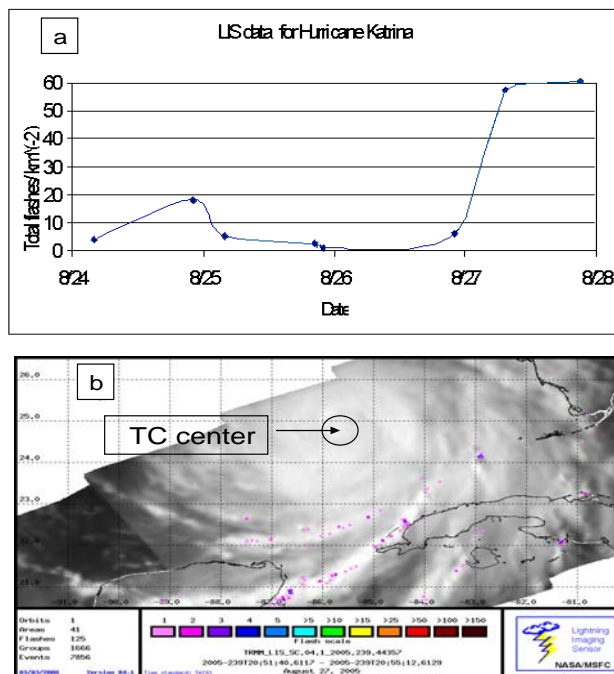

Figure 2. a) The time dependence of total flashes in Hurricane Katrina from 24 Aug. to 28 Aug, 2005. (b) Flash number in Katrina, August 27, 20.51-20.55. The figures are plotted using the Lightning Imaging Sensor (LIS) data. The data above is the total flashes for 6° x 6° area composite to hurricane Katrina best track center (http://thunder.nsstc.nasa.gov/data/#LIS_DATA

Figure 2a shows that intense lightning begins when the TC approach the land at a

comparatively low distance (intense lightning starts on 27-27 Aug. (see also Figure 1a corresponding to the period of the lightning ring formation). Figure 3b shows that lightning arises within the airmass downwind from the land. This air mass can contain a significant amount of aerosols that can affect the microstructure of the convective clouds.

(Note that at present it is impossible to determine the spatial structure of lightning activity in hurricanes far from the land, E. Williams and C. Price, personal communications 2008). The view of Hurricane Rita on September 23, 2005 at 12:52 pm EDT observed by the NASA's TRMM spacecraft is shown in **Figure 3**.

Figure 3. The view of Hurricane Rita on September 23, 2005 at 12:52 pm EDT observed by NASA's TRMM spacecraft. The cloud cover is evaluated by TRMM's Visible and Infrared Scanner and the GOES spacecraft (http://www.nasa.gov/vision/earth/lookingatearth/h2005_rita.html).

One can see that TC circulation penetrates far into the land even when the TC center is located several hundred kilometers from the cost line. The latter means that the TCs approaching the land involve a lot of continental aerosols into their circulation. This assumption will be tested using a numerical model. We will also show that these aerosols can lead to the formation of a narrow ring with the radii of 250-300 km and a high lightning flash density. Finally, we evaluate the possible effect of

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continental aerosols on the intensity, the structure of cloudiness and precipitation of landfalling TC.

The effects of aerosols on individual convective clouds under conditions typical of the TC periphery are investigated using the 250 m resolution 2-D mixed phase Hebrew University cloud model (HUCM) with spectral bin microphysics (Khain et al 2004, 2005). The aerosol effects on the cloudiness structure of a TC approaching and penetrating the land is investigated using a two nested grid Weather Research Model (WRF) (Skamarock et al., 2005). In the latter case the WRF was used to simulate the evolution of a TC moving along the track close to that of Hurricane Katrina from August 28 12z to August 30 00 z.

The combination of the two models for the investigation is natural: the high resolution HUCM can describe fine microphysical processes to evaluate the aerosol effects on the vertical profiles of liquid water and ice particles contents in individual clouds. These simulations will represent the main justification that an increase in the aerosol concentration can create conditions favorable for lightning in maritime clouds. The 3-D simulations with the WRF model cannot be carried out with so high resolution and with the microphysical scheme used in the HUCM. These simulations do not resolve small clouds and underestimate vertical velocities. Thus, the TC simulations provide less exact but useful evidence concerning aerosol effects on individual clouds. At the same time the 3D model illustrates the effects of aerosols on the structure of TC cloudiness and precipitation over large areas.

Taking into account the factors affecting the lightning rate (Fierro et al 2007) the lightning probability will be characterized by the product of the total ice content, the supercooled content and the vertical velocity above the 5-km level.

2. AEROSOL EFFECTS ON THE INDIVIDUAL TROPICAL MARITIME CLOUDS

2.1 DESCRIPTION OF THE CLOUD MODEL

A 2-D mixed phase Hebrew University cloud model (HUCM) with spectral bin microphysics has been used (Khain and Sednev 1996; Khain et al 2004, 2005) to investigate whether an increase in aerosol concentration in the zone of maritime tropical convection can change the cloud microphysical structure of individual clouds to make it favorable for lightning formation. The HUCM model microphysics is based on solving a kinetic equations system for size distribution functions for water drops, ice crystals (plate-, columnar- and branch types), aggregates, graupel and hail/frozen drops, as well as atmospheric aerosol particles (AP). Each size distribution is described using 43 doubling mass bins, allowing simulation of graupel and hail with the sizes up to 4 cm in diameter. The model is specially designed to take into account the AP effects on the cloud microphysics, dynamics, and precipitation. *The initial* (at t=0) CCN size distribution is calculated using the empirical dependence

$$N = N_o S_1^k , \qquad (1)$$

and applying the procedure described by Khain et al (2000). In (1) N is the concentration of activated AP (nucleated droplets) at supersaturation S_1 (in %) is with respect to water, N_o and k are the measured constants. The values of the constants used in the simulations are presented in Table 1. At t>0 the prognostic equation for the size distribution of non-activated AP is solved. Using the value of S_1 calculated at each time step, the critical AP radius is calculated according to the Kohler theory. The APs with the radii exceeding the critical value are activated and new droplets are nucleated. The corresponding bins of the CCN size distributions become empty.

Primary nucleation of each type of ice crystals is performed within its own temperature range following Takahashi et al (1991). The dependence of the ice nuclei concentration on supersaturation with respect to ice is described using an empirical expression suggested by Meyers et al. (1992) and applied using a semi-lagrangian approach (Khain et al 2000) allowing the utilization of the diagnostic relationship in the time dependent framework. The secondary

ice generation is described according to Hallett and Mossop (1974). The rate of drop freezing is described following the observations of immersion nuclei by Vali (1974, 1975), and homogeneous freezing according to Pruppacher (1995). The homogeneous freezing takes place at temperature about -38C. The diffusional growth/evaporation of droplets and the deposition/sublimation of ice particles are calculated using analytical solutions for supersaturation with respect to water and ice. An efficient and accurate method of solving the stochastic kinetic equation for collisions (Bott, 1998) was extended to a system of stochastic kinetic equations calculating water-ice and ice-ice collisions. The model uses height dependent drop-drop and drop-graupel collision kernels following Khain et al, (2001a) and Pinsky et al (2001). Ice-ice collection rates are assumed to be temperature dependent (Pruppacher and Klett, 1997). An increase in the water-water and water-ice collision kernels caused by the turbulent/inertia mechanism was taken into account according to Pinsky and Khain (1998) and Pinsky et al. (1999, 2000). Advection of scalar values is performed using the positively defined conservative scheme proposed by Bott (1989). The computational domain is 178 km x 16 km with the resolution of 250 m and 125 m in the horizontal and vertical directions, respectively.

2.2. DESIGN OF SIMULATIONS WITH THE CLOUD MODEL

Simulations of individual maritime deep convective clouds under the conditions typical of tropical oceans during the hurricane season (Jordan 1958) have been performed. The sounding used is quite close to that observed during the GATE-74 Experiment, 261 day used for cloud simulations in many studies (e.g. Ferrier and Houze 1989; Khain et al 2004, 2005). The sounding used indicates high about 90 %. humidity near the surface. The sea surface temperature is $27^{o}C$.

The averaged wind shear was 15 m/s per the atmospheric layer of 14 km depth. The zero temperature level is at 4.2 km. The reasons of the choice of the Jordan (1958) sounding in our simulations are as follows. The Jordan (1958) sounding was used by a great number of scientists simulating TC genesis and evolution (e.g, Khain et al 1983, 1984). This sounding was mentioned by Molinari et al (1999) in their attempt to explain, why the lightning at the TC periphery is stronger than in the eyewall. The main reason why this sounding has been used in our 2-D simulations is that it is quite unstable and leads to the formation of deep convective clouds with the vertical velocity maximum of 18-20 m/s (see below). Such vertical velocities are unusually high for maritime convection (with typical updrafts of 5-10 m/s). We used this sounding to check the "thermal" or "instability" hypothesis, namely, *is the atmospheric instability the sufficient condition for lightning formation, i.e., are the conditions favorable for lightning formation if vertical updrafts are high, but aerosol concentration is very low?*

Note that we have no data concerning the possible aerosol flux from the land to the sea area. Hence, we assume that the aerosol particle (AP) size distributions over the sea can be represented as the sum of the distributions typical of maritime and continental conditions. The aerosol particles in all simulations were assumed soluble. Under high winds aerosol size distribution can contain a significant amount of large cloud condensational nuclei (CCN) arising because of the sea spray formation. Hence, the penetration of continental aerosol must create aerosol size distributions containing a significant concentration of both small continental aerosols and tails of large aerosols. To investigate the effects of a high concentration of small continental aerosols on cloud microphysics and dynamics in the presence of a significant amount of large maritime CCN, the

following simulations have been performed (see table 1): a) the "M-case" corresponding to the typical maritime distribution outside the area of strong winds. In this case the maximum radius of dry AP was set equal to 2-μm. The CCN number (at S=1%) was set equal to 60 cm^{-3}. The activation of the largest APs leads to the formation of 10 μm-radius droplets. b) the "M_c case", in which the AP distribution represents the sum of a continental AP distribution (with the maximum AP radius of 0.6 μm) and the maritime distribution similar to the M-case described previously. We suppose that this case represents the AP size distribution over the sea when continental aerosol intrusion under weak and moderate winds occurs. c) In the "M_tail-case" the AP distribution is similar to that in the M-case within the dry AP radii range below 0.6 μm, but with a 100 times higher AP concentration with radii exceeding 0.6 μm. As a result, the concentration of dry CCN (at S=1%) with the radii exceeding 0.6 μm is 60 cm^{-3}, which includes the concentration of 2 μm-

radius CCN of 3.5 cm^{-3}. We suppose that this case may represent the AP distribution under hurricane winds in the central TC zone. d) In the "M_c_tail case" the AP is the same as in M_c, but with a large CCN tail of the size distribution as in the M-tail case. We suppose that this case represents the AP size distribution under the intrusion of continental aerosols and very strong wind.

Parameters N_o and k assumed in these simulations and the corresponding references are presented in **Table 1**.

We have not included the CCN with the dry radius above 2 μm into the simulations. In all simulations clouds were triggered by the initial heating within the zone centered at x=54 km, to allow the cloud hydrometeors be located in the computational zone during a longer time period. The maximum heating rate was set equal to 0.01 $^{o}C/s$ in the center of the 4.9 km x 2 km heating area and decreased linearly to the periphery of the zones. The duration of initial heating was 600 s in all experiments.

Table 1. List of simulations and parameters characterizing aerosol distributions

Type of cloud	Short title	$N_0(cm^{-3})$	k	references
Maritime cloud	M	60	0.308	Khain et al 2005
Sum of maritime and continental	M-c	2500 60	0.921 0.308	Khain et al 2001b; 2004; 2005; Khain and Pokrovsky, 2004
Maritime cloud with increased fraction of large CCN	M_tail cloud	60 tail: 60 CCN with the radii exceeding 0.6 μm	0.308	Sensitivity study
Sum of maritime and continental but with the increased fraction of large CCN	M_c_tail	2500 cm^{-3} 60 cm^{-3} tail: 60 particles with the radii exceeding 0.6 μm	0.92 0.308	Sensitivity study.

The maximum value of the dynamical time step was 5 s. Most simulations were conducted for 3 to 4 hours.

2.3 RESULTS OF SIMULATIONS

The first important result was the following: the tail of large CCN (within the 0.6 μm to 2 μm radii range) actually does not influence cloud microstructure and precipitation in the presence of a high concentration of small CCN. For instance, horizontally averaged accumulated rain amounts in the M_c and M_c_tail simulations are just similar (**Figure 4**).
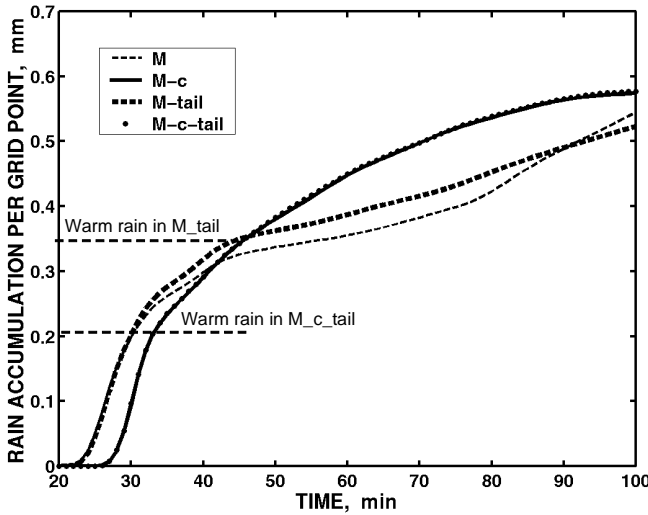

Figure 4. Time dependence of accumulated precipitation at the surface in the simulations with different concentrations and size distributions of aerosols (see Table 1). The horizontal dashed lines denote the approximate amount of warm rain in the simulations. The remaining accumulated rain is mostly cold (melted) precipitation. The warm rain from the clouds developing in the polluted air is smaller than that in the cloud developing in clean air. At the same time, the total accumulated rain is larger in clouds developing in polluted air.

This result can be explained as follows: the rate of the diffusion growth is determined by supersaturation. In the

case when about 1000 cm^{-3} droplets are nucleated (as in the M_c and the M_c_tail cases), the supersaturated water vapor is shared between a great number of droplets and the supersaturation is small. Moreover, the droplets growing on large CCN also contribute to the decrease in the supersaturation. As a result, the largest nucleated droplets with the initial radius of about 8 μm grow relatively slowly and reach the size necessary to collect smaller droplets at the heights of about 5.5-6 km. It is necessary to recall that smaller droplets increase their size by diffusional growth faster than the largest ones (Rogers and Yau 1989), so that at the 6 km level the contribution of the CCN tail to the concentration of raindrops is not substantial. A stronger effect of the large tail takes place in the case of low droplet concentration, when supersaturation is higher than in the case of high droplet concentration. Hence, raindrops in the M_tail form at about 3 km and not at 4 km as in the M-case. The accumulated rain in the M_tail run is larger than that in the M-run during the first 1.5 h. However, the difference between the rain amounts is not substantial in this case either. The low sensitivity of clouds developing under tropical maritime environmental conditions to the amount of large CCN allows us to discuss the effects of continental aerosols on TC clouds under a high uncertainty as regards the concentration of large CCN. Since we expect the existence of the tail of large CCN in the TC clouds, we will discuss the effects of continental aerosols on the cloud structure in the M_tail and the M_c_tail simulations below. Note that no giant CCN with the radii above 15-20 μm, which are able to trigger drop collisions immediately after their penetration into a cloud, were assumed in the simulations.

Figure 5 shows the fields of cloud water contents CWC (droplets with the radii below 50 μm) in the M_tail and M_c_tail simulations at t=25 min. One can see that while the CWC in clouds with a low AP concentration decreases dramatically above 3.5-4 km because of rapid raindrop formation, the CWC in clouds with a high AP concentration remains significant till
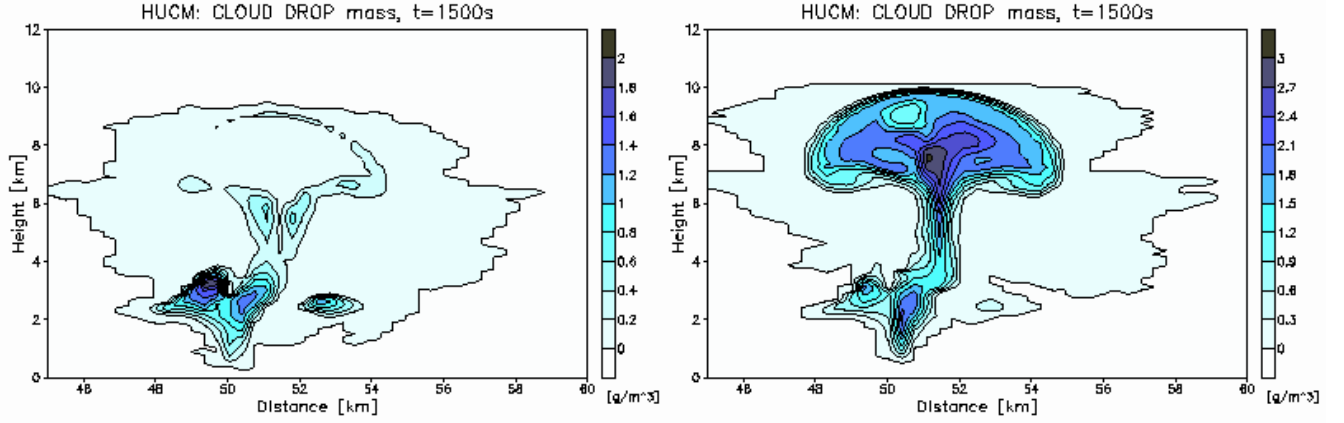
Figure 5. Fields of the cloud water content CWC (droplets with the radii below 50 μm) in the M_tail (left) and M_c_tail (right) simulations at t=25 min. The CWC is significantly higher and reaches higher levels in the clouds developing in polluted air.

the upper atmosphere. The latter is a typical feature of the clouds developing in highly polluted air (e.g., Andreae et al 2004; Ramanathan et al, 2001; Khain et al 2004, 2005). The specific feature of the present results is that high CWC forms in the polluted air in the presence of a high concentration of large CCN.
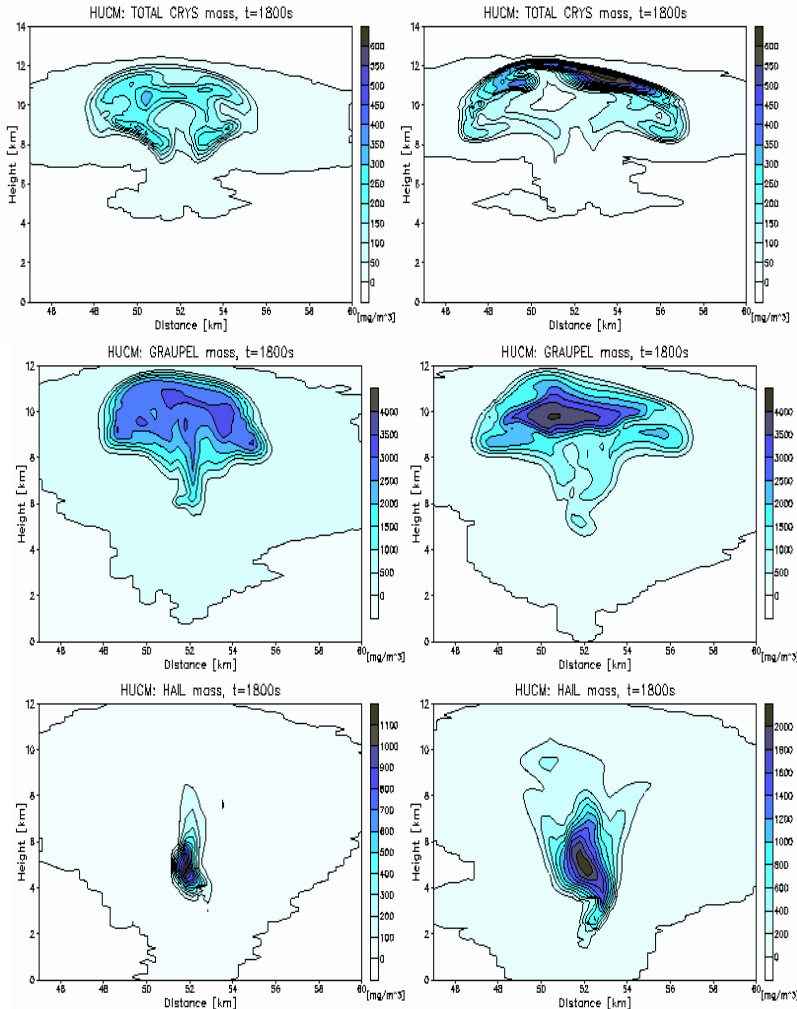

Figure 6. The fields of crystal (upper panels), graupel (middle) and hail/frozen drops contents in the M_tail (left) and the M_c_tail (right) simulations. One can see that these contents are higher in the clouds developing under high aerosol concentration.

Figure 6 shows the fields of crystal (upper panels), graupel (middle) and hail/frozen drops contents in the M_tail (left) and the M_c_tail (right) simulations. One can see that these contents are higher in the clouds developing in continental air, which can be attributed to a larger amount of the CWC penetrating above the freezing level in this case. On the average the total mass of graupel under high aerosol concentration is about 1.5 times higher than in case of low aerosol concentration.

Comparison of the fields of the vertical velocity in the M_tail and in the M_c_tail runs indicates (not shown) that the vertical velocities in the clouds developing within high AP concentration air are higher by a few ms^{-1} than in the clouds developing under low aerosol concentration cases. This aerosol effect on cloud dynamics was simulated and discussed in detail by Khain et al (2005) and then simulated in many studies (e.g. Lynn et al 2005 a,b). The increase in the vertical updrafts in tropical clouds developing in microphysically continental air can be attributed to an extra latent heat release caused by an extra condensational droplet growth (larger CWC) and an extra freezing (higher ice contents) (see Figures 5 and 6). Note that the difference between the vertical velocities (a few meters per second) is much smaller than the maximum updrafts of 15-18 ms^{-1}. Thus, the effect of aerosols on the maximum velocities is comparatively small. The vertical velocities are determined mainly by the CAPE in agreement with the results reported by Williams (1995) and Williams et al. (2004, 2005). The cloud base tops were actually similar in both clean and polluted air cases. This result agrees well with those reported by Khain et al (2005). The comparatively small effect of aerosols on the dynamical characteristics of clouds can be attributed to the following. On the one hand, the latent heat release in the clouds developing in the high CCN concentration case is larger, which can be seen by the generation of larger condensate mass. On the other hand, the mass loading in clouds developing in the microphysically continental air is larger. These two factors have opposite effects on the buoyancy and largely compensate each other. Thus, the dramatic difference between the microstructure of clouds seen in Figures 5 and 6 can be attributed mainly to aerosol effects. The comparison of the mass contents fields shown in Figures 5 and 6 indicates that favorable conditions for the charge separation (the coexistence of a significant amount of supercooled water, ice crystals and graupel at the temperatures below -13 ^oC) exists only in the case of high aerosol

concentration. Hence, *the second conclusion that can be derived from the results is that the instability of the atmosphere at the TC periphery corresponding to the Jordan's (1958) sounding is necessary, but the condition it is not sufficient to create the microphysical structure favorable for intense electrification.*

So, it appears that aerosols represent an important component as regards production of a significant amount of supercooled water at the upper levels even under unstable atmospheric conditions, and in particular under maritime conditions. Of course, we mean the results of the numerical cloud model with the parameters of the aerosol distributions chosen as mentioned in Table 1.

Figure 7 (upper panels) shows the radar reflectivity fields in the M_tail (left), and in the M_c_tail (right) simulations probably representing the clouds in the eye wall (clean air) and at the periphery cloud bands in case of continental aerosol intrusion, respectively.
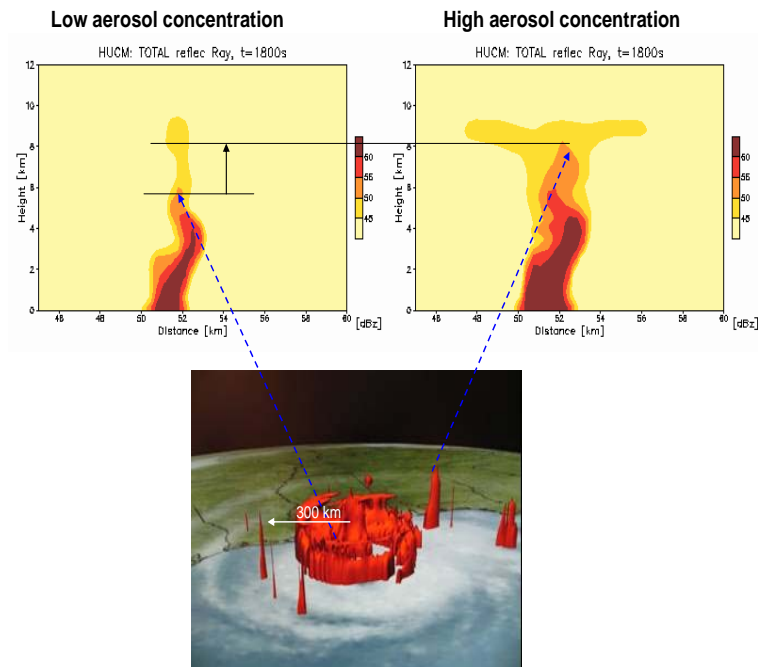

Figure 7. *.The upper row: Radar reflectivity fields in the M_tail (left), the M_c_tail (right) simulations. Low panel: The rain structure of hurricane Rita.*

The comparison shows that the high values of the radar reflectivity in the clouds forming in the case of high AP concentration reach (better to say, start) at higher levels as compared to the clouds developing in clean air (one can expect that the air in the TC eyewall can be regarded as clear, see below). In the low panel the rain structure of hurricane Rita measured by the TRMM's Tropical Microwave Imager (TMI) on September 23, 2005 at 12:52 pm EDT is shown. At this time the storm was the category 4 hurricane with the minimum pressure of 924 mb, and sustained winds of 60 ms^{-1}. The zones with the precipitation rate at least 2.0 inches (58 mm/h) of rain per hour are marked. One can see that the model reveals some similar features, namely: precipitation particles start forming at higher levels at the TC periphery than in the eyewall.

We interpret all these results as some justification of the hypothesis *that aerosols penetrating into the maritime convective clouds are able to change dramatically the cloud microphysics and dynamics. These effects plus the atmospheric instability make possible the coexistence of cloud ice and supercooled water at high levels with the temperatures below* $-13^{o}C$ *, i.e, the formation of thunderstorms over the ocean.*

Figure 4 shows the time dependence of accumulated precipitation at the surface in the simulations with different concentrations of small and large CCN (see Table 1). Horizontal dashed lines denote the transition from the warm rain regime to the cold (melted) rain. This transition is seen by the change in the slopes of the curves. Cold (melted) rain is less intense and the slope is lower. The ratio of the accumulated rain amount toward the end of warm rain and beginning of cold (melted) rain to the total accumulated rain represents the fraction of the warm rain within the cumulative precipitation. One can see that a) the accumulated precipitation from clouds developing in microphysically continental air is higher, which corresponds to a higher

convective heating of the atmosphere. As it was discussed in Khain et al (2004, 2005, 2008), a higher accumulated rain is related to the generation of larger condensate mass in clouds developing in polluted air. At the same time the precipitation loss in the wet air by evaporation is low. There was no secondary cloud formation in these simulations.

b) The warm rain amount decreases in clouds developing in microphysically continental air, so that most precipitation in dirty clouds is cold rain formed by melting graupel and hail. On the contrary, precipitation from clouds developing in clean air is mainly warm rain.

It is reasonable to assume that the concentration of aerosols decreases from the TC periphery toward the TC center. This decrease can be caused by the wash out of aerosols in TC rain bands, as well as by a significant time required by the aerosols to reach the TC center in the TC boundary layer (see below). Hence, the conditions at the TC periphery where the concentration of continental CCN is higher remain favorable for flash formation during the whole time period when the TCs are in the vicinity of the land.

The conclusions derived in this section help us to carry out the simulations using a 3-D mesoscale model.

3. AEROSOL EFFECTS ON THE TCs APPROACHING THE LAND

A two nested grid Weather Research Model (WRF) was used to simulate landfall of a hurricane in the Gulf of Mexico. The resolution of the finest and the outer grid was 3 km and 9 km, respectively. The number of the vertical levels was 31, with the distances between the levels increasing with the height. The Thompson et al. (2006) bulk-parameterization was applied. Hurricane Katrina (August 2005) has been TC chosen as the case study. The initial fields were taken from the Global Forecast System Reanalysis data. The lateral boundary conditions were

updated every six hours using the data as well. The Gulf of Mexico surface water temperatures were initialized on 28 August 12 Z, and were not updated during the experiments described below.

Note that the WRF used was not a TC forecast model, so that no specific adjustment procedures were used to adopt the TC structure derived from the crude resolution Reanalysis data to the intensity of the real TC at t=0. Hence, some relaxation period was required to get the model TC intensity close to the observed one. Note, however, that the accurate prediction of the Katrina's intensity was not the primary purpose of the study. The main purpose of the simulations was to compare the lightning spatial distributions in the simulations with and without aerosol effects on the clouds at the TC periphery in a strong hurricane, which would be able to involve aerosols from the continent. Because of the computer limitations, the simulations were performed in two stages.

At the first stage the TC was simulated on 27 August 00 Z (when it was located to the south of Florida) up to August 30, 00Z. The purpose of the simulation was to check whether continental aerosols, which were assumed to be located initially over the land (with zero concentration over the sea) may be involved into the TC circulation and penetrate into the TC to the distances up to about 300 km to the TC center toward the beginning of lightning in the Mexican Gulf (Fig 1). Aerosols were considered to be a passive scalar in the run.

Figure 8 (left panels) shows time dependence of the minimum pressure and tracks of TC simulated. The minimum pressure and track of TC Katrina is shown as well. One can see that the model TC (after some spin up period) reaches the intensity of hurricane Katrina. The errors in the location of TC center are small and hardly can affect significantly the

formation of the spatial aerosol distribution. We believe that these results indicate that the model TC creates the field of the continental aerosol concentration quite similar to that could be expected around hurricane Katrina.
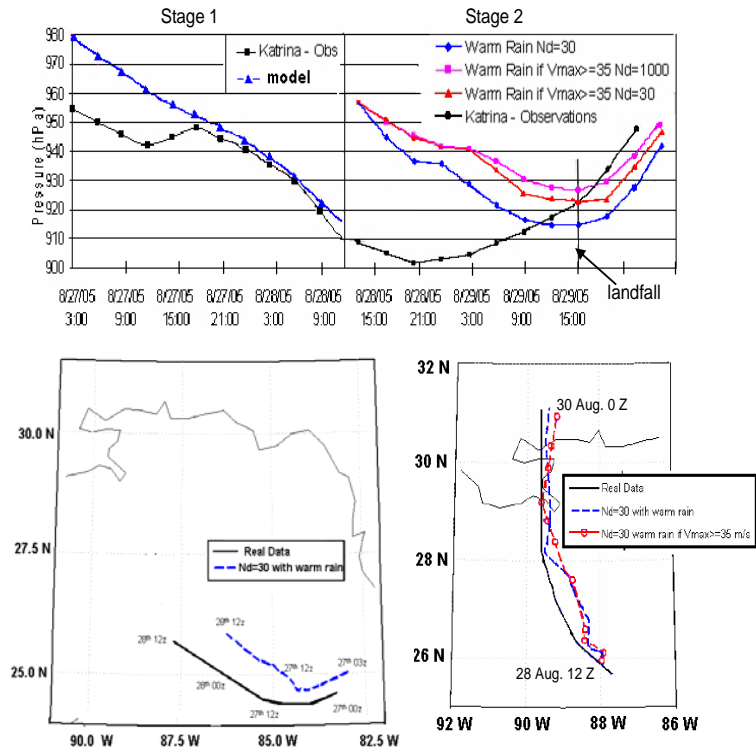

Figure 8. *Time dependencies of the minimum pressure and tracks of the model TC and of Katrina in two simulation stages: Aug. 27-Aug.28 (left panes) and Aug. 28- Aug 29 (right panels). The right upper panel shows that the warm rain prevention at the TC periphery increases the pressure by 7-12 mb, i.e., decreases the TC intensity.*

Figure 9 shows the field of aerosol concentration toward the end of the first stage of simulations. One can see that aerosols do penetrate the lower troposphere of the TC "in time". Towards August 28 , 12 Z aerosols form a front with the radius of about 250-300 km and a quite sharp gradient of aerosol concentration: while at the radii $r \geq 250 km$ the concentration in the

lower atmosphere was actually similar to that over the land, the central TC zone with the radius below 250 km was free from the continental aerosols. These results indicate that while lightning at the distances of a few hundred km from the TC center may be related to aerosols as it is hypothesized in the present study, the lightning in the TC eye wall seen in Fig 1 is not related to the continental aerosol effects. Note that developed rain bands form in the TC within zone of about 300 km radius. We suppose that the concentric lightning ring seen in Figure 1 forms in the zone of the aerosol "front", which transforms the maritime clouds into thunderstorms with more continental characteristics.

Figure 9 (left) shows that the aerosols penetrate closer toward the TC center on the south side. This effect can be attributed to the fact that aerosols are advected along spirals by the TC wind speed. Aerosols starting their motion at the continent (to the north of the TC) should approach the TC center during their motion along the spirals because of the radial wind directed inward. Thus, aerosols should be closer to the TC center on the southern and the eastern sides of TCs then in the west side. However, the AP concentration decreases along the stream limes. Hence, the concentration on the south side turns out to be higher than on the east side. It is interesting to note that the spatial distribution of the AP
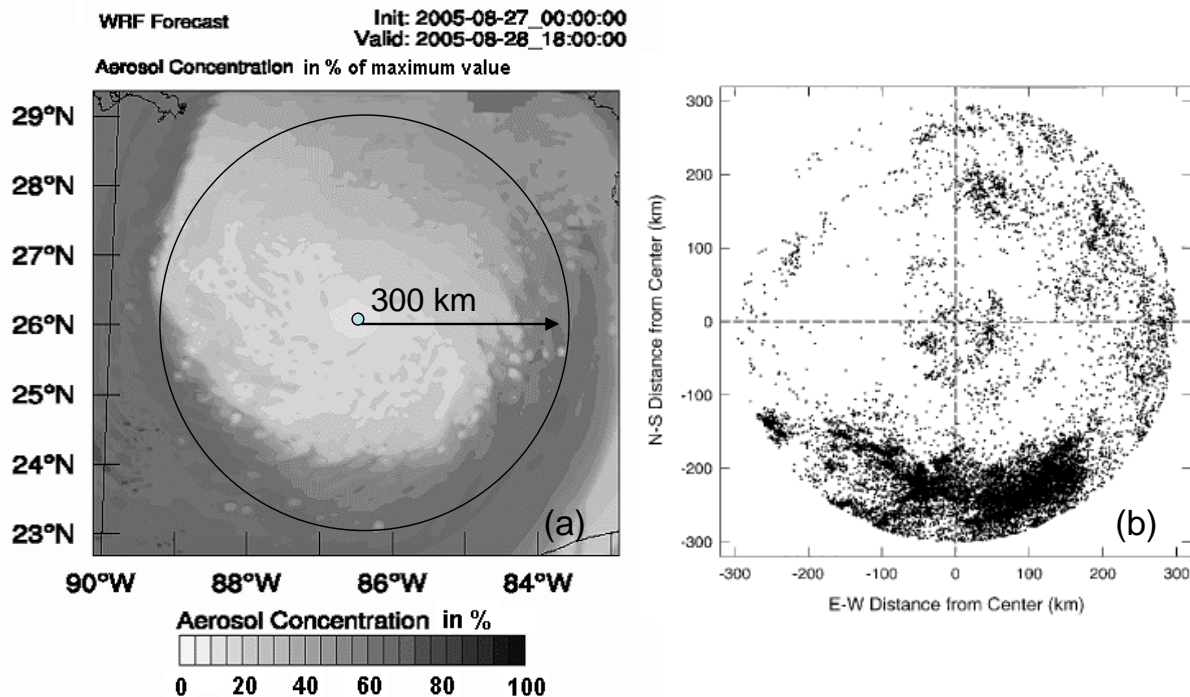

Figure 9. The field of the aerosol concentration simulated in the TC zone on 28 August, 12 00Z in % to the maximum value assumed to be over the continent (left panel). The right panel adopted from Molinari et al (1999) is composited with respect to the hourly center position of each hurricane. One can see a good correlation between the distribution of aerosols and the lightning density.

concentration obtained in the simulation resembles the locations of negative ground flashes observed during hurricane stage in the nine storms in the Western Atlantic (see Figure 9 (right) adopted from Molinari et al 1999), composited with respect to the hourly center position of each hurricane. One can see a good correlation between the distribution of aerosols and the lightning density. The similarity of the fields may be interpreted as some evidence of the validity of the aerosol hypothesis (at

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least, it does not contradict the hypothesis).

In the simulation, in which aerosols are treated as a passive scalar the aerosol front approaches the TC center in the inflow layer quite slowly, with the velocity close to the radial velocity of the flow. Passive aerosols reach the central TC towards the time, when Katrina was quite close to the land and the lightning in its eye wall had already terminated (Fig. 1, right panel). In the case cloud-aerosol interaction is taken into account, the concentration of continental aerosols must decrease from the TC periphery toward the TC center even more intensely because of the wash out of aerosols in the TC rain bands.

The results indicate the following:

a) The lightning in the TC eye wall is not related to aerosols involved into the TC circulation from the continent. To attain a significant concentration of supercooled droplets above the -13°C level in the TC eye wall the vertical velocities must be especially high (or some still unknown mechanisms should be involved). At the same time the conditions at the TC periphery where the concentration of continental CCN is high remain favorable for flash formation during the whole time period when the TC approaches or penetrates the land. This result agrees with the observations that the lightning in the TC eye wall takes place only during eyewall replacements, while the lightning is a permanent feature at the TC periphery of TCs approaching the land (see .e.g. Fig 1).

b) The problem whether giant CCN arise in the zone of the maximum winds or not is not very important for our purpose because continental aerosols do not reach the TC center (at least, in the TC simulated in the study).

At the second stage the calculations of the TC were performed for the cases when the aerosol effects on TC clouds were taken into account. The calculation was performed within the period Aug. 28, 12 00Z to 30 Aug. 00 Z. The way how the aerosol effects were accounted for requires some preliminary comments. The 3-km resolution and the relatively crude vertical resolution do not allow one to reproduce the fine microphysical cloud features discussed in Section 2.3. It is well known that crude resolution decreases vertical velocities, whose values are of crucial importance for the simulation of cloud dynamics, microphysics and precipitation. As it was shown by Khain et al (2004), the increase in the distance between the neighbouring grids in the 2-D HUCM from 250 m to 1 km has resulted in a decrease in the maximum vertical velocities from 25 ms^{-1} to 10 ms^{-1}. Since the fall velocities of different hydrometeors do not depend on the grid resolution, the cloud microphysical structure turns out to be quite different under different grid resolutions. For instance, the terminal velocity of raindrops is about 10 ms^{-1}. In case the vertical updraft velocity is less than 10 ms^{-1} raindrops will rapidly fall out. Thus, the utilization of the crude resolution artificially makes the clouds "maritime" by their character, even if the concentration of AP is high. Besides, most bulk-parameterization schemes are, as a rule, less sensitive to aerosols than the spectral bin microphysical scheme (Lynn et al 2005b). To "sidestep" this problem, we simulated the effects of continental aerosols in the simulations by preventing warm rain in the scheme altogether. As seen in Figure 4, it is hardly possible to prevent warm rain in maritime deep clouds by an increase in aerosol concentration up to the magnitudes typical of continental conditions. However, small aerosols significantly decrease the warm rain amount, and even more relevantly, the prevention of warm rain in the 3-D model allows the reproduction of aerosol

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effects qualitatively similar to those obtained in the 2-D 250 m resolution cloud model, namely, a substantial transport of large CWC upward and an increase in the ice content.

The following simulations have been performed with the 3-D WRF model at the second stage. The control run allowed warm rain (WR) formation by drop-drop collisions (the WR run). In this run droplet concentration at the cloud base N_d was set equal to 30 cm^{-3}. This case corresponds to the M-run with the 2D model, where CCN concentration (at S=1%) was assumed to be 60 cm^{-3} (usually about half of the available CCN are activated). In the second run referred to as No Warm Rain at the Periphery (NWRP-30), aerosol effects were parameterized by shutting off the drop-drop collisions only at the hurricane periphery, where the surface wind was smaller than 35 m s⁻¹. This threshold was chosen because the continental aerosol concentration in the TC central zone where wind speed exceeds 35 m s⁻¹ should be negligibly small, as it was discussed above. Besides, very high wind speed supposedly is able to produce a significant amount of giant CCN in the eye wall which most likely renders any effects of small aerosols ineffective. Hence, warm rain is shut off only in that part of the hurricane that has winds speeds less than the threshold value. A similar approach has been used by Rosenfeld et al (2007b). The third simulation, referred to as NWRP-1000, is similar to NWRP-30, but the droplet concentration at the cloud base was set 1000 cm^{-3}. This is a supplemental run, whose purpose was to illustrate the difficulties connected with simulation of the continental aerosol effects on maritime convection using the model with crude horizontal and vertical resolution and changing only the droplet concentration.

Figure 8 (right lower panel) shows the tracks of the simulated storms in the WR and NWRP-30 runs (The TC track in the NWRP-1000 was actually similar to that in NWRP-30) as well as the track of hurricane Katrina. One can see that the model reproduces Karina's track well enough. The deviation of the model TC track from that of the real TC can be attributed to the errors in the initial TC intensity derived from the Reanalysis data, rather than to aerosol effects. The simulated minimum pressures of the storm with time are shown in **Figure 8** (right upper panel). After a relatively short spin up period, the simulated TC reaches the super hurricane intensity with the minimum pressure of about 915-920 hPa, also observed in Katrina. In spite of the differences between the intensity of the simulated TC and that of Karina during the spin up period, the intensity of the simulated storm was quite close to that of Katrina when the intense lightning at the Katrina's periphery was observed (Fig. 1). The main result that follows from Figure 11 and those presented below is that aerosols affecting the hurricane clouds affect both TC structure and the intensity. One can see that the TC moving within polluted air at its periphery has a smaller intensity than that moving in clean air during the course of the entire simulation. This effect (see below) can be attributed to the convection invigoration at TC periphery, which was caused by turning off the warm rain (representative of the potential impacts of high concentrations of aerosols on the formation of droplets). As a result, some fraction of the air moving within the inflow layer ascends at the periphery instead of reaching the TC eye wall. Hence, the rate of the latent heat release within the eye wall decreases, which results in some increase in the central pressure.

Figure 10 shows the maximum values of the super cooled CWC above 5 km in simulations WR (with warm rain permitted) (left) and NWRP-30 (right) with no warm rain at the TC periphery. One can see a good qualitative correspondence of values obtained in
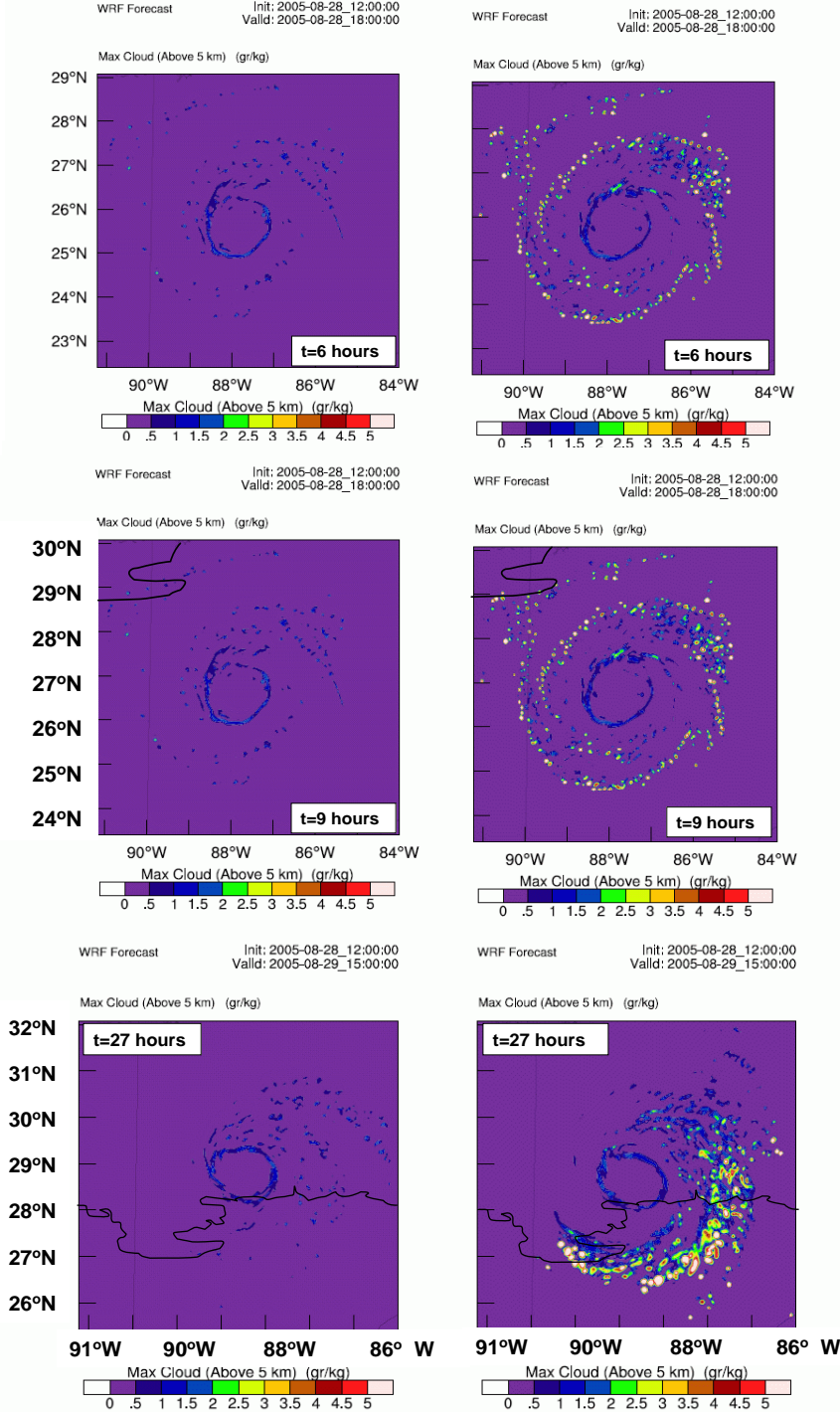
Figure 10. The maximum values of the super cooled CWC above the 5 km-level in the simulations with warm rain permitted (left) and no warm rain at the surface wind speeds exceeding 35 ms^{-1} at different time instances.

the 3-D simulation and those found in the 2-D individual cloud simulations: the CWC maximum in microphysically continental air is about 5 g/kg, while in the clean air it does not exceed 2 g/kg.

The amount of ice in the high AP concentration air is also larger than that in the clean air (not shown). The aerosol-induced increase in the total ice content at the TC periphery in agreement with the results of the 2D model with the spectral microphysics. The maximum values of the total ice content are also in good agreement with the 2-D results.

The difference between the condensate mass contents in the WR and NWRP-30 indicates the differences in the latent heat release and the vertical updrafts. The maximum vertical velocity in many small clouds arising in polluted air exceeds 10 ms^{-1}, while in the TC central zone the typical maximum vertical velocity ranges mainly from 4 to 10 ms^{-1}. Thus, aerosols increase the intensity of convection within the cloud bands located at the distance of 250-300 km from the TC center. This radius corresponds to the TC most remote cloud bands.

As it was discussed above, an increase of the CWC and the ice content in the zones of high vertical velocity should foster the lightning formation. **Figure 11** shows the fields representing the maximum values of products of updrafts, the CWC and the total ice content calculated in each grid point in a column. The maximum values of the products will be referred to as the lightning probability (LP). The two upper panels correspond to the time periods of the lightning rate depicted in Figure 1. One can see that the LP fields calculated in NWRP-30 resemble quite well the structure of the lightning observed: a) the maximum lightning in both cases takes place within a comparatively narrow ring of the 250-300 km radius; b) lightning (in the calculations
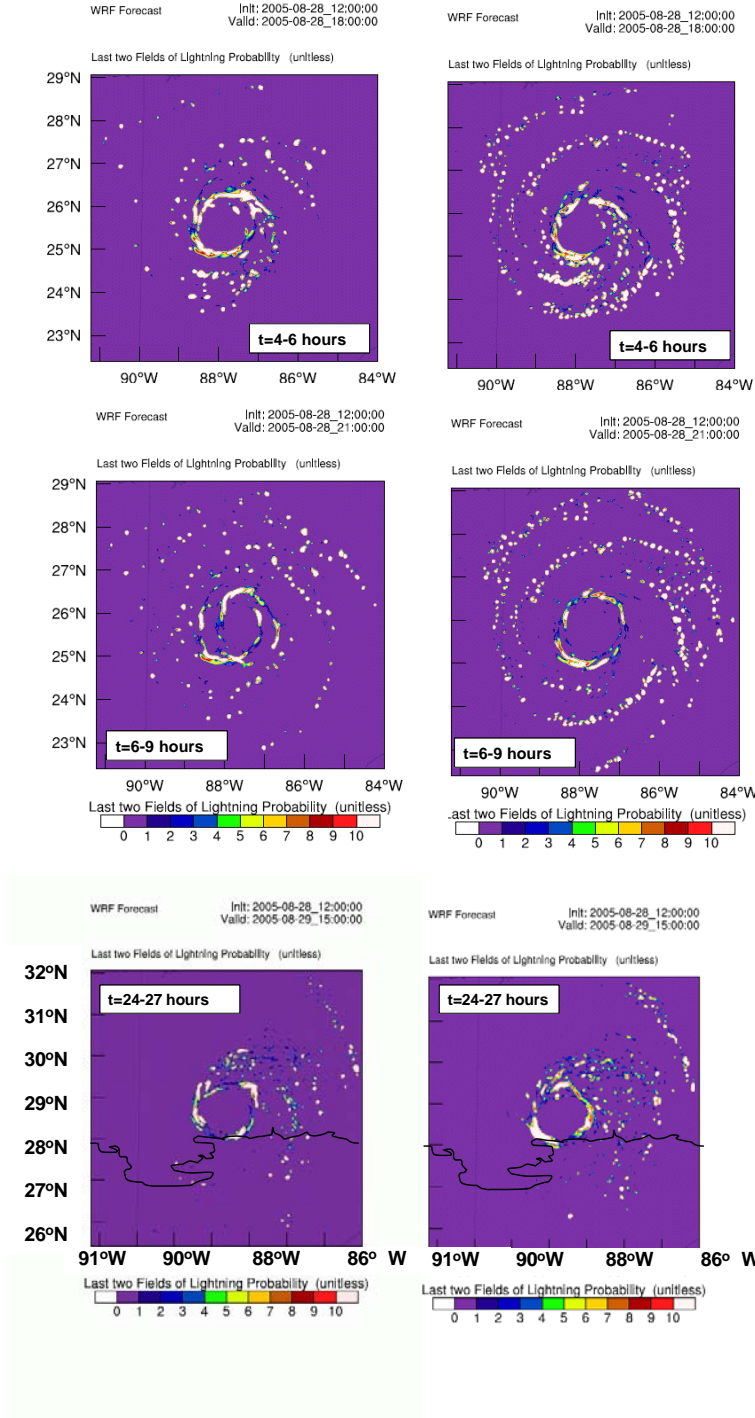
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Figure 11. Lightning probability (LP) at different time periods in the case of clouds developing in clean air (left) and continental air (right). The LP is calculated as the maximum product of updrafts, CWC and total ice contents. The LP plotted represents a sum of the LP fields calculated within the time ranges shown in the panels.

LP) at the TC central zone is as a rule weaker than in the rain bands at the TC periphery. Contrary to it, in the WR simulation lightning is much weaker and is concentrated in the eye wall (similarly to Fierro's et al 2007 results), which does not agree with the observations.The lower panels correspond to the time when the entire TC has penetrated into the land and rapidly decays. **Figure 11** indicates that while in clean air the lightning probability decreases over the land, it remains high in polluted air. Moreover, the lightning rate increases in the TC central zone. These figures show that aerosols significantly affect the spatial distribution of intense convection in TCs.

The results also indicate that aerosols affect the cloud structure, the intensity and spatial distribution of the precipitation of TCs approaching and penetrating the land (**Figure 12**).

4. DISCUSSION AND CONCLUSIONS

The potential effects of continental aerosols penetrating into the clouds (via the cloud bases) at the periphery of a TC approaching the land on the cloud structure and the lightning rate has been investigated using a 2-D cloud model with spectral bin microphysics and a 3D mesoscale WRF model with the bulk microphysics. Numerical experiments with the 2D cloud model with the resolution of 250 m and 125 m in the horizontal and vertical directions, respectively, show that the continental aerosols with the CCN concentrations of about 1000 cm^{-3} significantly increase the amount of supercooled cloud water, as well as ice (mainly ice crystals, graupel and hail) even under the high concentration of large CCN. In addition, the vertical updrafts were several m/s higher in the polluted clouds. All these factors taken together lead to the coexistence of ice and cloud water within a suppercooled cloud zone, which
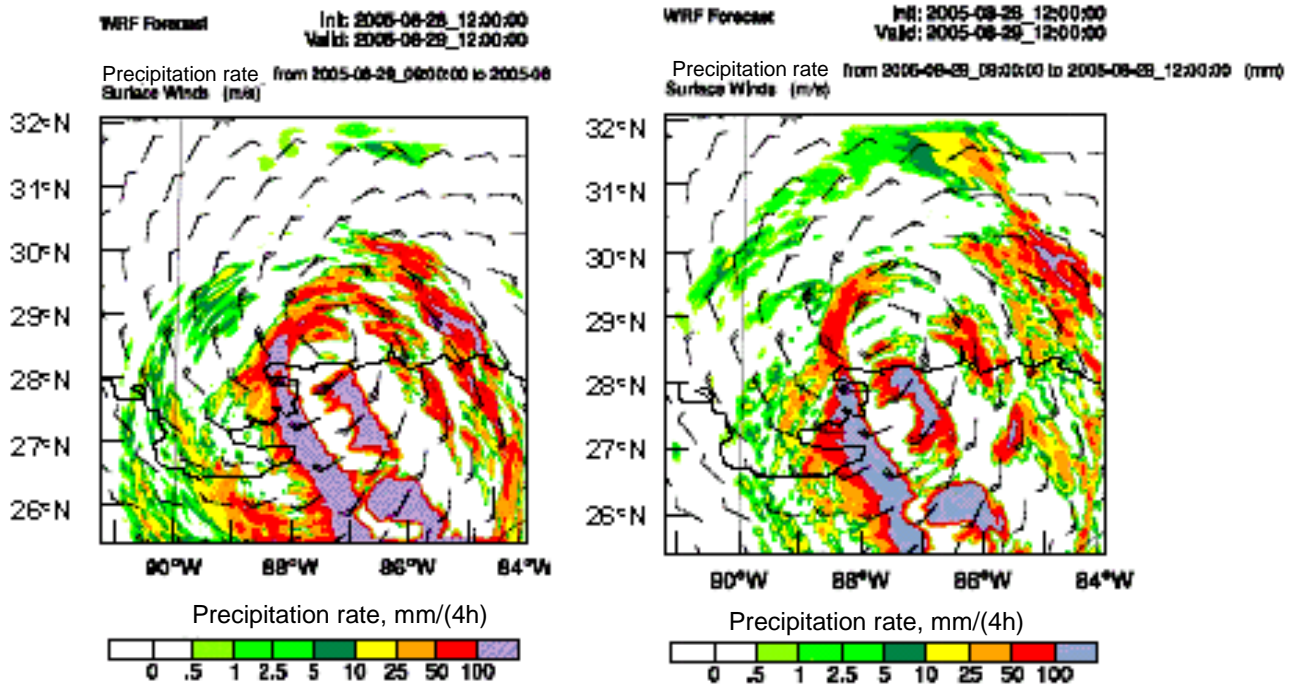
Figure 12. The field of the precipitation rate in the run with warm rain allowed over the entire computational area (WR-simulation)(left) and in the case when warm rain has been turned off at the TC periphery (NWRP-30) (right) during the TC landfall.

is considered to be favorable for charge separation and lightning formation.

The purpose of the simulations using a mesoscale 3-km resolution model was to investigate the possible effects of aerosols on lightning at periphery of hurricanes approaching the land as well as, on the cloud structure, precipitation and the TC intensity. The utilization of the 3-km resolution and crude vertical resolution, as well as the bulk-parameterization scheme used do not allow one to reproduce aerosol effects related to the fine balance between the fall velocity of growing droplets and the vertical velocity. For instance, although the results of the supplemental simulation with the droplet concentration N_d of 1000 cm^{-3} indicate some increase in central pressure and an increase in the supercooled CWC, but in general the results were quite similar to those obtained in the case when the droplet concentration was set equal to 30 cm^{-3}.

As a result, the aerosol effects were parameterized by "switching off" droplet collisions, and preventing warm rain at the TC periphery, where the surface wind speed was under 35 ms^{-1}. This simple parameterization of the aerosol effects leads to the results, which agree better with those, obtained using a spectral microphysical model with the 100-250 m grid resolution.

The product of the ice content, the supercooled cloud water content and the updraft velocity was chosen as the measure of the lightning activity (lightning probability, LP). It was shown that the LP field calculated in the model resembles quite well the structure of lightning observed: a) the maximum lightning takes place within a comparatively narrow ring with the radius 250-300 km; b) lightning in the TC central zone is, as a rule, weaker than that in the rain bands at the TC periphery. The LP minimum was found to be related to the suppression of the deep convection within the ring of the 50-150 km radius range. This minimum becomes more pronounced when the outermost rain bands are invigorated. In the simulation, where no aerosol effects were taken into account, the magnitude

of the LP parameter was smaller and concentrated in the eye wall, which does not agree with the observations.

The analysis of the intensity variations of simulated TCs, as well as the observed variation of the intensity of hurricanes Katrina (2005) and Rita (2005) (Fierro et al 2007), shows that the disappearance of lightning in the TC central zone and its intensification at the TC periphery can be a good indicator of the TC decaying. Such behavior of the TC lightning may be useful for a short range TC intensity forecast. Note that the simulations of TC lightning in an idealized TC performed by Fierro et al (2007) showed a negligible lightning activity at the TC periphery as compared to that in the TC eye wall. The instability of the atmosphere at the TC periphery was higher than in the TC center in those simulations. However, it did not lead to convective invigoration and lightning at the TC periphery. Thus, the comment of Molinari et al (1999) that the higher instability of the atmosphere at the TC periphery *cannot be a complete explanation"* of the lightning maximum at the TC periphery seems to be correct. We attribute the result obtained by Fierro et al (2007) to the fact that no aerosol effects have been taken into account in their simulations. Their results indicate that the changes of the instability of the atmosphere can lead to the variability of the lightning in the TC eye wall, but not within a narrow ring at the TC periphery. Hence, the comparison of the results obtained by Fierro et al (2007) and those in this study supports the assumption that lightning at the TC periphery is caused by the synergetic effect of continental aerosols and higher atmospheric instability.

The results also indicate that aerosols affect the cloud structure, the intensity and the spatial distribution of precipitation of TCs approaching and penetrating the land, or the TCs located in the air of continental nature containing a significant amount of aerosols. According to the results obtained, aerosol-induced convection invigoration at TC periphery leads to a TC weakening. From observations (e.g., Dvorak, 1984) and numerical results (e.g., Khain et al, 1984) it follows that TC formation and intensification takes place when the convection is concentrated in the TC center leading to latent heat release and the pressure fall there. According to the results of a great number of simulations with a TC model (Khain 1984) the intensification of convection at TC periphery decreases the TC intensity. This decrease can be attributed to the fact that the convection invigoration at the TC periphery leads to a) a decrease in the air mass penetrating to the TC eyewall and to a decrease the mass flux there. Correspondingly, the latent heat release also decreases in the eyewall. The second mechanism is as follows. Any deep cloud induces compensating downdrafts in its surrounding. As a result, deep clouds are usually surrounded by quite small ones, because of the competition. This competition between clouds was studied in several studies. The convection at the TC periphery affects the convection in the TC center in the similar way.

The mechanism of the TC weakening discussed in our paper resembles the hypothesis of the field experiment Storm Fury (Simpson and Malkus, 1964, Willoughby et al., 1985) according to which intensification of convection at the periphery of the eyewall should decrease air mass penetrating further toward the TC axis and weaken the wind speed maximum. However, in Storm Fury it was hypothesized that there is a significant amount of supercooled water at high levels in the eyewall. More detailed analysis (Willoughby et al., 1985) showed that there was too little supercooled water and too much hail in these clouds. Correspondingly, the cloud microphysical structure did not match the Storm Fury hypothesis. In our study the convection invigoration is caused by small aerosols. Such invigoration agrees both with the observations (e.g., Koren et al 2005) and the results of numerical cloud models with detailed microphysics

(Khain et al, 2005, 2008). A possible application of such finding to the TC mitigation by seeding is discussed by Rosenfeld et al. (2007b).

Zhang et al (2007) simulated the evolution of the idealized TC beginning with a weak initial vortex and found that the maximum TC intensity decreases when the aerosol concentration increases. We suppose that the reason of the TC weakening was similar to that found in this study: Zhang et al (2007) took aerosol scavenging into account, so the aerosol concentration was, supposedly, larger at the TC periphery as it was simulated in the present study.

Precipitation in the TC zone of the 350-400 km radius range decreases in the high AP concentration air mainly due to the convection weakening in the central zone of the TCs. At the same time precipitation at the TC periphery, as well as the precipitation area increase. More detailed simulations are required to conclude, whether continental aerosols increase or decrease precipitation from landfalling TCs over the land. We speculate that a weekly variation of the anthropogenic aerosols concentration can lead to a weekly cycle of intensity and precipitation of landfalling TCs. Note that the weekly cycle of the intensity and precipitation of landfalling TCs was reported first by Cerveny and Balling (1998). They attributed the variations to the weekly variations of the anthropogenic aerosol production. According to the hypothesis by Cerveny and Balling (1998), the aerosol loading increase leads to the increase in the solar heating in the boundary layer around a cyclone. In the present study we considered another mechanism that potentially can contribute to the weekly cycle of precipitation and the wind speed. Cerveny and Balling (1998) reported an increase in the precipitation under aerosol effects. Figure 12 indicates some possibility that such increase is caused by effects of aerosols on microphysical cloud structure. Note that the radiative effects of aerosols have not been taken into account in the present study. Thus, we cannot conclude what

mechanism is dominating. More investigations are required to make a definite conclusion.

A decrease in the TC intensity under the influence of the Saharan air was also reported by Dunion and Velden (2004). They attributed the decrease in the intensity to thermodynamical effects. We suspect that microphysical effect of the Saharan dust on TC convection may be also significant (see e.g. Zhang et al, 2007). Further investigations are required.

According to the results obtained using the 2-D cloud model, aerosols lead to an more significant increase of the CWC at higher levels than that it has been simulated using the WRF mesoscale model. We suppose, therefore, that the aerosol effect on cloudiness, precipitation and intensity of TCs may be more pronounced than that demonstrated in the study using the model with the 3-km resolution.

Note in conclusion that in spite of the encouraging and consistent results, this study must be considered as a plausible hypothesis as to how continental aerosols might affect hurricanes. Much more work must be done in the area of aerosol mapping and the trajectory analysis to formulate a consistent picture of TC/aerosol interactions. The utilization of high resolution models with spectral bin microphysics is desirable to make the results quantitative. More observational studies are required to investigate the microphysical structure (e.g., supercooled water, cloud ice contents) of clouds in TCs. Observational and numerical studies are also needed to determine aerosol properties (e.g., size distributions), as well as aerosol fluxes from the land to tropical cyclones.

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