Simulation of a landfalling hurricane using spectral bin microphysical model: effects of aerosols on hurricane intensity (the HAMP contribution)

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Abstract

The evolution of a super hurricane (Katrina, August 2005) was simulated using the Weather Research and Forecasting Model (WRF; version 3.1) with explicit (non-parameterized) spectral bin microphysics (SBM). The new computationally efficient spectral bin microphysical scheme (FAST-SBM) implemented to the WRF calculates at each time step and in each grid point size distributions of atmospheric aerosols, water drops, cloud ice (ice crystals and aggregates) and graupel/hail. The TC evolution was simulated during 72 hours beginning with its bypassing the Florida coast (27 Aug 2005) to its landfall just east of New Orleans, Louisiana (near the end of 29 August). The WRF/SBM was used to investigate the potential impact of aerosols ingested into Katrina's circulation during its passage through the Gulf of Mexico on Katrina's structure and intensity. It is shown that continental aerosols invigorated convection largely at TC periphery, which led to its weakening prior to landfall. Maximum weakening took place ~24 h before landfall, just after its intensity had reached its maximum. The minimum pressure increased by ~15 mb, and the maximum velocity decreased up to 15 m/s. Thus, the model results indicate the existence of another (in addition to a decrease in the surface fluxes) mechanism of weakening of TCs approaching the land. This mechanism is related to effects of continental aerosols involved into the TC circulation.

It is shown that aerosols substantially affect the spatial distribution of cloudiness and hydrometeor contents. The evolution of lightning structure within the TC is calculated and compared with that in Katrina. The physical mechanisms of aerosol-induced TC weakening are discussed.

Key words: tropical cyclones, spectral bin microphysics, cloud-aerosol interaction, numerical modeling

1. Introduction

1.1. Factors affecting TC intensity

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 prediction of TC intensity represents a difficult task. Well known factors affecting TC intensity are heat and moisture surface fluxes (which in turn are determined by the sea surface temperature (SST) and wind shear (e.g. Anthes 1982, Khain and Sutyrin, 1983; Khain 1984; Emanuel 2005). The implementation of TC-ocean coupling into prognostic TC models led to a significant improvement in the forecast of TC intensity (Falkovich et al, 1995; Bender and Ginis, 2000). The mechanisms mentioned above represent thermodynamical factors affecting the convection intensity, i.e. the availability of latent heat for release in cumulus clouds. During the past decade it was found that aerosols (including anthropogenic ones) substantially affect cloud microphysics and consequently the rate of latent heat release, the dynamics and precipitation (see, overviews by Levin and Cotton, 2008; Khain et al 2009; Rosenfeld et al, 2008). In particular, it was found that small aerosols invigorate tropical convection increasing vertical velocities and cloud top heights of deep convective clouds (Khain et al, 2004, 2005, 2008b; Koren et al 2005; Lynn et al, 2005a,b, Wang 2005; Lee et al 2008; Khain 2009). Thus, aerosols affect cloud microphysics and dynamics.

Until recently the possible effects of aerosols on TC intensity did not attract the attention of investigators. This can be attributed to the lack of sophisticated microphysical schemes in most current TC models. Besides, both experimental and numerical investigations of cloud microphysics in TCs are quite limited. Microphysical observations in TCs are usually limited by the zones of the melting level (McFarquhar and Black 2004). The lack of knowledge of the microphysical structure of clouds in TCs led in the past to the failure of the hurricane mitigation project STORMFURY (Simpson and Malkus, 1964, Willoughby et al., 1985) based on the hypothesis that TC weakening can be achieved by glaciogenic seeding of cloud tops at the periphery of the eyewall. It was hypothesized that there is usually 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 ice in these clouds (Black and Hallett 1986). Correspondingly, the cloud microphysical structure did not match the Storm Fury hypothesis (see Rosenfeld et al 2007; Khain et al 2008a for detail). A numerical study by Khain and Agrenich (1987)-was probably one of the first ones where a possible effect of the Saharan dust on the TC development via heating of polluted air by solar radiation was investigated using an axisymmetric TC model. Some observations indicating possible aerosol effects on TC development via their influence on the cloud microphysics of TC clouds have been performed recently by Jenkins et al (2008) and Jenkins and Pratt, (2008).

1.2 Lightning and microphysical processes

An indirect evidence of aerosol effects on TC intensity can be derived from enhanced lightning in TC. It is well known that lightning over the sea is a much rarer phenomenon than over the land (e.g. Williams and Satori, 2004; Williams et al, 2005). Charge separation needed for lightning formation is a result of collisions between low density ice crystals with high density graupel in the presence of a significant amount of supercooled water at temperatures below -13 ^{o}C (e.g., Takahashi, 1978; Saunders 1993; Black et al 1996; Black and Hallett 1999; Cecil et al 2002a,b; Sherwood et al 2006, Betz et al 2008). Such microphysical structure is not typical of maritime convection in clean air where efficient drop collisions below freezing level lead to the formation of raindrops falling down without freezing (warm rain). As a result, maritime clouds contain usually very low amounts of supercooled water. Thus, the formation of intense lightning in a TC indicates substantial changes in the cloud microphysical structure as compared to that of clouds in TC surroundings. Note that enhanced lightning intensity over the oceans takes place downwind of continents. It is reasonable to hypothesize that this increase in lightning takes place in zones of aerosols intrusion from continents. An increase in lightning rate indicates invigoration of convection (stronger updrafts with increasingly larger volumes of graupel or small hail/frozen drops aloft), and an increase in 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 (Lyons and Keen, 1994; Orville and Coyne 1999, Molinari et al 1999; Shao et al 2005; Demetriades and Holle, 2006; Fierro et al 2007, Price et al 2009). Rodgers et al (2000) found that the closer the lightning is to the storm center, the more likely the TC is to intensify. Molinari et al (1994, 1999) analyzed the radial distribution of lightning in hurricanes approaching the US coast using the National Lightning Detection Network data. 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. The lightning activity in hurricane Katrina (2005) at several successive time instances was presented by Shao et al (2005). An analysis of the location of lightning within the zone of Katrina indicates that the appearance of strong lightning in the eyewall of Katrina was a precursor of the TC deepening, while the disappearance of lightning in the TC center and appearance of intense lightning at the TC periphery was the precursor of TC weakening. Usually enhanced lightning at the TC periphery is attributed to atmospheric instability (e.g. Molinari et al., 1994, 1999). Lightning in the TC eyewall is often attributed to instabilities that result in "convective bombs" or explosive convection in the eye wall that produces the large updrafts and conditions suitable for lightning. Khain et al (2008a) analyzed the mechanisms of lightning formation in Katrina (as well as in other landfalling hurricanes) in more detail. They showed that a penetration of continental aerosols to clouds at the TC periphery and successive convection invigoration is the mechanism contributing substantially to lightning formation at the TC periphery. An increase in concentration of small aerosols increases droplet concentration and decreases droplet size. The net effect is the decrease in the collision rate, delay in raindrop formation and warm rain production. As a result, small droplets ascend in cloud updrafts and continue growing by condensation. It leads to an increase in supercooled water content, which intensifies the riming, i.e. ice-water collisions accompanied by freezing of liquid water. Both processes (diffusion drop growth and freezing by riming) are accompanied by extra latent heat release leading to increase in cloud updrafts and sometime to increase in cloud top height (see overview by Khain 2009). Aerosol-induced increase in supercooled cloud water content (CWC) and vertical velocities foster the formation of conditions favorable for charge separation and lightning formation, when collisions of ice crystals and graupel take place in the presence of supercooled droplets.

1.3 Modeling with microphysical treatment

Khain et al (2008a) simulated the evolution of hurricane Katrina (August 2005) during its movement in the Gulf of Mexico using a two nested grid Weather Research Model (WRF, NCAR version). The resolution of the finest and of the outer grid was 3 km and 9 km, respectively. The Thompson et al. (2006) one-moment bulk-parameterization was used to describe microphysical processes in clouds and the corresponding latent heat release. Parameters determining autoconversion rate were calculated in this scheme by presetting the droplet concentration. Supplemental simulations with droplet concentration varying from $100\,cm^{-3}$ to 1000 cm^{-3} showed, however, that the scheme is not sensitive to the aerosol concentration. To "sidestep" this problem, Khain et al (2008a) simulated the effects of continental aerosols by preventing warm rain in the scheme altogether by shutting off the drop-drop collisions only at the hurricane periphery. A similar approach was used by Rosenfeld et al (2007). The results obtained in these idealized simulations led to an increase in supercooled water and ice contents in the rainbands at the TC periphery and allowed Khain et al. (2008a) to conclude that continental aerosols that penetrated the TC periphery caused enhanced lightning flashes in the areas of penetration. It was also shown that aerosols, invigorating clouds at 250-300 km from the TC center, decrease the convection intensity in the TC eyewall leading to some TC weakening. Similar results were reported by Rosenfeld et al (2007), who proposed a method of TC mitigation by seeding of clouds at the TC periphery near their cloud base with small aerosol particles of 0.05 μm to 0.1 μm in radius.

Simulations of the evolution of an idealized TC using Regional Atmospheric Meteorological System (RAMS) (Zhang et al, 2007) supported the conclusion that aerosols (for instance, Saharan dust) can substantially affect the intensity of TCs.

To take into account microphysical factors (such as aerosols) properly, advanced microphysical schemes are required. At the same time the problem of the adequate description of convection in TC models, remains one of the most difficult problems in the TC modeling and forecasting. The current operational TC forecast model developed at the Geophysical Fluid Dynamics Laboratory used large scale convective parameterizations till 2006 (Kurihara 1973, Arakawa and Shubert 1974). Since 2006 this model has used a simplified Arakawa-Schubert scheme for its cumulus parameterization, and a simplified version of the Ferrier bulk-parameterization (Ferrier 2005) to calculate large-scale condensation under conditions of supersaturation (Bender et al 2007). The simplified bulk scheme treats only the sum of the hydrometeor classes (referred as the total condensate) in the advection in both horizontal and vertical direction. Both schemes are insensitive to aerosols.

The development of one and two-moment bulk parameterization schemes and their application in mesoscale models such as the Penn State/NCAR Mesoscale Modeling System Version 5 (MM5) (Dudhia et al., 1993), RAMS (Pielke et al, 1992) and the Weather Research Forecast Model (Skamarock et al, 2005), the operational numerical weather prediction model of the German Weather Service (COSMO) combined with an extended version of the 2-moment bulk scheme by Seifert and Beheng (2006), etc. was an important step toward the improvement in the description of convective processes and precipitation in numerical models. A priori prescription of the shape of size distribution functions of different cloud hydrometeors (such as cloud droplets, rain drops, graupel, aggregates) in the form of exponential Marshal-Palmer distributions or gamma distributions reduces the system of equations that describes cloud microphysics to a relatively small number of equations for integral quantities such as mass contents (one moment schemes) and mass contents and concentrations (two moment schemes). The comparatively small number of prognostic equations makes the schemes computationally efficient, so they are widely utilized in simulating different cloud-related phenomena such as supercell storms, squall lines. Recently these schemes were used for simulation of TCs (e.g., Zhang et al, 2007; Fierro et al, 2007).

The bulk-parameterization schemes, especially one-,moment schemes have substantial limitations in describing microphysical processes affecting the shape of size distributions (such as aerosol effects). Besides, these schemes as a rule do not solve the equation for diffusion growth of drops (this equation is replaced by transformation of all supersaturated water vapor into cloud water mass, so that the final supersaturation is assumed equal to zero). Instead of solving the stochastic equation of collisions, all one moment and most two-moments bulk schemes use semi-empirical relationships for autoconversion rates often with hidden or internal assumptions about cloud droplet number concentrations. A substantial shortcoming of most bulk schemes is utilization of a single settling velocity and single collision efficiency for an entire distribution of cloud droplets and given ice species*.

The second approach to simulate microphysical processes is the utilization of spectral bin microphysics (SBM), in which a system of kinetic equations for size distributions of particles of different classes is solved. Each size distribution function is described using several tens of mass (size) bins. The equation system solves the equations for advection, settling, collisions, freezing, melting, etc. for each mass bin (each particle size). The SBM describes aerosol effects on cloud microphysics and dynamics taking into account the aerosol budget, i.e. a limited source of aerosol particles. This method is much more accurate than the bulk-parameterization as regards its ability to simulate cloud dynamics and microphysics and

precipitation [see comparisons of SBM vs bulk schemes in Lynn et al., 2005b, Lynn and Khain 2007; Li et al. 2009a,b, Iguchi et al., 2008; Khain and Lynn, 2009; Khain et al 2009]. As was mentioned by Khain et al (2008a), the utilization of high resolution TC models with SBM could grant greater credibility to simulations of TC and the effects of aerosols on TCs. However, the basic SBM scheme described by Khain et al all (2004, 2008) requires ~50 times more computer time than standard one-moment bulk-parameterization schemes, which hinders their wide application in mesoscale models, and in particular, in TC models.

In this study the evolution of hurricane Katrina over the Gulf of Mexico is simulated with WRF, in which cloud microphysics is described using a computationally efficient spectral bin microphysics scheme, in which all microphysical processes are described explicitly.

2. Model and experimental design

1. Spectral bin microphysics scheme

The SBM scheme implemented into the WRF (Skamarock et al., 2005, Version 3) has been described by Khain et al (2004) and Lynn et al (2007). The scheme is based on solving the kinetic equation system for the size distributions of seven classes of hydrometeors: water drops, three types of crystals (columnar-, plate- and branch-type), aggregates (snow), graupel and hail. Each hydrometeor class is described by a size distribution function defined on the grid of mass (size) containing 33 mass bins. A doubling mass grid is used, so that the mass of drops belonging to the (i+1)-th bin is twice as large as the drop mass in the i-th bin. The mass grids are similar for all hydrometeors to simplify the transition from one type to another during freezing, melting, etc. The minimum particle mass corresponds to that of the $2\,\mu m$ radius droplet. The model is specially designed to take into account aerosol effects on cloud microphysics. It contains an aerosol budget, which describes two-way cloud-aerosol interaction. Aerosol particles are also described by a size distribution function containing 33 size bins. In contrast to the standard bulk parameterization schemes the size distributions of cloud hydrometeors and aerosols are not prescribed a priori, but rather calculated in the course of the model integration. Supersaturation is calculated using an accurate analytical method (Khain et al 2008b) representing an extension of the approaches developed earlier by Tzivion et al. (1989) and Khain and Sednev (1996). Using the values of supersaturation, the critical size of aerosol particles to be activated to drops is calculated. Aerosol particles exceeding the critical size are activated and the corresponding mass bins in the aerosol size distribution become empty. It means that within the cloud updraft we have two fluxes: the flux of droplets and flux of non-activated CCN. When supersaturation exceeds its local maximum at the cloud base, a new portion of CCN will be activated to droplets (in-cloud nucleation). This process is especially important in maritime clouds where supersaturation is high and often increases with height because of a decrease in droplet concentration and increase in vertical velocity with height. The inclusion of this process leads to the formation of bi-modal droplet spectra with realistic droplet spectra dispersion (Pinsky and Khain, 2002; Segal and Khain 2006, Segal et al, 2003). The SBM also takes into account possible droplet nucleation during dry air entrainment through the lateral cloud boundaries.

A new approach has been applied to eliminate artificial spectrum broadening typical of previous spectral microphysical schemes (including the earlier version of this model, see Khain et al 2000). In this method, the remapping of size distribution functions obtained after diffusion growth to the regular mass grid conserves the three moments of size distribution (zero, third and six moments) to prevent artificial formation of large sized tails in the drop distribution (Khain et al, 2008b). This approach allows the formation of very narrow droplet spectra found in smoky and pyro-clouds measured during biomass burning over Brazil (Andreae et al 2004). Note that in the SBM, water drops are not separated artificially into cloud water and rain water (in contrast to in all bulk-parameterization schemes). It means that the SBM does not separate the collision process of water drops into accretion (collisions of cloud droplets) and collection (collisions of rain drops and cloud droplets), which is performed in all bulk schemes under the simplification of continuous growth. Instead, in SBM the cloud particle collisions are calculated by solving the stochastic kinetic equations for collisions. 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 collision kernels for each pair of particles are calculated using accurate superposition method (Pinsky et al, 2001, Khain et al 2001) and used in the form of lookup tables. The collisions kernels are calculated taking into account the particle shape and density that are represented as the functions of particles mass following Pruppacher and Klett (1997) (see Khain and Sednev 1996, Khain et al 2004 for details). The ice nuclei activation is described using an empirical expression suggested by Meyers et al. (1992) and applying a semi-lagrangian approach (Khain et al 2000) to allow the utilization of the proposed diagnostic formulas in a time dependent framework. Secondary ice generation is described according to Hallett and Mossop (1974). The rate of drop freezing follows the observations of immersion nuclei by Vali (1974, 1975), and homogeneous freezing according to Pruppacher (1995). Breakup of raindrops is described following Seifert et al (2006).

The SBM model does not requires any tuning of the scheme parameters and was successfully used without any changes for simulation of deep maritime convection (Khain et al, 2004, 2008b), continental clouds including pyro-clouds (Khain et al, 2008b), squall lines (Lynn et al, 2005a,b; Tao et al 2007; Li et al 2009a,b; Khain et al, 2009), supercell storms (Khain and Lynn 2009) and arctic stratiform clouds (Fan et al, 2009).

Note that the treatment of 8 size distributions (advection of all bins, collisions between particles belonging to different bins) require a significant computer time, which is about 50 times longer than that required by a standard one-moment bulk-parameterization scheme (in a 3-dimensional simulation).

To reduce computer time, a Fast-SBM has been developed and applied in the study. In the Fast-SBM all ice crystals and snow (aggregates) are calculated on one mass grid (one distribution function). The smallest ice crystals with sizes below 150 μm are assumed to be crystals, while larger particles are assigned to aggregates (snow). Similarly, high-density particles (graupel and hail) are also combined into one size distribution (graupel). No changes in the description of microphysical processes compared to the Full-SBM have been made. As a result, the number of size distributions decreases from 8 to 4 (aerosols, water drops, low density ice, high density ice). Note that Fast-SBM keeps the main advantages of SBM: a kinetic equation system is solved using the non-parameterized basic equations, particles of each size have their own settling velocity, particles depending on their mass have different densities, etc. The test simulations showed that Fast-SBM requires less than 20% of the time of the full SBM, which makes it possible to use the Fast-SBM on standard PC-clusters.

The detailed comparison of results obtained by the Full and Fast SBM was described by Khain et al (2009), where a tropical squall line was simulated using both SBM schemes. It was shown that the Fast SBM produces the microphysical and dynamical structure of the squall line as well as accumulated rain at the surface quite similar to those simulated with Full SBM. We suppose that for the simulation of tropical maritime deep convection it is not so important to reproduce many types of ice crystals, as well as specific properties of hail, because deep maritime clouds contain comparatively small amount of such hydrometeors.

2.2 Experimental design

A set of simulations were used to study possible aerosol effects on the evolution of Hurricane Katrina (August 2005) in the Gulf of Mexico during about three days (beginning with 27 August 00 UTC) prior to and including landfall (on about 12 UTC 29 August). A two nested gridded WRF (version 3.1) was used, and the nest moved using a cyclone-following algorithm. 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 SBM is applied at the finest grid of size 400 x 400 km. On the outer grid, the bulk-parameterization of Thompson et al. (2005) was used. Integral parameters of clouds (mass contents) penetrating from the outer grid into the fine grid were recalculated into size distribution functions assuming a Marshall-Palmer size distribution as defined in the bulk-parameterization scheme. If clouds formed in the internal area penetrated the outermost grid, the size distribution functions of hydrometeors in these clouds are used to calculate mass contents used by the bulk parameterization.

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 temperature was initialized on 27 August 00 UTC, and was not updated during the experiments. According to the reanalysis data the SST taken along the TC track reached its maximum near the shore (the place of the TC landfall).

Cloud droplets arise on aerosol particles (AP) playing the role of Cloud Condensational nuclei (CCN). The initial (at t=0) CCN size distribution is calculated using the empirical dependence of concentration of activated CCN N_{ccn} at supersaturation with respect to water S_w (in %):

$$N_{ccn} = N_o S_w^k \quad (1)$$

This method for calculating initial AP size distribution is based on the Kohler theory and described by Khain et al (2000) in detail. N_o and k are the measured constants for determining the AP concentration and shape of the AP size distribution. At t>0 the prognostic equation for the size distribution of non-activated AP is solved. The initial AP concentration was assumed constant within the lowest 2 km layer and decreased exponentially with height with characteristic scale of 2 km. Aerosols were transported over the entire computational area similarly to other scalars like the mixing ratio.

To investigate aerosol effects on microphysics and the dynamics of the TC two simulations were carried out: a) in the first “MAR” simulation N_o was set equal to 100, cm^{-3} typical of maritime atmosphere over the whole computational area; b) in the second, semi-continental MAR_CON case the initial CCN concentration over the land N_o was set equal to 1500 cm^{-3}, typical of continents under not very polluted conditions. Initially, over the sea N_o was set equal to 100 cm^{-3} in all simulations. In all simulations the slope parameter k was set equal to 0.5. When the TC entered the Gulf of Mexico, its circulation transported aerosols from the land to sea, so that some continental aerosols penetrate clouds within the TC and affect their microphysics and dynamics. The impact of continuing emissions was taken into account by continuing a flux of aerosols into the boundaries of the coarse domain. Corresponding comments are included into the revised paper. In all simulations the maximum size of dry AP was equal to 2 μm, which gave rise to droplets of radius 8 μm at cloud base. No giant CCNs that could arise at high winds as a result of spray formation were assumed in the simulations. The role of CCN with sizes of 2 μm was investigated in detail by Khain et al (2008a). It was shown that in the presence of large concentrations of small aerosols concentration of droplets is high and supersaturation is low. As a result, droplets forming on large CCN grow slowly and do not affect substantially the microphysical and dynamical structure of clouds. The role of giant CCN may be significant in the central zone of TC and will be investigated in a future study.

Note that there are some uncertainties as regards the concentration and size distributions of continental AP that may form over the land under strong winds. A concentration of 3000 cm^{-3} (instead of 1500 cm^{-3} used) seems also to be realistic. In this sense our simulations can be considered as a sensitivity study of TC to continental aerosols involved by TC circulation.

3. Results of simulations

3.1 Effects of aerosols on minimum pressure in TC

Figure 1 shows the time dependence of minimum pressure in all simulations and in Katrina. One can see that the modeled TC had lower intensity during the first ~50 h of simulations as compared to that of Katrina.

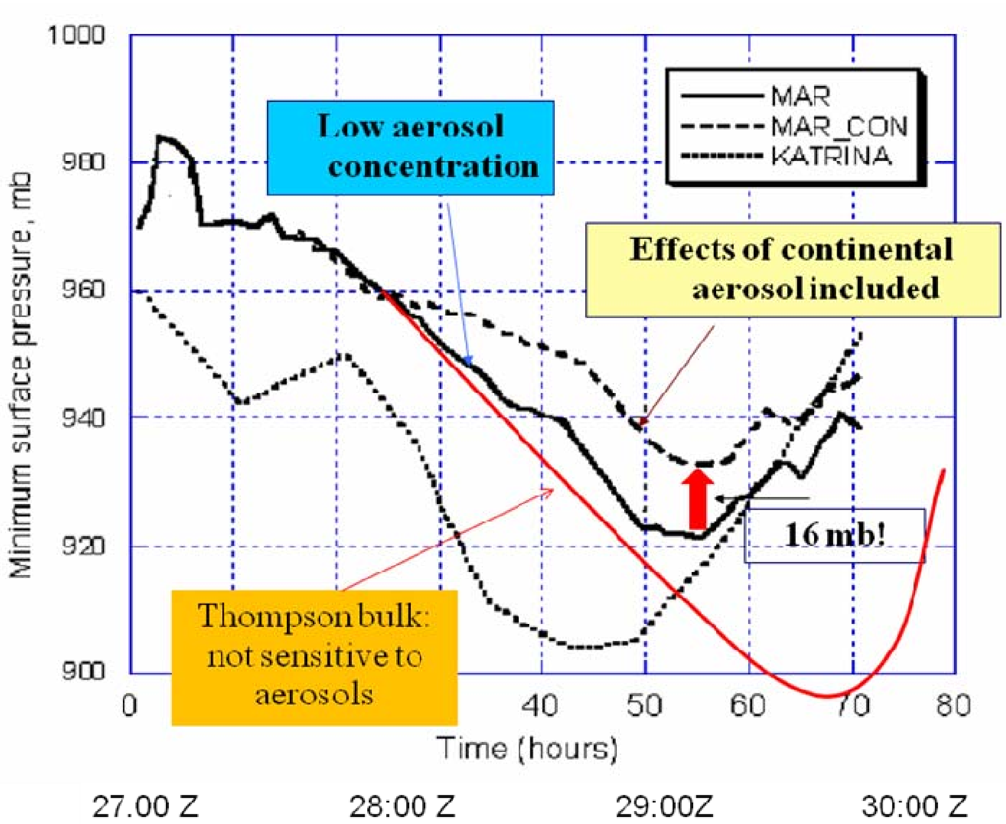

Figure 1. Time dependence ofminimum pressure in numerical experiments and hurricane Katina (August 2005)

Note in this connection that the WRF model used did not have specific adjustment procedures typically used in the TC forecast models to adopt the TC structure. In our case the initial data were derived from the crude resolution (100 km) reanalysis data at t=0 (27 Aug 00 UTC). Hence, some relaxation period was required to get the model TC intensity close to the observed one. Yet, 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 TC intensity and structure in the simulations with and without aerosol effects on the TC clouds in a strong hurricane, which is able to ingest aerosols from the continent. Figure 1 shows that TC in the "MAR_CON run turned out substantially weaker, so that at the time instances when the TC reached its maximum intensity the minimum pressure in its center was about ~16 mb higher than in the MAR run. Note that lower (as compared to Katrina) intensity of the model TC leads likely to an underestimation of aerosol effects, because of a weaker TC transports lower

AP amounts into the TC circulation. Results shown in Figure 1 indicate that the aerosol effect is an important factor affecting the TC intensity.

3.2 Aerosol effect on microphysical and dynamical fields

In the analysis of the aerosol fields we addressed two main questions. The first one was: whether a significant aerosol concentration can enter the TC periphery when it is located at comparatively large distance from the coast line, and second, whether aerosols can penetrate the TC eye. Aerosol fields simulated in the MAR run (not shown) indicate a very uniform distribution of AP concentration (which is very low) because the AP concentration over land was assumed equal to that over the sea. **Figure 2** shows the fields of the vertical velocity and cloud top height in MAR and MAR_CON**.**

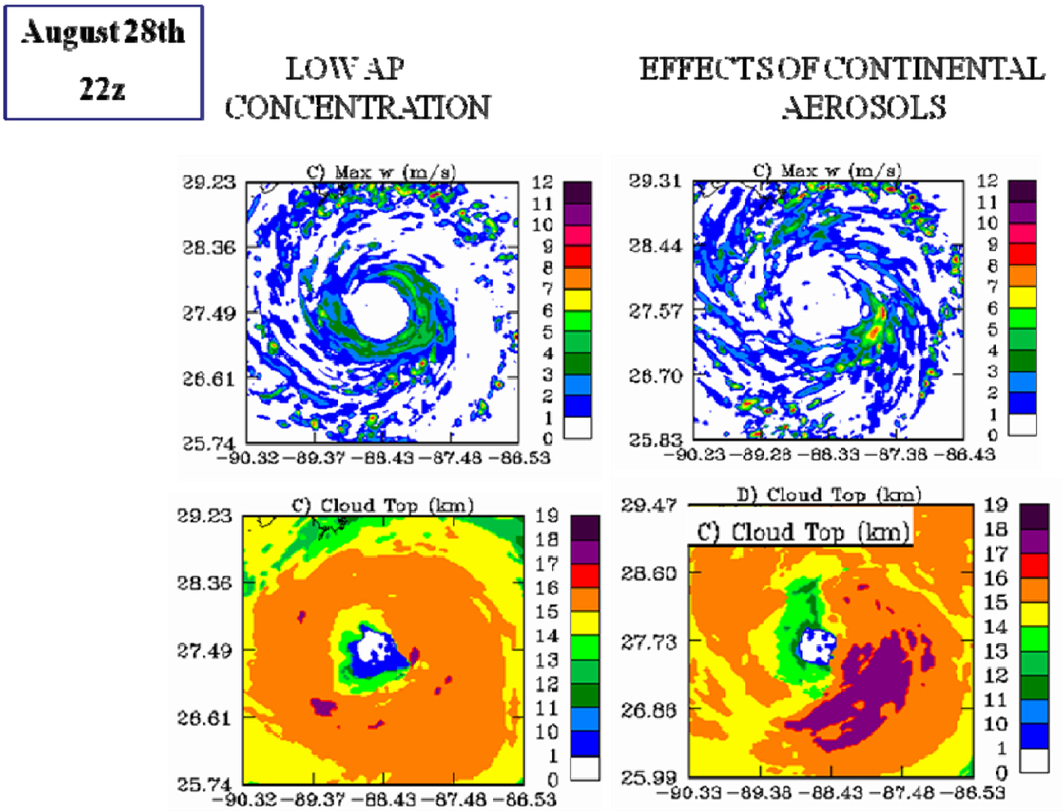

Figure 2. Fields of vertical velocity and cloud top height. One can see that W is higher in MAR-CON, especially in zones of high AP concentration. Cloud top height is also higher in MAR_CON. These results reflect the fact that aerosols invigorate convective clouds at TC periphery.

Figure 3 compares the fields of the column-maximum droplet concentrations (upper row) and cloud water mass content (CWC) (clouds with radii below 40 μm) in clouds in simulations MAR (left) and MAR-CON (right) at August 28th 22 UTC on the fine grid (46h, Figure 1).

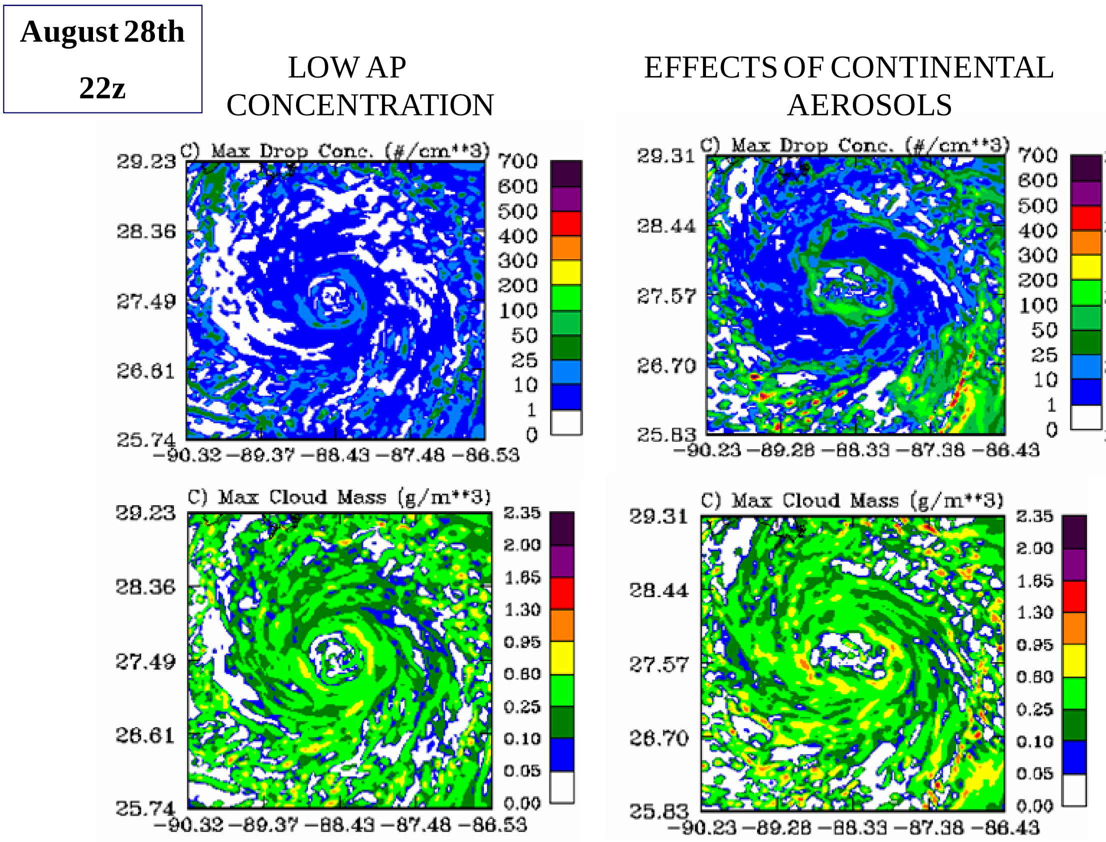

***Figure 3** Fields of the column-maximum droplet concentrations (upper row) and cloud water mass content (CWC) (clouds with radii below 40 μm) in clouds in simulations MAR (left) and MAR-CON (right) at August 28th 22 UTC on the fine grid (46h, Figure 1).*

One can see that in the MAR run droplet concentration does not exceed 50-100 cm^{-3}, which is typical droplet concentration in clouds arising in clean maritime air. Zones of maximum droplet concentration in the MAR run at the TC periphery indicate zones of higher vertical velocities in rain bands. In MAR-CON, the penetration of continental aerosols led to an increase in droplet concentration at the TC periphery in the zone of high aerosol concentration, as well as in the eyewall. In the MAR-CON run the maximum droplet concentration reached 500 cm^{-3} (especially high concentrations are at the TC periphery), which is substantially higher than those in typical maritime clouds. An increase in droplet concentration within the eyewall in the MAR-CON run indicated that aerosols penetrated to the TC eyewall in the simulations.

The CWC dramatically increased (mainly supercooled water content) as a result of aerosol penetration: while the maximum CWC reached 0.6 gm^{-3} in the MAR run, the CWC in the MAR-CON run exceeded 1.6 gm^{-3}. The CWC increased largely at the TC periphery where concentration of aerosols was higher. To illustrate the difference in the structure of CWC in MAR and MAR_CON, **Figure 4** shows the cross-section of azimuthally averaged CWC in these runs at the time instance when the maximum difference in the TC intensities took place. One can see that while small droplets in MAR reach about 3.5 km and produce warm rain, small droplets in MAR-CON reach ~8 km level indicating the formation of a high amount of supercooled water. One can see that aerosols led also to increase in the radius of the eyewall that usually indicates TC weakening.

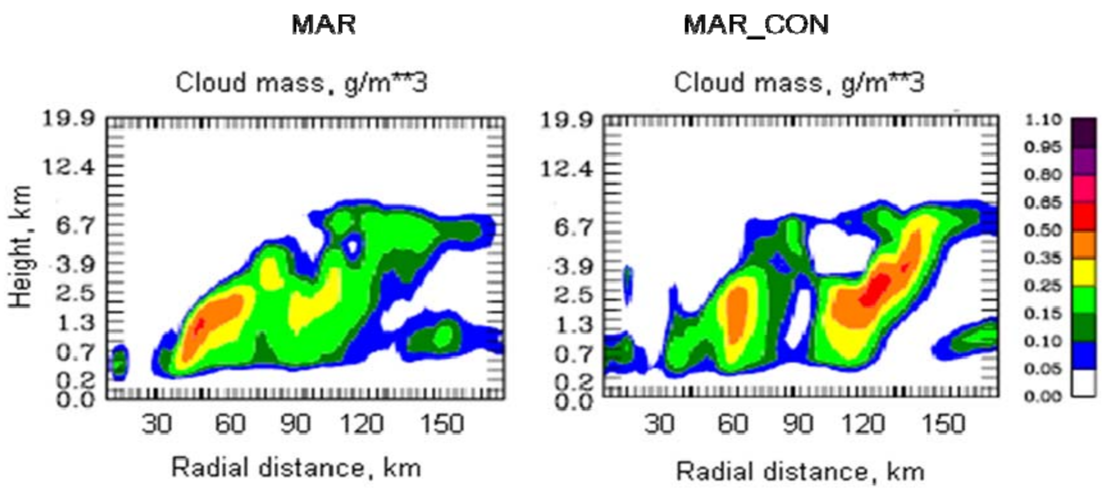

Figure 4 *The cross-section of azimuthally averaged CWC in these runs at the time instance when the maximum difference in the TC intensities took place.*

The aerosol-induced changes in warm microphysics resulted in corresponding changes in ice microphysics. The penetration of larger amounts of drops above the freezing level led to an increase in graupel and snow (aggregates) contents at the TC periphery (**Figure 5**).

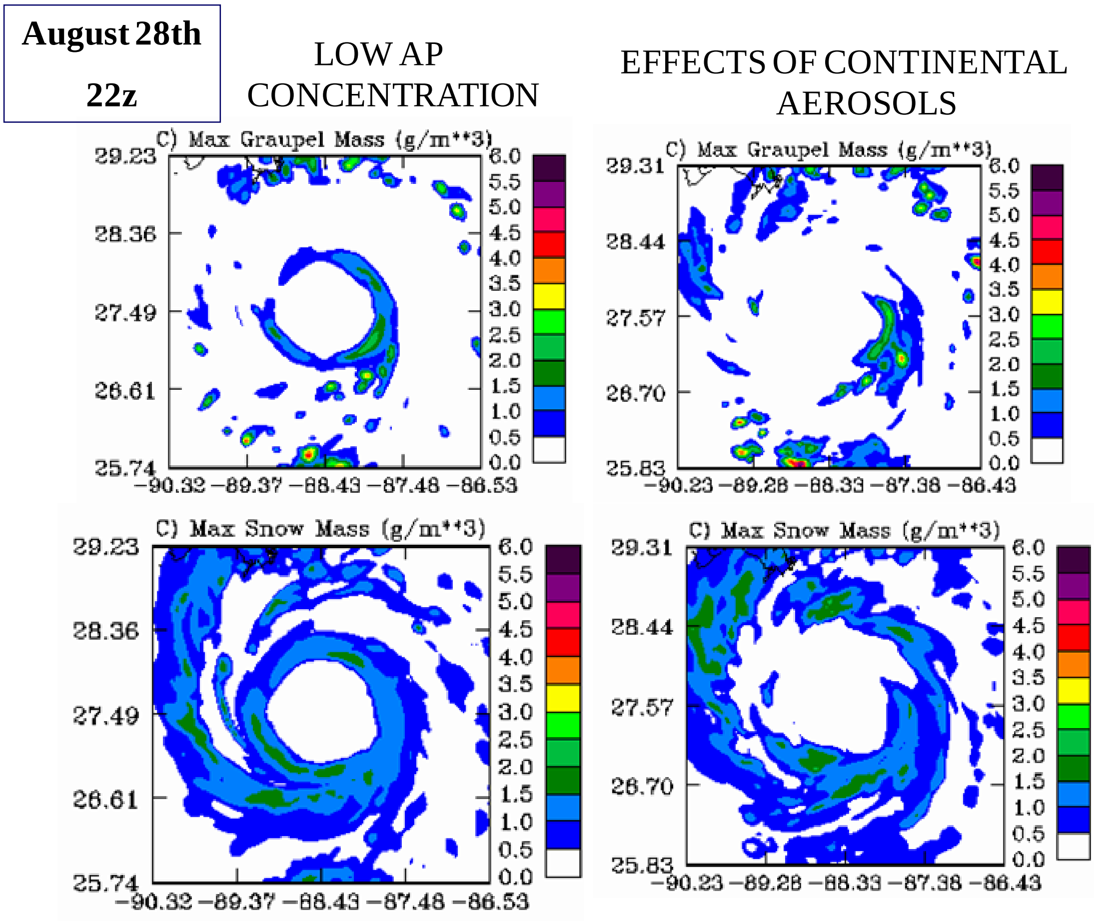

Figure 5. Fields of maximum values of graupel and snow contents in clean air and in polluted air (right)

Note that convection invigoration of clouds at the TC periphery weakened the updrafts in the eyewall, which immediately resulted in the decrease and even disappearance of graupel and snow in the eyewall. The values of maximum vertical velocities exceed 10 m/s which is a quite rare situation for maritime TC clouds (Jorgensen et al, 1985; Jorgensen and LeMone 1989). At the same time namely such high velocities are required to form lightning. The increase in cloud top height within polluted air was observed from satellites (Koren et al,

2005) and simulated in many recent studies dedicated to aerosol effects on cloud dynamics (see review by Khain 2009).

Figure 6 shows fields of the maximum wind speeds in the MAR (left) and MAR_CON (right) runs at the same time instances of 11 UTC, and 12 UTC. One can see that the maximum wind speed in MAR-CON is substantially lower than in MAR.

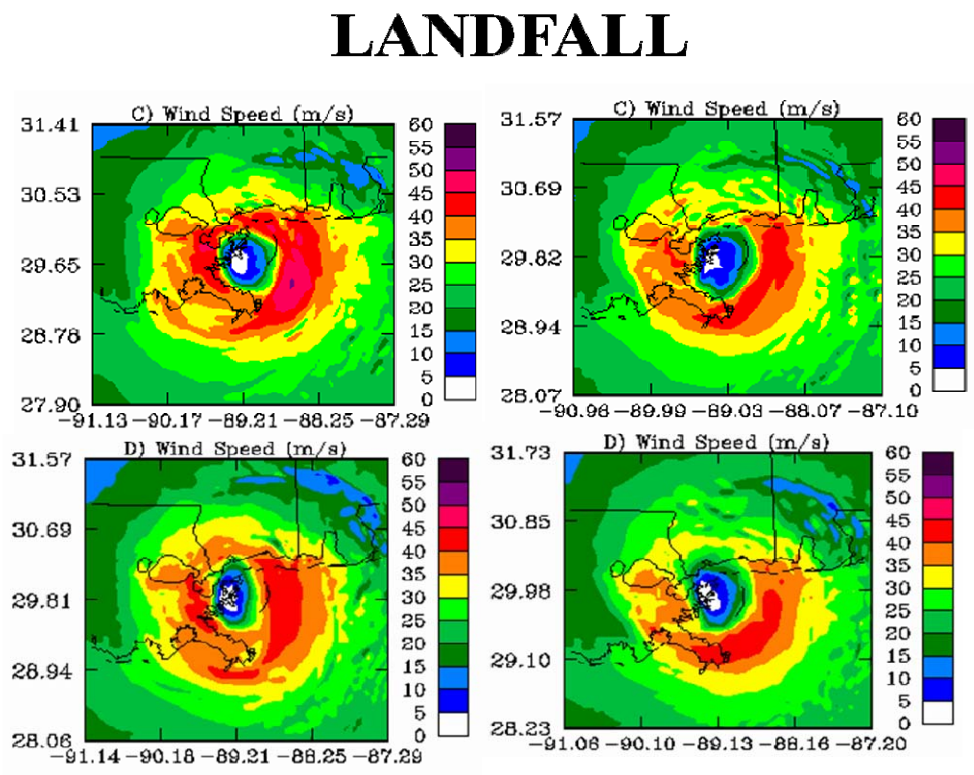

Figure 6 *Fields of the maximum wind speeds in the MAR (left) and MAR_CON (right) runs at the same time instances of 11 UTC, and 12 UTC. One can see that the maximum wind speed in MAR-CON is substantially lower than in MAR during landfall.*

Besides, the area of strong wind is also substantially lower in MAR-CON. Note the existence of a sharp decrease in the maximum wind speed over the land. We attribute this decrease to the effect of surface friction. These figures indicate that the aerosol –induced TC weakening, which was the strongest when TC was located several hundred km offshore, remains significant during landfall. Continental aerosols change the spatial distribution of wind, precipitation, lightning and other important parameters of landfalling TCs.

4. Discussion and conclusions

4.1 The main results

For the first time tropical cyclone evolution was calculated using explicit (non-parameterized) spectral bin microphysics (SBM). Simulations with resolution of 3 km, were made with the WRF/ SBM. The evolution of Katrina was simulated during 72 hours beginning after it had just bypassed Florida to 12 hours after landfall. These simulations were used to investigate the effects of continental aerosols ingested into its circulation TC on the TC structure and intensity. It is shown that at distances of a few hundred kilometers to the TC center aerosol concentration becomes similar to that over the land. It also was shown that some fraction of aerosols penetrated clouds in the eyewall affecting their micorphysical processes within the eyewall.

It is shown that continental aerosols invigorate convection (but largely at the TC periphery), which leads to TC weakening. Maximum TC weakening took place ~20 h before landfall, just after the TC intensity reached its maximum. The minimum pressure increased by ~16 mb, and maximum velocity decreased by about 15 m/s. However, the difference in the intensities remains significant even during the TC landfall. Thus, the results indicate that there is another (in addition to decrease in the surface fluxes) mechanism of weakening of TCs approaching the land. This mechanism is related to effects of continental aerosols involved into the TC circulation.

A scheme of the aerosol-induced TC weakening mechanism is illustrated in **Figure 7**. The aerosol –induced intensification of convection leads to a) an increase in W at TC periphery. As a result mass updraft increases at the periphery and–a smaller amount of air mass and water vapor penetrates to the central part of the TC; b) extra convective heating at the periphery lowers the surface pressure at the TC periphery decreasing horizontal pressure gradient; and c) competition between two zones of convection arises (TC eye and at TC periphery). Compensation downdrafts caused by convection at TC periphery also tends to damp the convection in the TC eye. As a result of the interaction between the radial circulation caused by the two convective zones compensating downdrafts increase between these zones (at radii between 30-50 km to ~ 100-150 km) as it was demonstrated by Khain et al (2008a). The scheme shown in Figure 7 agrees well with results of Saharan dust effects of TC intensity reported by Zhang et al (2009). They found that convection in TC rainbands was negatively correlated with that in the eyewall in all simulations. The results agree well with recent observations by Jenkins et al (2008) and Jenkins and Pratt (2008).

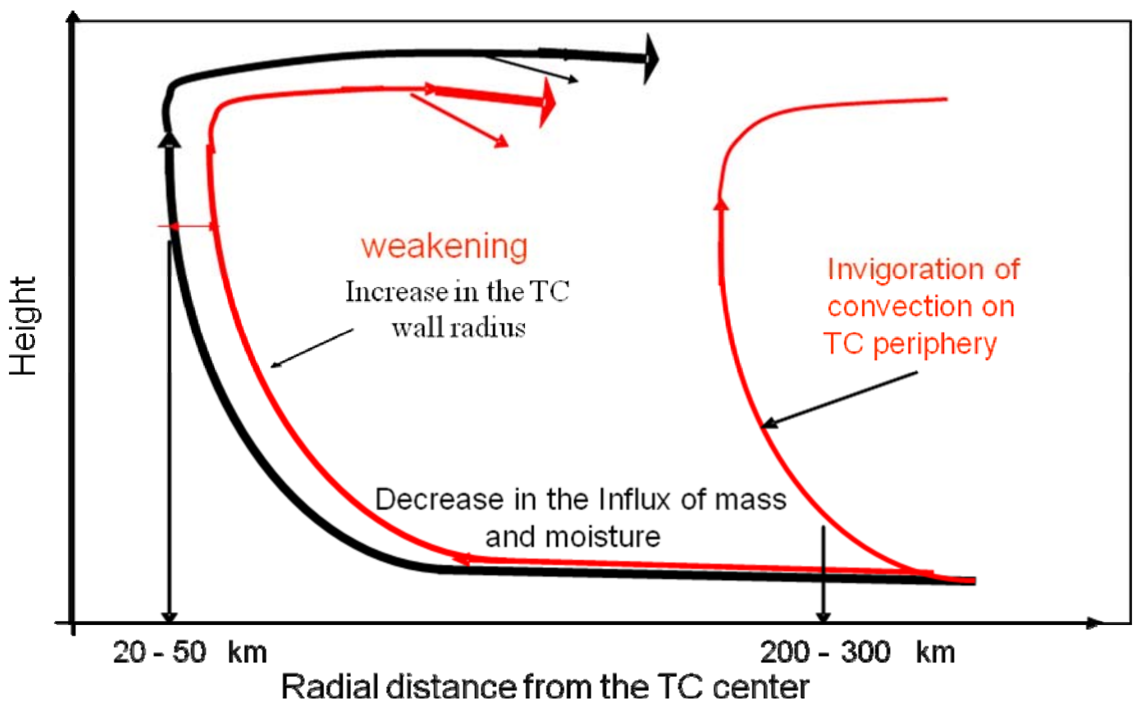

Figure 7. A scheme of aerosol effects on the TC structure leading to TC weakening (see text for detail).

The aerosol effects on TC intensity obtained in the present study turned out to be stronger than those reported by Rosenfeld et al (2007) and Khain et al (2008a). We attribute this difference to the utilization of SBM that accurately describes cloud dynamics and microphysics (Khain and Lynn, 2009; Khain et al 2009). For instance, the convection invigoration and increase in cloud top at the TC periphery (accompanied by intensification of lightning) were observed in hurricane Katrina and Rita when they were located in the Gulf of Mexico (see Khain et al 2008a for more detail). At the same time simulations with low aerosol concentration everywhere as well as simulations with bulk-convective parameterizations (Fierro et al, 2007) do not reveal these features. The changes of the TC structure during landfall resembles well that observed in Hurricane Katrina during its approach to the land. (Katrina reached its maximum intensity around Aug. 28 th, 18 UTC and weakened as it then approached land). The inner core of Katrina had collapsed during this period, and the storm became a very broad hurricane with near 100 knot winds extending almost 100 km from the storm center. This broadening and weakening during the last 6-8 hours before landfall has been difficult to capture in other simulations at NCAR and it can be speculated that aerosol effects played a role that has not been accounted for up till now. For instance the eye radius and the radius of maximum winds during TC landfall were significantly larger in case aerosol effects were taken into account. It should be stressed that the weakening and the inner core collapsing was simulated in spite of the fact that the SST maximum was located near the coastal line, and no TC weakening caused by the TC-ocean interaction was taken into account. The structure of TC cloudiness in simulation accounting for the effects of continental aerosol (MAR-CON) seems to be much more realistic than in the case of no effects of continental aerosols (MAR).

Note that these results concerning deep maritime cloud response to aerosols agree qualitatively and even quantitatively with those obtained using high resolution simulations

(350m x 125 m) using the Hebrew University Cloud model with SBM (Khain et al 2008a). However, an increase in the WRF-SBM resolution is very desirable to resolve better single clouds. It is especially important for description of clouds at the TC periphery.

4.2 Perspective of utilization of the TC models with spectral bin microphysics

The application of the explicit spectral bin microphysics in the TC models allows one to reproduce fine microphysical structure of clouds in TCs and to take into account mechanisms and processes that affect cloud microphysics and dynamics. The SBM can allow explicit treatment of sea spray and its effect on surface fluxes as well as on cloud microphysics and precipitation, because SBM describes explicitly evolution (evaporation, differential settling, collisions, etc) of drop size distribution of sea spray during droplet ascent in the boundary layer till cloud base. Hence, the SBM is an ideal tool allowing improvement description of TC-ocean interaction. The TC models with SBM coupled with the ocean can be the basis for TC models of the next generation. Since TC intensity, structure and precipitation depend on latent heat release determined by microphysical processes, the utilization of TC models with explicit SBM seems to open the way to improve prediction of TC intensity, wind and precipitation and lightning of landfalling hurricanes. Many studies have focused on reproducing the rapid intensification, but not on the collapse of the inner core and subsequent weakening as the storm approached the coast. The simulations with the SBM allowed reproduction of this effect. The utilization of TCs with explicit microphysics will allow simulation of TC genesis taking into account effects of Saharan dust. As it was shown by Zhang et al (2007, 2009) and Jenkins et al (2008), this effect may be quite significant. The utilization of explicit microphysics also allows investigation of the possibility to mitigate TC intensity by seeding of clouds at the TC periphery with small aerosol particles as it was discussed by Rosenfeld et al (2007), Cotton et al (2007) and Khain et al (2008a).

In the conclusion we would like to stress that the simulations of the current version of the TC model with spectral bin microphysics for 72 hours were performed using a standard PC cluster with 8 processors and a limited memory volume. The simulations require 10 days of computer time. It means that WRF with spectral bin microphysics can be used for scientific purposes using computers available in most scientific centers.

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