The Effects of Aerosols on the Evolution of a Tropical Depression off the Coast of Africa: As Seen From Simulations with the WRF Model with Spectral (Bin) Microphysics.

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Abstract

The evolution of tropical storm (TS) Debby that formed near the African coast on 21 August 2006 was simulated using the Weather Research and Forecasting Model (WRF; version 3.1) with explicit (non-parameterized) spectral bin microphysics (SBM). The spectral bin microphysical scheme implemented in the WRF-SBM calculates at each time step and at each grid point size distributions of atmospheric aerosols, water drops, and cloud ice, snow and graupel. The storm's evolution was simulated from 12 GMT 20 August until 12 GMT 23 August. The WRF-SBM was used to investigate the potential impact of Saharan dust on structure and intensity of the tropical storm via its effect on cloud microphysics. Results of two simulations with SBM are compared. In the first simulation, the aerosol concentration was assumed to be typical of maritime conditions with CCN of 100 cm^{-3} (at 1% of super-saturation). In the second simulation, the initial CCN distribution over the sea was assumed to be maritime, while over the land the CCN concentration was assumed (as reported) to be of high concentration within the lowest atmospheric 2 km layer. The continental aerosols (dust) penetrate the Atlantic Ocean by an easterly wind. As a result, aerosols penetrated into the storm circulation and affected its evolution. It is shown that aerosols intensify convection over a large area, and cool the air in the boundary layer, hindering the development of the Eye Wall and

weakening the storm. This concept is further demonstrated through comparison of the simulated lightning structure within the two cases.

1. Introduction

1.1. Factors affecting Tropical-Storm genesis

Despite many worthy observational and numerical modeling studies in recent decades, our understanding of the detailed physical processes associated with the early stages of tropical cyclone (TC) formation is still inadequate, and, operational forecast skill remains to be quite low (Gray 1998; Vizy and Cook, 2009). General parameters affecting the TC genesis such as wind shear, air humidity and SST, effects of Coriolis force, etc., were revealed in early studies and led to deriving different expressions for seasonal potential genesis (e.g., Gray, 1979; Anthes, R.A., 1982; Ivanov and Khain 1983; Khain 1983; Emanuel, 2005).

At the same time the seasonal genesis potentials can only present some estimation of the probability of development of particular tropical depressions (TD) and cannot be used for the purposes of predicting the evolution of particular TDs. Note that "genesis efficiency" in nature is quite low: among approximately 1000 TDs arising during the tropical season, only about 100 reach the intensity of tropical storms and only about ten of them reach hurricane intensity (Gray, 1979, 1998). Detailed analysis performed in studies by McBride (1981a,b), McBride and Zehr (1981), Erickson and Gray (1977), Zehr (1976, 1992) showed that TDs tend to develop when their vorticity is spatially concentrated and exceeds some critical value. At the same time the particular mechanisms of formation of high values of vorticity remains unclear.

Owing to the dominating role of latent heat release, Charney and Eliassen (1964) proposed the mechanism of interaction of convective and TC-scales leading to TC genesis. According to this mechanism convection interacts with the TC-scale vortex via the friction in the boundary layer (so called CISK-mechanism). The convective parameterizations based on the CISK allowed developing the first models predicting TC formation. The major physical meaning of the CISK is the creation of friction-induced radial velocities in the boundary layer concentrating convection (and convective heating) in the central region of a TD.

In a set of studies by Montgomery and Farrell (1993), Hendricks et al (2004), Montgomery et al (2006), the formation of a cloud "coherent" structure in TD and the TD development is attributed to the merger of small scale vortices in conjunction with so-called vortical hot towers (VHT). Tory et al (2007) showed that the exact detail of the vortex interactions was unimportant for qualitative genesis forecast success. Instead the critical ingredients were found to be sufficient net deep convection in a sufficiently cyclonic environment in which vertical shear was less than some destructive limit. Thus, again the main factors determining the TC genesis were found to be vorticity, vertical wind shear, and proper spatial distribution of convective forcing.

The possible sensitivity of TC intensity to the spatial distribution of latent heat release can be derived from lightning data. Lighting indicates zones with the strongest convection. 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.

An increase in computer power has allowed the investigation of TC genesis using models with a high resolution of several kilometers (e.g. Wang 2002; Zhu and Zhang 2006; Frisius and Hasselbeck, 2009). Wang (2002) concluded that spiral rain bands forming at the TD periphery have a weakening effect on the tropical cyclone because they introduce low-entropy air by downdrafts into the boundary layer and hinder the boundary-layer inflow towards the eyewall. Sensitivity studies produced by Wang (2002) and Zhu and Zhang (2006) showed that the most intense tropical cyclone resulted when latent cooling processes (evaporation and melting) were switched off. More detailed investigation of spatial distribution of convective heating (latent heat release) and cooling by evaporation and melting was performed by Frisius and Hasselbeck (2009) using axi-symmetric and fully 3-D mesoscale models with bulk-parameterization of microphysical processes. In a set of sensitivity studies they showed that the fact whether a TD will develop or not dramatically depends first on the spatial distribution of heating/cooling within an area of a TD. For instance, switching off drop evaporation led to a decrease in the TD eye diameter and the convective inhibition outside the eyewall. As a result, their TD rapidly developed into a TS when there was no evaporation of water droplets (as well as in the simulations in which the air cooling caused by melting was switched off).

The spatial distribution of latent heat within a TC area may depend on the concentration of atmospheric aerosols. The major hurricanes that affect the United States can often be traced back to weak disturbances moving westward across the tropical Atlantic Ocean. Many hurricanes that have reached the US coast have also been found to contain Saharan dust. The transport of Saharan dust may be so strong as to affect precipitation of Florida storms (e.g. Van Heever et al 2006).

In some studies it was shown that the penetration of Saharan dust into TDs hinders their development into TS (e.g. Evan et al 2006). This effect is typically attributed to two factors: a) the penetration of warm air from Africa to the marine layer just or above the inversion increases the stability of the atmosphere and hinders the development of convection. In this case dust plays the "passive" role of tracer of warm air; b) to the radiative effects of dust. For instance, Carlson and Benjamin (1980) showed that heating rates associated with Saharan dust can be in excess of 1 C per day. Khain and Agrenich (1987) was one of the first where the effects of solar heating of dust on the TC intensity was investigated using an axisymmetric TC model. Recent studies by Lau and Kim (2007 a,b), Evan et al (2008), and Wang et al (2008) suppose that dust intrusions are responsible for chilling the SST in the North Atlantic. It was assumed that the outbreak of Saharan dust led via this mechanism to a decrease in TS activity. In particular, it was assumed that the excess of Sahara dust in the 2006 premonsoon season, as compared to 2005, contributed to the big decrease in TC activity from 2005 to 2006.

Substantiality lower attention was paid on the potential effect of aerosols (in our case Saharan dust) on the intensity and evolution of TD via their effect on microphysics of deep convective clouds. During the past decade it was found that aerosols substantially affect cloud microphysics and consequently the rate of latent heat release, dynamics and precipitation (see, overviews by Khain et al 2009; Levin and Cotton, 2009 and Rosenfeld et al, 2008). In particular, it was found that aerosols with sizes below ~0.5-1 μm invigorate tropical convection increasing vertical velocities and, sometimes, 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 microphysics and the dynamics of convective clouds. Some observations indicating the possible aerosol effects on TS development via their influence on the cloud microphysics of TS clouds have been performed by Jenkins et al (2008) and Jenkins and Pratt (2008).

Rosenfeld et al (2007) and Khain et al (2008, 2010) showed that continental aerosols penetrating the circulation of TC approaching the land invigorate convection at the TC periphery and weaken its intensity in the Eye Wall leading to the weakening of landfalling TC.

The weakening takes place over the sea a few hundred km from the coast line even in cases when the SST was assumed to be unaffected by TC-ocean interaction. Zhang et al (2007, 2009) invested the effects of aerosols on the evolution of idealized TC. They used RAMS with a 2-moment bulk parameterization of microphysical processes. These simulations suggested that under some conditions aerosols could lead to TC weakening.

Khain et al (2010) used the WRF model with spectral bin microphysics (SBM). This microphysics allows the explicit calculation of the effects of aerosols on cloud microphysics through their effect on size distributions of droplets and other hydrometeors. Khain et al (2010) concluded that the intensification of convection at TC periphery caused by continental aerosols (in addition to decrease in the surface latent heat flux) leads to a decrease in intensity of landfalling TC. In the present study, the evolution of TS Debby (2006) is simulated using the WRF model with SBM.

2. Model and experimental design

The SBM scheme implemented into the WRF (Skamarock et al., 2005, Version 3) has been described by Khain et al (2004, 2010) and Lynn et al (2007). The scheme is based on solving the kinetic equation system for the size distributions of four classes of hydrometeors: water drops, ice crystals/aggregates and graupel/ hail. Each hydrometeor class is described by a size distribution function defined on the grid of mass (size) containing 33 mass bins. The particles with sizes below 150 μm in the crystals/aggregate class 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). Characteristics of particles belonging to different classes such as bulk density, fall velocity, aspect ratios, etc. are expressed as functions of their mass following Khain and Sednev (1995), Khain et al (2004) and Pruppacher and Klett (1997).

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 μm 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, and aerosol particles are 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 due to advection, sedimentation and different microphysical processes.

Super-saturation 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 super-saturation, the critical size of aerosol particles to be activated to drops is calculated. When aerosol particles exceeding the critical size are activated, the corresponding mass bins in the aerosol size distribution become empty. It means that within the cloud updraft there is the flux of droplets and flux of non-activated CCN. When super-saturation in updrafts exceeds its value at lower levels, a new portion of CCN will be activated to droplets (in-cloud nucleation). This process is especially important in maritime clouds where super-saturation is high and often increases with height because of a decrease in droplet concentration and increase in vertical velocity with height. The inclusion of secondary nucleation processes leads to the formation of a 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 the possible droplet nucleation on aerosols penetrating the lateral cloud boundaries at upper levels.

Note that in the SBM, water drops are not separated artificially into cloud water and rain water (in contrast to 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 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 an accurate superposition method (Pinsky et al, 2001, Khain et al 2001) and used in the form of lookup tables. These 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). 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).

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).

The SBM model does not requires any tuning of the scheme parameters and was successfully used without any significant changes (at the exception of different initial soundings and aerosols) 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).

2.2 Experimental design

2.2.1 Case study

A pair of simulations were used to study possible aerosol effects on the evolution of TS Debby (AL042006) 21-26 August 2006 (Zawislak and Zipser 2010). Debby formed from a vigorous tropical wave that moved across the west coast of Africa on 20 August. Almost immediately after moving offshore, the wave developed convective banding and a broad closed circulation. The “best track” chart of the tropical cyclone’s path is given in **Fig. 1**. The depression initially moved west-northwestward to the south of the subtropical ridge. Around 1200 GMT 22 August, the center of the cyclone passed about 100 n mi to the southwest of the southernmost Cape Verde Islands, bringing thunderstorms and gusty winds to the southern islands of Fogo and

Brava. The depression strengthened as it moved away from the islands, becoming a tropical storm around 0000 GMT 23 August. By 1200 GMT that day Debby's sustained winds reached 45 Knots, and there was little or no change in strength for the next two days as the cyclone moved between west-northwest and northwest at 15-20 Knots over the open waters of the eastern Atlantic. Intensification during this period appeared to be limited by a dry and stable air mass surrounding the cyclone, along with marginal sea-surface temperatures. On 25 August, southerly shear began to increase in association with an upper-level trough, displacing the deep convection to the north of the center. Debby started to weaken, and became a depression around 0600 UTC 26 August. (Franklin 2006).

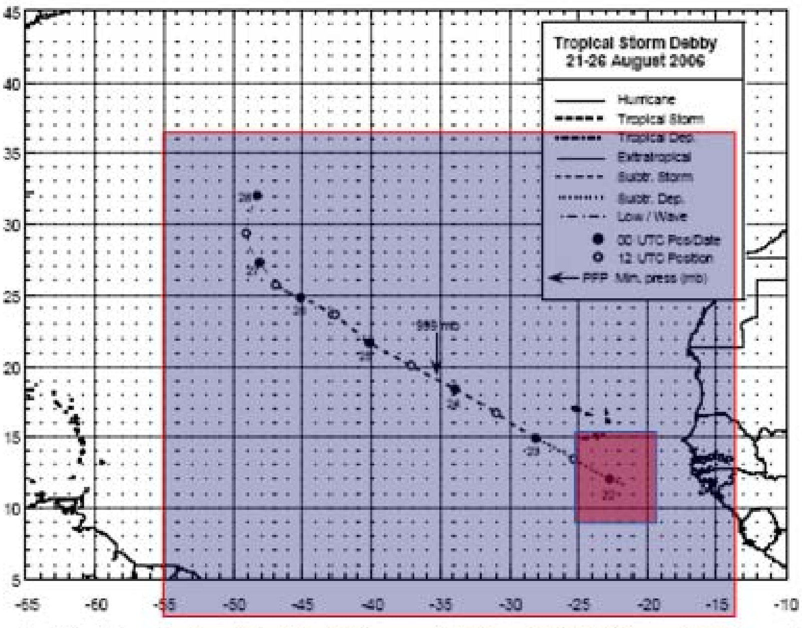

Figure 1. Best track for Tropical Storm Debby, 21-26 Aug, 2006 and the structure of model meshes at t=0.

In situ measurements in TD Debby (Zipser et al 2009) showed the existence of strong wind shear in the 5-8 km layer. This shear inhibited convection, so that maximum updraft velocity was quite low (~5m/s). Microphysical measurements showed the dominating contribution of graupel to ice habits. A warm core of 4-5 C was found. The wind maximum was observed on the east side of the storm and exceeded 30 ms^{-1} . Observations also show that Debby was completely surrounded by a large Saharan Air Layer (SAL), with a significant aerosol particle layer with depth up to ~ 5 km. According to Zipser et al (2009) the SAL was often adjacent with deep convective clouds, sometimes penetrating them and sometimes not. At the same time TD Debby can be considered as TD which clouds in its center region were only weakly affected by continental aerosols (Heymsfield 2010, personal communication).

2.2.2 The simulation design

A two-nested gridded WRF (version 3.1) was used, and the nest moved using a cyclone-following algorithm. The spacing 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 720 x 720 km^2. The inner mesh is movable and follows the height minimum at z=850 mb.

The outer grid was motionless and was 3240 x 3240 km^2 with spacing of 9 km. The bulk-parameterization of Thompson et al. (2004) was used on the outer grid. The grid structure at t=0 is shown in **Figure 1.** 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 were 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 this data as well. The surface water temperature was initialized at 12 GMT on 20 August, and was not updated during the experiments.

Cloud droplets arise on aerosol particles (AP) playing the role of Cloud Condensational nuclei (CCN). According to observations (Zipser et al 2009; Thohy et al 2009) dust in the Saharan air layer (SAL) is slightly hygroscopic and acts as CCN. The distribution of dust had a maximum within the diameter range 0.1-0.2 μm (Zipser et al 2009). Thus, the high concentration of submicron APs may influence microphysical properties of cloud systems interacting with the SAL. The initial (at t=0) CCN size distribution near the surface is calculated using the empirical dependence of concentration of activated CCN N_{ccn} at supersaturation with respect to water S_w (in %):

$$N_{ccn} = N_o S^k \qquad (1)$$

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 a characteristic scale of 2 km. Aerosols were transported over the entire computational area similarly to other scalars like the mixing ratio.

To investigate the possible aerosol effects on microphysics and the dynamics of the TS, two simulations were carried out: a) in the first simulation N_o was set equal to 100 cm^{-3} typical of maritime atmosphere over the whole computational area (hereafter, referred to as "MAR"); b) in the second, the initial CCN concentration over the land N_o was set equal to

$3000\,cm^{-3}$, typical of continents under not very polluted conditions. Initially, over the sea N_o was set equal to 100 cm^{-3} in all simulations. For the maritime aerosol concentrations, the slope parameter k was set equal to 0.46. It was 0.31 for the continental AP concentrations. The magnitudes of these parameters are typical of maritime and continental conditions, respectively (Pruppacher and Klett 1997). The vertical profiles of the aerosols decreased exponentially with height beyond two kilometers.

The evolution of the background fields was initially simulated for six hours to allow time for the dust to move from over the continent to over the ocean in case MAR-CON. Then, weak temperature sinusoidal heating along the vertical line passing through the point of minimum geopotential at ~800 gPa was used to trigger the TS. The heating zone was moved with the TD. The heating maximum of about 8 C day^{-1} was applied during 18 hours. This magnitude of the heating is quite small: it is only a few times larger than the radiative forcing. For instance, it is much lower than that in natural clouds, where the heating reaches 100 (and more) C day^{-1} . At the same time the heating allowed triggering of TD without using bogusing techniques of implementation of initial vortex. Such triggering was necessary because the crude resolution of reanalysis data does not resolve the TD scales. The weak heating for TD triggering does not require any adaptation procedure. The initial heating has the same meaning as that used for triggering of convective clouds. The heating allows the system "to recognize" its center, but the further evolution is determined by the cloud formation within the TD, interaction of the TD with the surrounding, etc., and does not depend on the heating.

3. Results of simulations

Figure 2 shows that the two simulations reproduce the observed *changes* in surface pressure and wind speed changes in good agreement with observations. Toward 72 hr of simulation time, the errors in the prediction of minimum pressure and maximum velocity were very low in both simulations when compared to the observed values. The simulation MAR-CON shows, however, better agreement with the observations, especially after 50 hours of simulation time. Trajectories of simulated storms coincided reasonably well with the observed trajectory in both simulations (not shown).

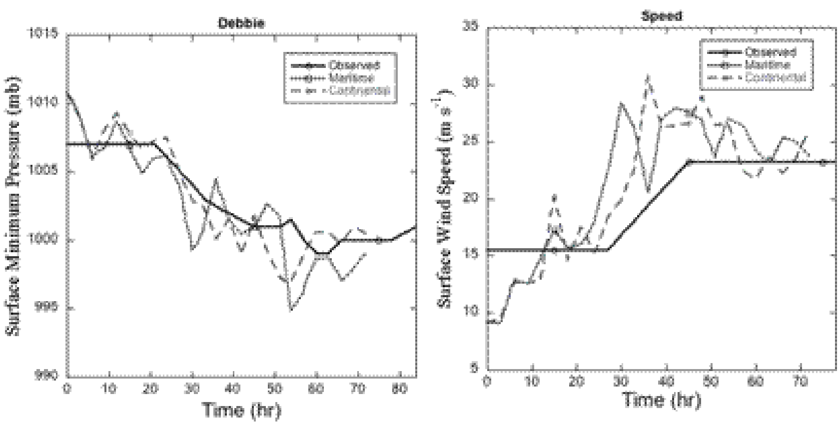

Figure 2. The time dependences of minimum pressure (left) and maximum wind speed (right) in two simulations: MAR and MAR-CON.

Figure 3 shows the fields of maximum AP concentrations in the MAR and MAR-CON simulation at 63 and 72 hours of simulation. This is the time near the greatest penetration of aerosols into the TS in MAR-CON. The MAR simulation shows a fairly consistent field, except along the previously traversed path of the TS rain bands. One can see a decrease in the AP concentration caused by CCN nucleation into drops and precipitation. In contrast, MAR-CON shows a spiral of high aerosol concentrations beginning in the northwest quadrant of the domain and reaching into the center of the TS. Note that Fig. 3 shows fields of only non-activated APs, whose concentration is high because it includes a great number of smallest APs which do not activate to droplets at the supersaturations calculated in the model. In MAR the CCN concentration is low which leads to a very low droplet concentration with maximum below ~50 cm^{-3} (**Figure 3, panels a,b**).

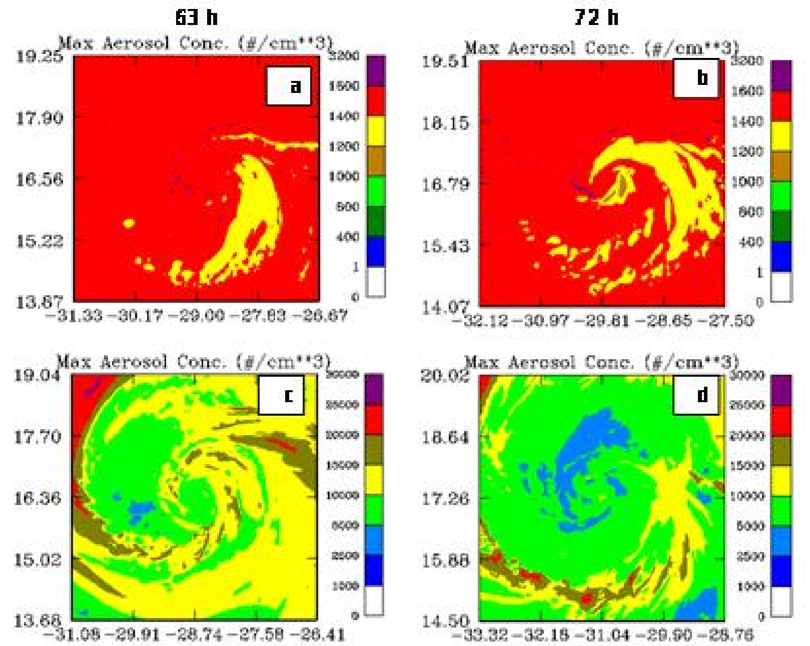

Figure 3. Fields of maximum AP concentrations in the MAR (panels a,b) and MAR-CON (panels c and d) simulation at 63 and 72 hours of simulation.

Low droplet concentration leads to high supersaturation, fast diffusional drop growth, and an early beginning of intense collisions. Rain drops form at low levels and fall down without ascending first to the freezing level. The rapid formation of rain decreases cloud water content (CWC) in MAR run, so that CWC does not exceed ~0.3 gm^{-3} (**Figure 4, panels a,b**). Associated with the high aerosol concentration in MAR-CON is a relatively high cloud drop concentration (up to 250 cm^{-3}) and high cloud mass content (up to 0.6 gm^{-3}) relative to MAR (panels c and d in **Figures 4 and 5** respectively).

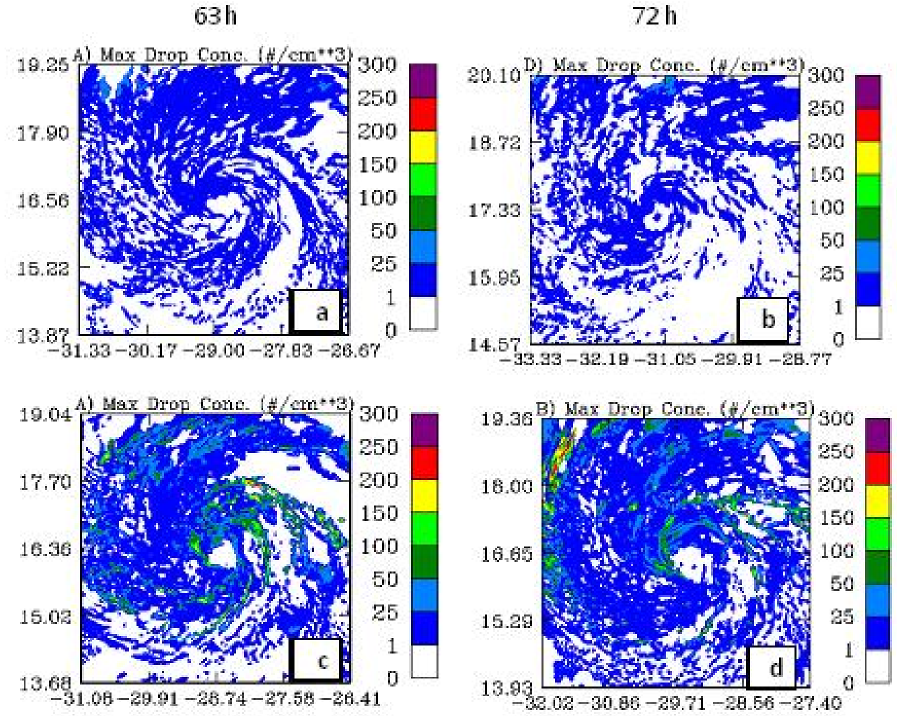

Figure 4. Fields of column maximum cloud droplet concentration in the MAR (panels a,b) and MAR-CON (panels c and d) simulation at 63 and 72 hours of simulation.

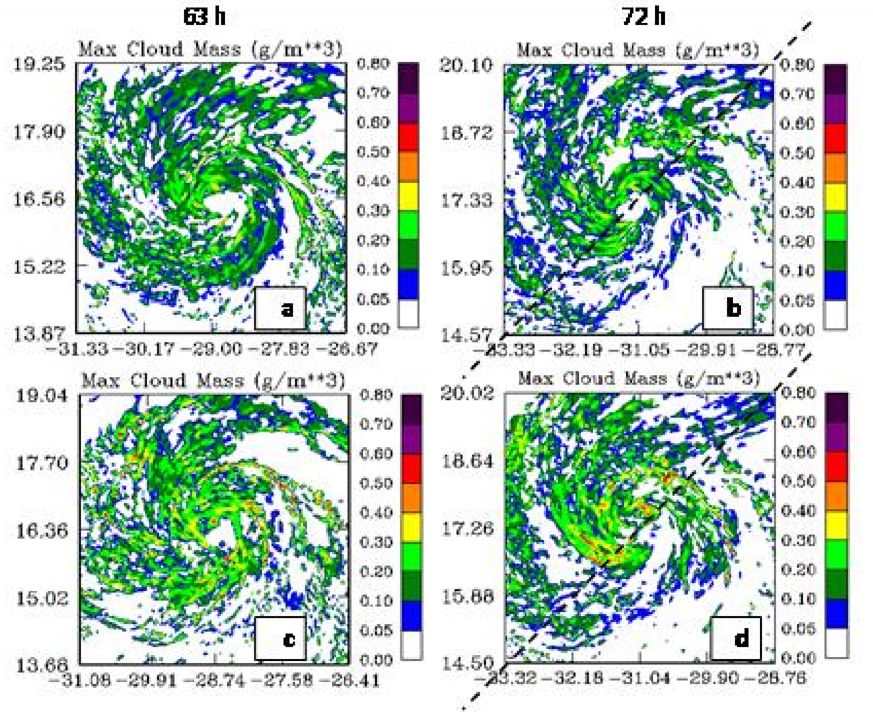

Figure 5. The same as in Figure 4, but for cloud water content. Dashed straight lines in panels b and d indicate the location of vertical cross-sections presented in Fig. 6

Figure 6 shows the fields of total liquid water content (QC+QR) in the north-east – south-west vertical diagonal cross –section at t=72 h in MAR and MAR-CON (the line showing the cross-section is shown in Figure 5 (panels b and d) by dashed lines. One can see the dramatic effect of APs on microphysics and structure of cloudiness. Droplets in MAR-CON are smaller than in MAR and ascend to higher levels producing supercooled rain drops at ~5 km. While in MAR, the spatial distribution of convection is more or less symmetric with respect to the TD center, it is highly asymmetric in MAR-CON. In MAR-CON droplets penetrate to higher levels and produce stronger supercooled rain that starts at height of 6-7 km.

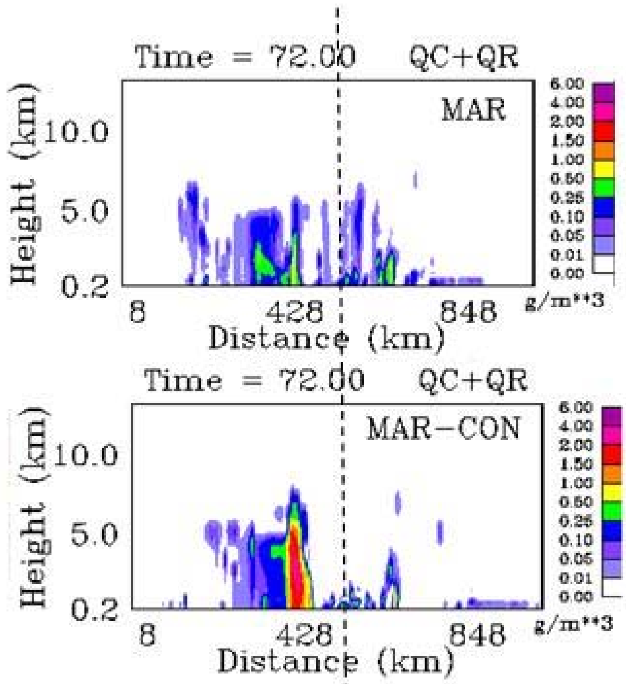

Figure 6. Vertical north-east – south-west cross section in the field of liquid water content

Figure 7 shows fields depicting the maximum updrafts. Prior to the penetration of aerosols into the storm, convection was concentrated in the storm central region in both simulations (not shown). Later the differences in the W fields become significant: W are lower in MAR, but elevated values are more strongly concentrated with respect to the TS center. In MAR_CON, the vertical velocities are higher (up to 7 m/s), but zones with high updrafts are located more that 100 km from the TS center (panel d). It is obvious that convection at the TC periphery is stronger in MAR-CON. This comparison reveals the aerosol –induced convection invigoration reported earlier in many observational and numerical studies (see Khain 2009 for detail).

Differences in warm microphysics and updrafts of clouds in MAR and MAR-CON affect the dramatic difference in ice microphysics. **Figures 8 and 9** show field of maximum values of graupel and snow, respectively. One can see that graupel is the dominating solid hydrometeor in MAR-CON, while snow dominates in MAR. The difference on the snow and graupel contents can be attributed to the fact that in MAR-CON supercooled CWC is higher which leads to more intense riming of snow, which converts it to graupel. In the MAR case there is very little supercooled water and, respectively, snow does not transfer to graupel as rapidly as in MAR-CON.

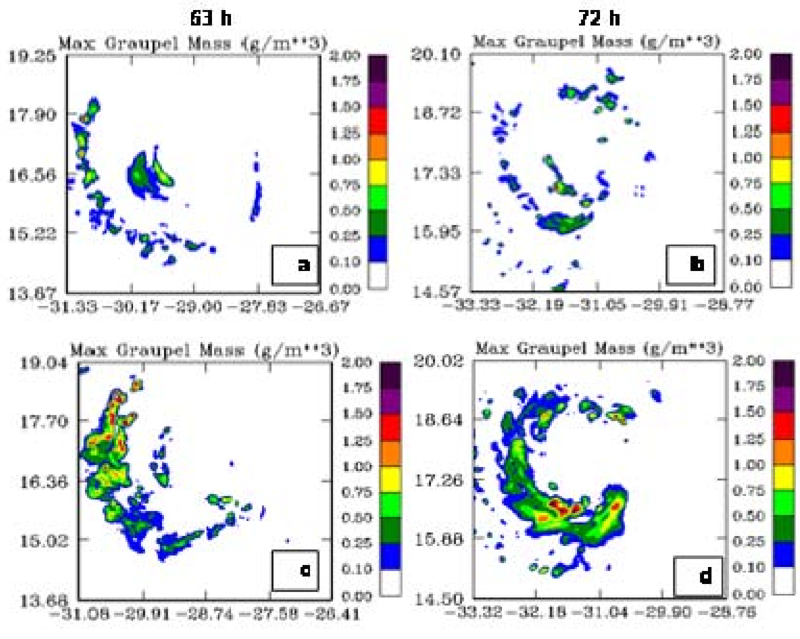

Figure 8. The same as in Figure 4, but for maximum graupel content.

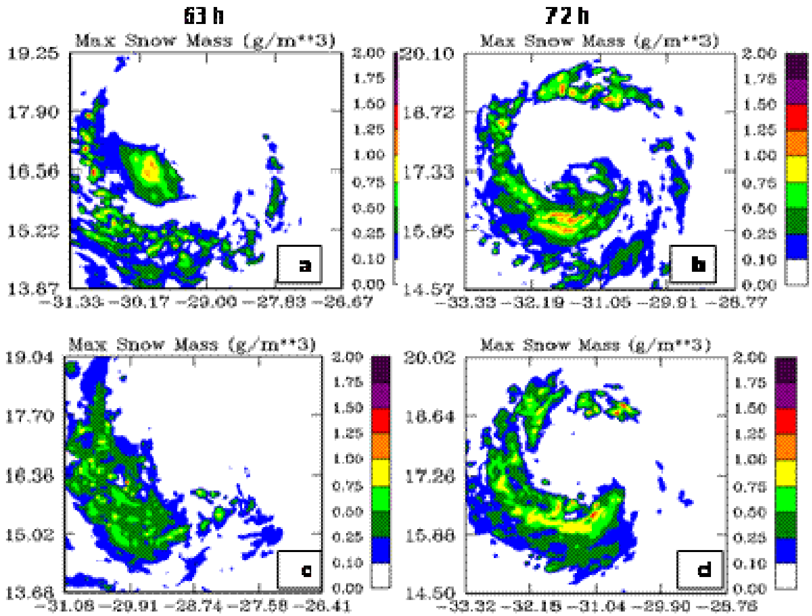

Figure 9. The same as in Figure4, but for maximum graupel content.

Figure 10 shows that clouds in both runs reach 14 km with maximum graupel mass content in MAR-CON of 2-3 gm^{-3}, and snow mass content of 1 gm^{-3}. Again one can see that convection is stronger in the TS center in MAR, while it is stronger in MAR-CON at the TS periphery.

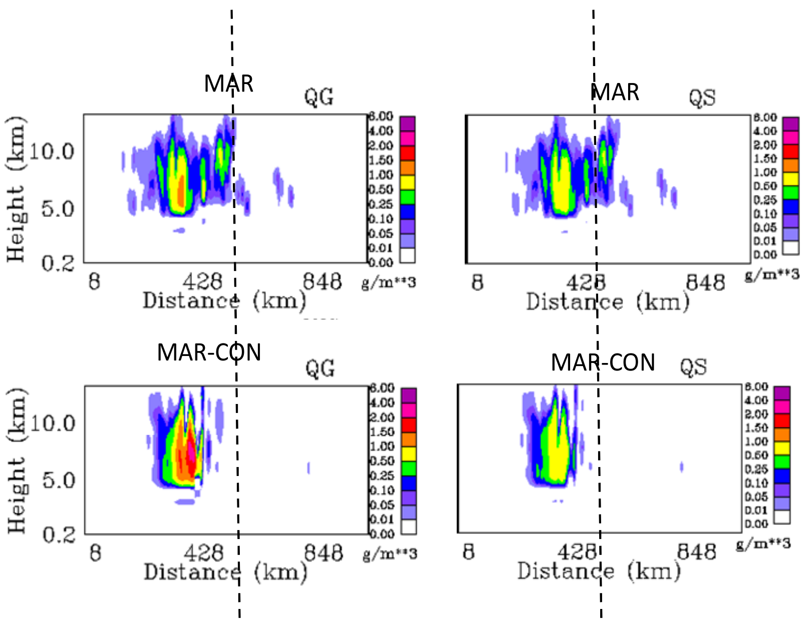

Figure 10. Vertical north-east – south-west cross section in the fields of graupel (QG) and snow (QS) mass contents at t=72 h.

Figure 11 shows time dependences of area averaged maximum CWC, rainwater content, snow and graupel. One can see that CWC and graupel mass contents are larger in MAR-CON for all periods of the simulation. At the same time snow dominates in MAR. Note that according to the in-situ observations (Zipser et al 2009), graupel was dominating in the TC Debby, which corresponds well with the measurements of significant concentrations of AP in the surrounding cloud bands of this TC. As regards to rain mass content, there is no any significant difference in area averaged values.

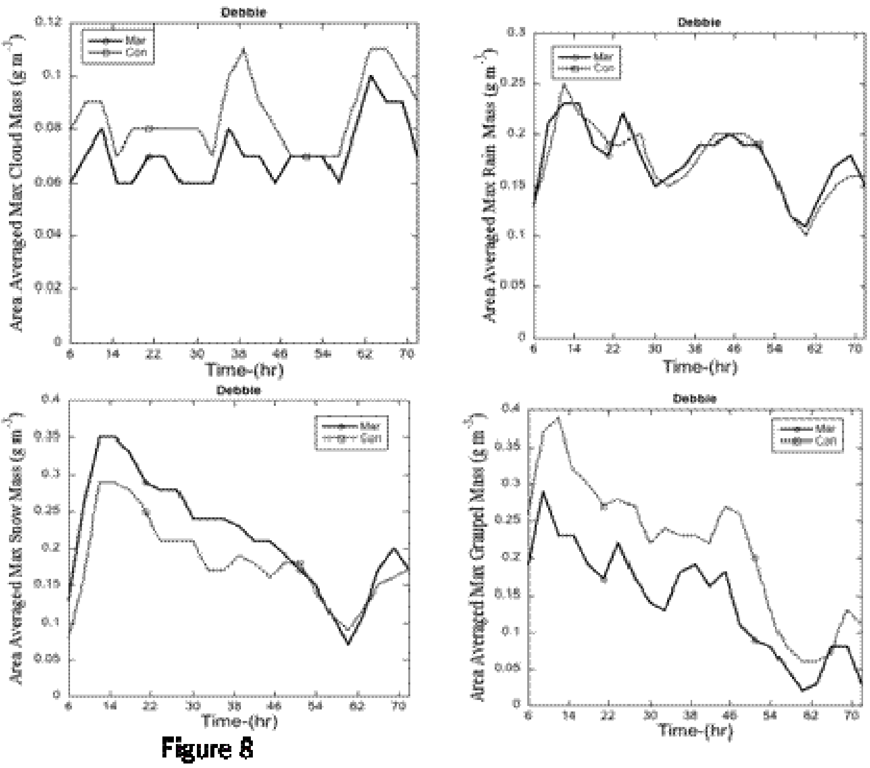

Figure 11. Time dependences of area averaged maximum CWC, rainwater content, snow and graupel.

Figure 12 shows fields of accumulated rain at 72 h in MAR (left) and MAR-CON (right). The fields look quite similar as regards to maximum rain amounts. Note, however, a significant difference: in MAR precipitation is centered closer to the TS center, while in MAR-

CON precipitation covers a larger area at quite a high distance from the TS center. Since accumulated rain is the measure of the heating of the atmosphere, one can see that aerosols lead to heating at larger distances from the TS center.

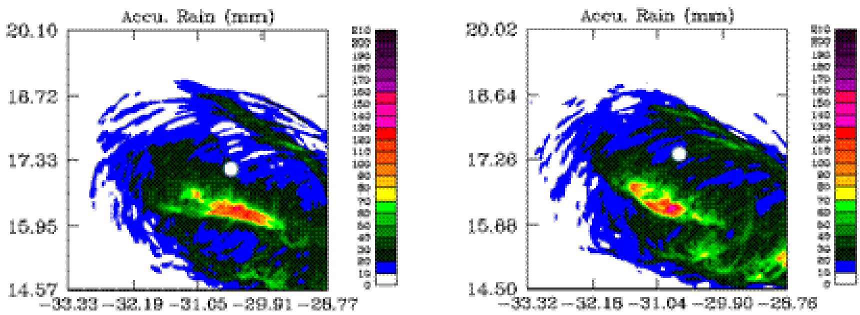

Figure 12. The fields of accumulated rain at 72 h in MAR (left) and MAR-CON (right). White circles denote the center of TS in these simulations.

The fields of the temperature at 72 h and the temperature difference of the temperature fields between MAR-CON and MAR within the north-east –south-west vertical cross-section (marked by dashed line in Figure 12, panel d)are shown in **Figure 13**. In agreement with observations the results indicate 5–6 ^{0} C temperature rise associated with a warm core at 700 hPa (center temperature of 14 ^{0} C) (Zipser at al 2009; Zawislak and Zipser, 2010). Note that the extra heating in MAR-CON takes place above the boundary layer at the TC periphery. Below the cloud base the evaporation of precipitation water leads to air cooling which is stronger in the CON-MAR (**Figure 13**). It means aerosol-induced cloud invigoration at the TS periphery leads to the formation of cold pool in the BL, so that comparatively low temperature air penetrates the central part of TS, hindering the formation of deep convection there.

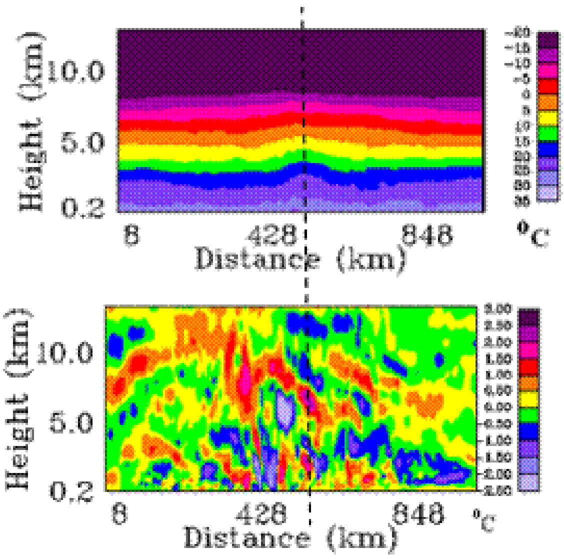

Figure 13. North-east – southwest vertical cross-sections of the temperature field in MAR (upper panel) and the difference between MAR-CON and MAR. Dashed line denotes the center of the computational area which is close to the center of the TS. One can see warm core of about 5 oC at ~3 km (700 hPa). In MAR-CON temperature at the TS periphery is higher above the boundary layer, while in the BL temperature is lower which represents formation of cold pool in MAR-CON in zone of active convection.

The microphysical and dynamical factors mentioned above lead to significant differences in the fields of wind speed (**Figure 14).** Note that both in MAR and MAR-CON maximum wind speed takes place to the east from the TS center which is in agreement with observations (Zipser at al 2009). At it was shown also in Figure 2, the maximum values of wind speed are close in both runs. At the same time Figure 14 indicates substantial difference in the TS wind structure: while in MAR strong winds cover a significant area around the TS center, zones of strong wind in MAR-CON are smaller, but zones of significant winds cover a much larger area indicating a much slower decrease of wind speed with the increase in the distance from the TS center as compared to that in MAR. One can see that zone of strong winds in MAR-CON is much less concentrated than in MAR. Taking into account that the TC Debby started weakening at 72 h, results obtained in MAR-CON seem to be closer to observations.

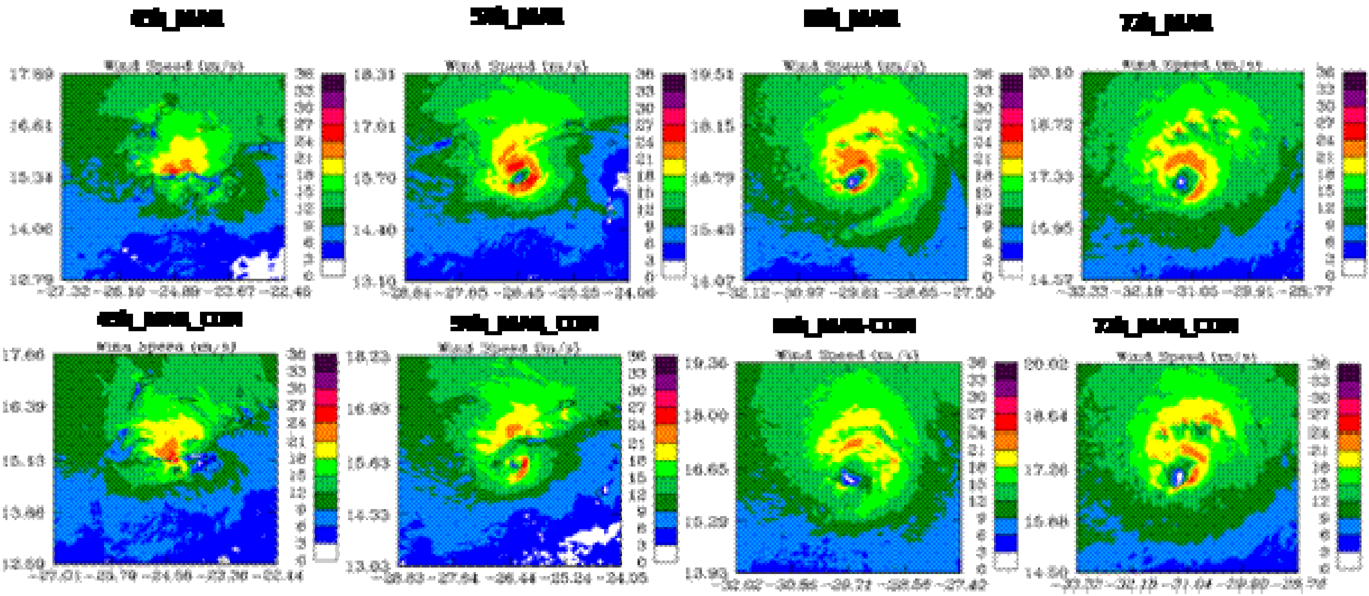

Figure 14. Fields of wind speed in the boundary layer at different time instances in simulation MAR (upper row) and MAR-CON (lower row).

The comparison of the fields of cloud top height (**Figure 15**) shows that aerosols affect significantly the structure of cloudiness, and increase the asymmetry of the TS. Note that in MAR-CON clouds reach higher levels, which is the consequence of the higher vertical velocities (aerosol induced invigoration of convection).

Comparison of radar reflectivity calculated in the same cross-section with the observed one (Figure 7 in Zipser et al 2009) indicates a good agreement (not shown): the structure of convection is substantially asymmetric with the maximum of convection within south-east quadrant; maximum radar reflectivity of 60 dBZ is reached within the layer from 3 to 5 km.

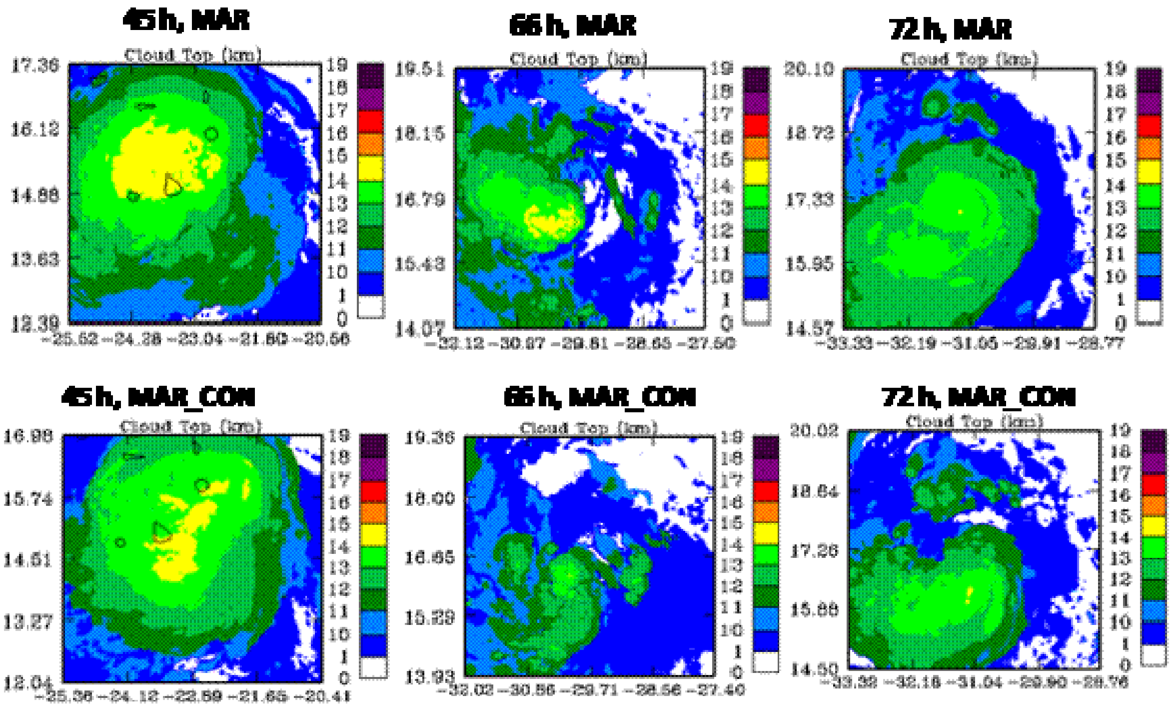

Figure 15. The same as in Figure 14, but for cloud top heights.

We complete our analysis by presenting **Figure 16** that compares the fields of Lightning Potential Index (LPI) in these two simulations. The LPI was introduced by Yair et al (2009) and Lynn and Yair (2009) and was used by Khain et al (2009) to estimate the lightning intensity at the periphery of landfalling TCs. The LPI is the volume integral of the total mass flux of ice and liquid water within the "charging zone" (0 to -20°C) of the cloud. It is calculated using the simulated grid-scale vertical velocity, and simulated hydrometeor mass mixing-ratios of liquid water, cloud ice, snow, and graupel as $LPI = 1/V \iiint \varepsilon w^2 \, dx \, dy \, dz$, where V is the model unit volume, w is the vertical wind component in m s^{-1}. In essence, ε is a scaling factor for the cloud updraft, and attains a maximal value in clouds with strong updrafts when when the mixing ratios of super-cooled liquid water and ice ice species (cloud ice, graupel and snow) are equal. It signifies the fact that charge separation requires all these ingredients to operate synergistically within the charging zone, as shown by many laboratory

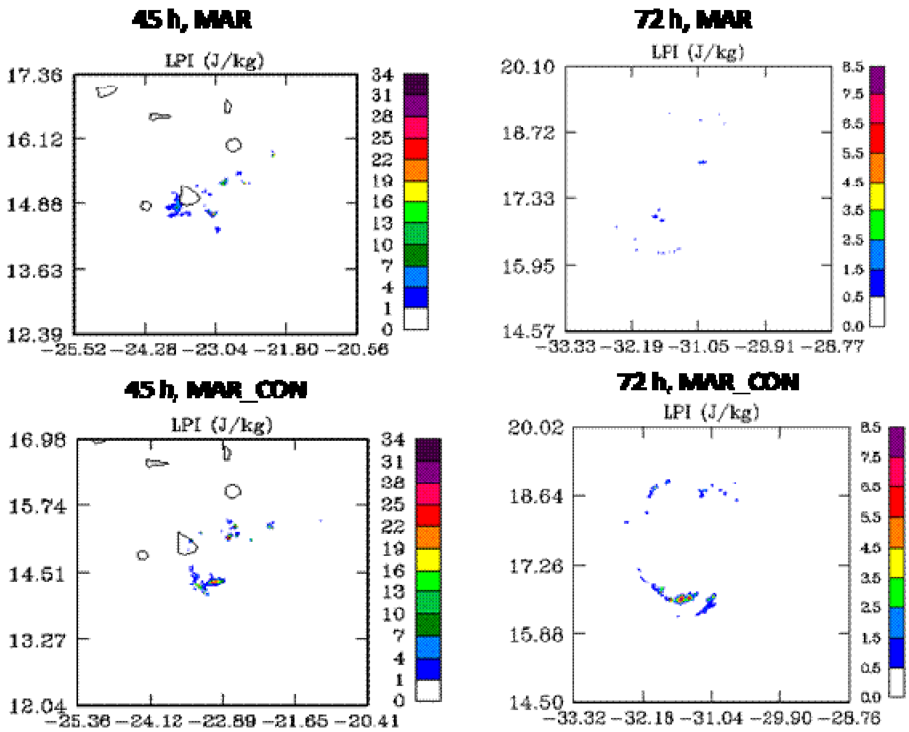

Figure 16. The same as in Fig. 9, but for Lightning Potential Index (LPI)

experiments summarized by Saunders (1993). The LPI has the same meaning as the lightning probability parameter introduced by Khain et al (2008a).

Figure 16 shows that in the MAR case there is no lightning or the lightning intensity is very low. At the same time lightning in the MAR-CON (especially at 72 h) is substantially stronger than in the MAR. This effect can be attributed to the stronger W, graupel and supercooled CWC in the MAR-CON, which coexistence is the necessary condition for lightning formation. It is interesting that the zone of enhanced lightning intensity in the MAR-CON form the circle of with radius of ~150 km. We see here the same effect of continental aerosols as was found in the case of a landfalling TC (Khain et al 2010).

4. Discussion and Conclusions

The evolution of TC Debby (2006) forming from a weak vortex, intensifying to the central pressure of ~ 1000 mb with the maximum wind up to 30 m/s and then decaying was simulated using WRF model with SBM. Simulations were performed with low maritime aerosol concentrations as well as with aerosol concentrations obtained as a result of Saharan dust outbreaks. Good agreement of simulated pressure and maximum velocity with observations was obtained in both simulations. Besides, the simulations using the SBM indicated errors in the maximum velocity of about 1-2 m/s.

Note that the prediction skill of TC genesis performed by current convective parameterizations is quite low. According to Vizy and Cook (2009) only one of 6 tested parameterization combinations realistically captured the TD Debby intensification. Three combinations did not predict any intensification at all, while two combinations predicted intensification to hurricane intensity with minimum pressure of ~970 mb.

Other features such as concentration of zone of strongest winds in the eastern part of storm, the dominating role of graupel, low vertical velocities (below 5m/s) are also agree with observations. This result indicates the significant potential of SBM to improve the prediction of TS genesis, as well as to understand better the factors affecting TC evolution.

The comparison of structures of TCs forming in the MAR and MAR-CON simulations shows that aerosols intensify the convection at the TD periphery and hinder the TD development by decreasing the rate of the concentration of convection in the TD center. Another possible reason of the convection weakening in the center of TS in MAR-CON is the penetration of cold air in the boundary layer which increases atmospheric stability of the central zone of the TS.

Thus, the microphysical mechanisms of aerosols effects on the genesis of TC via their effect on cloud microphysics do not contradict thermodynamic effects caused by dust leading also to a decrease in hurricane intensity and hurricane activity. It is interesting to note that the relative TD weakening in MAR-CON as compared to MAR is accompanied by an increase in precipitation at the TC periphery, the invigoration of convection in zones of high aerosol concentration, etc. This result indicates that the convective invigoration of individual clouds or even cloud clusters does not inevitably leads to the intensification of larger scale phenomena such as tropical cyclones. As regards TS in general, the effect of aerosols is strongly related to the spatial distribution of latent heat within the TS area in good agreement with results of many studies mentioned in the Introduction.

According to Vizy and Cook (2009) " There is currently no parameterization accounting for the influence of the variability of aerosol forcing in the model (WRF-the authors)". Thus, the WRF with spectral bin microphysics seems to be an efficient tool for investigation of TC genesis and the role of desert dust in this process (as well as a benchmark model for calibration of bulk parameterization schemes).

The results reported in this study should be considered as only preliminary. First, along with non-developing storms, it is necessary to simulate genesis and further evolution of TDs developing into hurricanes. The correct prediction of evolution of developing TDs will serve as a justification of the model ability to reproduce TC genesis. In these simulations the role of dust outbreaks will be also investigated. Note that the resolution used in the present study was 3 km which is not high enough to resolve comparatively small clouds, especially at the TC periphery. The use of supercomputers would allow simulations of tropical storms with even higher resolution to obtain, quite possibly, even greater sensitivity to aerosols.

Comparison of the results obtained using the spectral bin microphysics with those obtained using different bulk parameterizations implemented in WRF is also an important topic and will be performed in the next study.

Acknowledgements. The study was supported by the Israel Science Foundation (grant 140/07) and scientific project HUMP (Project FY2008-06-16). The authors express their deep gratitude to Prof. Baik and to H. Zvi Kruglyak for help in solving of computer problems.

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