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5 **High Resolution WRF Simulations of Hurricane Irene: Sensitivity to Aerosols and Choice**
6 **of Microphysical schemes**
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8 A. Khain, B. Lynn and J. Shpund
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10 Department of Atmospheric Sciences, The Hebrew University of Jerusalem, Jerusalem
11 91904, Israel
12

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22 Communicating author : Alexander Khain, Department of Earth Sciences, The Hebrew
23 University of Jerusalem, Jerusalem, Israel. email: alexander.khain@mail.huji.ac.il
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Abstract

Recent studies have pointed to the possible sensitivity of hurricanes to aerosols via aerosol effects on microphysical and thermodynamic processes in clouds. Hurricane Irene, occurring in August 2011, is an excellent case study for investigating aerosol effects on tropical cyclone (TC) structure and intensity: it moved northward along the eastern coast of the United States, and weakened much faster than was predicted by the National Hurricane Center. Moreover, the minimum pressure in Irene occurred, atypically, about 40 h later than the time of maximum wind speed. In this study, we simulate Hurricane Irene with 1-km grid spacing using Spectral Bin Microphysics (SBM) and various bulk microphysical schemes in WRF. Simulations with SBM showed that aerosols penetrating the eyewall of Irene from the Saharan Air Layer (SAL) led to an intensification of convection at Irene's eyewall and to a deepening of the hurricane. When Irene moved along the eastern coast of the United States, continental aerosols led to an intensification of convection at Irene's periphery, which interfered with the re-forming of the inner eyewall and to Irene weakening. Sensitivity tests using different "bulk" microphysics schemes indicated a large dispersion of simulated minimum pressure and maximum wind between different simulations. This showed that the simulated hurricane intensity was very sensitive to microphysical processes. Moreover, in consequence, forecast hurricane intensity was highly dependent on the choice of microphysical scheme. New bulk-parameterization schemes simulated the tropical storm intensity of Irene reasonably well. Most bulk schemes that used saturation adjustment indicate the weak sensitivity to aerosols that prevents them from precisely predicting the time evolution of TC intensity and structure.

Key words: *intensity of hurricanes, microphysics, cloud-aerosol interaction, numerical modeling*

1. Introduction

This study investigates factors affecting the intensity of Hurricane Irene (2011). Irene moved northwestward across the Bahamas, and then northward along the eastern US coast, grazing Cape Lookout, North Carolina. Irene was upgraded to hurricane status on 22 August 2011, reaching Category 3 and then downgraded to tropical storm status just south of New Jersey on 28 August 2011. The intensity of Irene was not well forecast. Irene was forecast to be a Category 3 storm through landfall in North Carolina, while in reality Irene weakened to a Category 1. It is notable that its minimum surface pressure (942 mb) occurred about 40 hours later (06 UTC 26 August) than the time of its maximum wind (105 kt; 12 UTC 24 August). This nontrivial behavior of wind and pressure in Irene means that the decrease in Irene's central surface pressure (TC deepening) was accompanied by a decrease in the radial pressure gradient leading to a decrease in the maximum wind (TC weakening). Such storm behavior indicates that there was an increase in the radius of the maximum wind.

Indeed, according to the National Hurricane Center (NHC) synopsis, a developing rainband seen (their Fig. 5; 23:23 UTC 25 August) in the northeast quadrant outside the eyewall gave rise to the formation of the second eyewall and to the suppression of the internal eyewall of Irene (http://www.nhc.noaa.gov/data/tcr/AL092011_Irene.pdf). As noted, Irene was anticipated to remain a Category 3 hurricane through landfall in North Carolina, given that she was forecast to remain in an environment of relatively light wind shear while moving over a warm ocean. However, the new eyewall wall did not shrink, and the radially wind profile became very broad. Reconnaissance from 26 – 27 August found that the radius of hurricane-force winds reached 80 n mi in the northeast quadrant. Irene's structure was characterized by continuing rainband development, resulting in a broad and diffuse wind field that slowly decayed until weakening to a tropical storm.

As Irene moved along the coastline for a long time, we assume that her intensity and structure was affected by continental aerosols penetrating from the land to the tropical cyclone (TC) circulation. The entrance of these aerosols accords very well with the observed pattern of the aerosol optical depth (**Figure 1**). Besides, Irene crossed a wide band of Saharan dust stretching from 15° latitude to 30° latitude (based on an examination of MODIS data). The depth of the Saharan Air Layer (SAL) was approximately equal to 4 km.

Numerical simulations using spectral bin microphysics (SBM) and bin-emulating schemes showed that the penetration of aerosols into the periphery of a TC prior to landfall leads to the intensification of periphery rainbands and to a subsequent TC weakening (Khain et al., 2008b, 2010; Cotton et al., 2007; Carrio' and Cotton, 2011; Rosenfeld et al., 2012; Herbener et al., 2014). Herbener et al. (2014) also showed that the penetration of aerosols into the eyewall of a TC can lead to TC intensification.

The goals of the study are: a) to investigate whether the impact of aerosols from the US mainland and the SAL was one cause of the unusual changes in the observed minimum surface pressure and maximum wind; b) to check whether existing microphysical schemes are able to simulate the observed TC evolution; and c) to demonstrate the sensitivity of TC intensity to microphysical processes.

The structure of the paper is as follows. In Section 2 we present a brief comparison of different approaches to the description of cloud microphysics. In Section 3 the main aerosol effects on deep convective clouds are discussed. Section 4 describes the experimental design. Section 5 shows the results, while a discussion and conclusions are presented in Section 6.

2. Two main approaches to the description of microphysical processes

In cloud-resolving models, two main approaches to the description of microphysical processes are used: *bulk-microphysics parameterization* (hereinafter, bulk parameterizations, or

bulk schemes) and *spectral (bin) microphysics* (hereinafter, SBM). Despite the fact that in both approaches the same microphysical processes are described and have similar outputs (cloud and rain water mixing ratios or contents, CWC and RWC, respectively, cloud ice content, precipitation rates, etc.), the two approaches diverge quite notably with regard to the description of microphysical processes.

Bulk-parameterization schemes aim to represent the most general microphysical cloud properties using a semi-empirical description of particle size distributions $f(m)$ (hereinafter, PSD) (here m is particle mass). The computational efficiency of bulk-parameterization schemes lies in their consideration of microphysical equations, not for PSDs of different hydrometeor types (e.g., cloud droplets, raindrops, ice crystals, aggregates, graupel, hail) themselves, but for a number of PSD moments. The k -th moment of the PSD is defined as $M^{(k)} = \int_0^{\infty} m^k f(m) dm$ with k primarily an integer value. Schemes that use only one moment (typically, the mass densities of variables hydrometeors, $k=1$) are known as one-moment or single-moment schemes, while schemes using two moments (typically, number concentrations, $k=0$ and mass contents, $k=1$) are known as two-moment schemes. Less frequently, three-moment bulk schemes are used. In these, the variables are number densities (concentrations), mass densities, and radar reflectivity ($k=2$). PSDs are typically represented in the form of gamma or exponential distributions. Since the number of moments used in bulk schemes is less than the number of parameters determining the PSD shape, some parameters are usually fixed. A major advantage of using gamma (and exponential) functions is that microphysical process rates can be then derived in a straightforward manner (Morrison and Gettelman, 2008).

The analysis of observational data in clouds of different types shows that real local PSDs have a complicated shape and often cannot be approximated well by gamma or exponential distributions (Costa et al., 2000; Geoffroy et al., 2010). An analysis of observational data by

Tampieri and Tomasi (1978) and later on by Dooley (2008) showed that the shape of drop size distributions (DSDs) in clouds evolves with time and height in such a way that all parameters of gamma distributions should be changed to preserve the approximation of DSDs. Taking into account the relationships between different parameters of gamma distributions may substantially improve the accuracy of the bulk scheme (Thompson et al., 2004; 2008; Thompson and Eidhammer, 2014). In contrast, fixing one or several parameters in the gamma distribution may lead to substantial errors in the representation of cloud microphysics.

The specific feature of microphysical processes is a dramatic, highly non-linear dependence of rates of microphysical processes on particle size. The averaging of PSD over particle spectra in bulk parameterizations significantly reduces the sensitivity of bulk schemes to particle size. The necessity of parameterizing microphysical processes with fewer PSD moments leads to serious drawbacks. For instance, most bulk schemes used in this study apply the saturation adjustment hypothesis to calculate the production of new cloud mass content. According to this hypothesis, the process of diffusion growth of drops in warm and mixed-phase clouds leads to zero supersaturation. The assumption behind the saturation adjustment hypothesis is that air ascends along the moist adiabat. The typical procedure of the saturation adjustment is the following. After advection substep new temperature T and mixing ratio q are calculated. The mixing ratio is compared to the saturation value $q_s(T)$. If $q > q_s(T)$, (i.e. supersaturation takes place) a simple iteration procedure is performed during which q decreases and T increases, so to the end of the procedure $q_{new} = q_s(T_{new})$. Then the difference $q - q_{new}$ is added to cloud water content. This procedure does not depend on aerosols, so CWC formed at the stage of diffusional growth will be the same at any CCN concentration. New parameters of the gamma distributions are then calculated using the droplet concentration obtained as a result of nucleation and the new drop mass content. Note that in convective updrafts, supersaturation never falls to zero, and in contrast, may reach several percents as a result of an increase in vertical velocity and a decrease

in drop concentration by collisions [*Khain et al.*, 2012; *Prabha et al.*, 2011]. This contradicts clear theoretical considerations, according to which supersaturation in clouds developing in clean air is larger than in clouds developing under polluted conditions (Pinsky et al., 2014). This also contradicts observations showing lower CWC in clean maritime conditions clouds (Prabha et al. 2011). Since supersaturation decreases with an increase in the droplet concentration (i.e. CCN concentration) the effect of this procedure is equivalent to the utilization of infinite large CCN concentration. Thus, the utilization of saturation adjustment overestimates the production of cloud water content (CWC) and the latent heat release. The only bulk scheme considered in this study that does not use saturation adjustment is NSSL-2MOMCCN (Mansell et al., 2010; Mansell and Ziegler, 2013).

In different schemes the accretion rates differ by several orders of magnitude (Gilmore and Straka, 2008). Especially important may be the errors in the description of particle sedimentation, leading to errors in the spatial distribution of microphysical parameters (Milbrant and Yau, 2005a,b; Milbrandt and McTaggart-Cowan, 2010). The errors are especially large in one-moment schemes. In many studies, significant errors in the description of evaporation, melting, freezing and other processes are mentioned in the literature (e.g., Li et al., 2009) as leading to divergent forecast results. Table 1 shows the main properties of the bulk-parameterization schemes used to simulate Irene. Further details are provided later where needed.

The SBM approach aims to simulate as accurately as possible cloud microphysical and precipitation processes in clouds of different types. Fundamental to bin-microphysics is the calculation of PSDs by solving explicit microphysical equations. Hence, no a priori information about the form of PSDs is required or assumed. Instead, PSDs are calculated, for instance, on a finite-difference mass grid containing several tens to several hundred mass bins in different models. A massive increase in computing power has made it possible to explicitly describe cloud

microstructure using the knowledge accumulated in cloud-physics processes. Numerous studies have proved the ability of SBM to simulate accurately individual clouds (Fan et al., 2009; Benmoshe et al., 2012), mesoscale rain events (Lynn and Khain, 2007), hail storms (Khain et al., 2011), hurricanes (Khain et al., 2010) and climatic properties of cloudiness under different aerosol conditions (Fan et al., 2014), even in real time under certain model configurations.

Equations used in bin-microphysics do not depend on particular meteorological situations. Thus, the advantage of the SBM is its universality; that is, the same scheme can be used without any modification for simulations of different atmospheric phenomena ranging from stratiform arctic clouds to tropical cyclones. By way of contrast, in bulk schemes microphysical parameters may need to be chosen based on the cloud system being simulated.

The number of equations to be solved in SBM schemes is proportional to the number of bins and the number of hydrometeor types (including aerosol). Typically SBM has about 200-300 prognostic variables, while in comparison bulk schemes have 6-15 variables. Therefore, computational requirements in these schemes are substantially higher than in bulk parameterization schemes. Accordingly, the bin-microphysics approach requires ~ 5 to 20 times more computer time than current bulk-parameterization schemes, depending on the complexity of the particular bulk scheme. However, the comparative gain in computer throughput in bulk schemes comes at the cost of a loss in accuracy and a decrease in sensitivity to such fine effects as the effect of aerosols on cloud drop growth and water-ice interactions. Table 2 summarizes the differences in the representation of the main microphysical processes in SBM and bulk-parameterization schemes.

In this study, we used the SBM scheme described by Khain et al. (2010). The scheme explicitly calculates the PSD of drops (including cloud droplets and raindrops), ice, aggregate (snow) and graupel/hail distributions. Each PSD is defined on logarithmically equidistant mass grids containing 33 bins. Droplet nucleation is calculated using the Kohler theory, as described

in detail by Khain et al. (2000). Diffusion growth/evaporation of drops and deposition/sublimation of ice particles are calculated by solving the coupled equation for diffusional growth and supersaturations over water and ice (Khain and Sednev, 1996). Collisions are calculated by solving a system of stochastic collision equations, as described in detail by Khain and Sednev (1995) and Khain et al. (2004). The collision efficiencies between droplets, as well as between drops and ice particles, were calculated using the precise superposition method of Pinsky et al., 2001 and Khain et al., 2001. The collision efficiencies depend on height. The nucleation of ice crystals is described using the parameterization formula of Meyers et al. (1992), with modifications described by Khain et al. (2000). The processes of sedimentation are explicitly described. Fall velocities are different for different mass bins and different hydrometeor classes, and fall velocities increase with height because of a decrease in the air density. In the SBM scheme that is used, melting is described in a simplified form. It was assumed that ice particles of different sizes have different time scales of melting. The melted water was assumed to be immediately shed. The SBM is described further in Khain et al. (2004).

We refer the reader to the original papers for a detailed description of the schemes. The treatment for aerosols is described in the next section.

3. Aerosol effects

Latent heat release in clouds is the main energetic source of atmospheric motions. This is especially true for hurricanes. Thus, a special branch of cloud physics was developed that aims at the investigation of the effects of aerosols on precipitation formation and on rain amounts. Early hypotheses put forward that an increase in aerosol concentration leads to the suppression of precipitation formation, and to an increase in the liquid water path (Albrecht, 1989). Subsequent studies showed that the picture is more complicated, and that many dynamic and thermodynamic processes are involved in the process of precipitation formation and in the determination of surface precipitation (Khain, 2009; Tao et al., 2012)

Three aspects of aerosols affect cloud microphysics and dynamics: aerosol concentrations, aerosol sizes (characterized by the size distribution), and the ratio of soluble and non-soluble fractions. Aerosols affect precipitation and the dynamics of clouds and cloud systems by means of different physical mechanisms. The relative contributions of these mechanisms depend on spatial scales of the phenomena considered and cloud types. The main microphysical effect of aerosols on single deep convective clouds is the control of droplet concentration; namely, an increase in CCN concentration increases droplet concentration. When droplet concentration is low, the supersaturation over water in the clouds is high, and droplets rapidly form raindrops at comparatively small distances above cloud base. These droplets grow rapidly, giving rise to the onset of warm rain. New droplets are efficiently collected by raindrops, which has the effect of shortening the residential time of droplets in clouds.

When droplet concentration is high, the supersaturation over water is low and droplets grow slowly. Since the rate of collision for small droplets is low (Pruppacher and Klett, 1997), small droplets reach altitudes of several kilometers above cloud base before the first raindrops are formed (Andreae et al., 2004; Khain et al., 2005; van den Heever et al., 2006; Freud and Rosenfeld, 2012). Thus, an increase in the CCN concentration leads to a delay in raindrop formation and increases the residential time of droplets in clouds. The delay leads to an elevation of the height of raindrop formation, sometimes by several kilometers (Freud and Rosenfeld, 2012; Benmoshe et al., 2012).

At the stage of diffusion growth, CWC in a polluted ascending volume of air will be larger than in an analogous volume of clean air (Pinsky et al., 2014). This means that there will be a larger rate of latent heat release in the polluted air mass.

The increase in the rate of condensation and the intensification of freezing with increasing CCN concentrations leads to an increase in the latent heat release, which in turn increases the vertical velocity and sometimes cloud top height and the area of cloud anvils (Fan et al., 2014). This dynamical effect, especially pronounced for convective clouds with a warm cloud base and

high altitude of freezing levels, is known as aerosol-induced *convection invigoration*. Aerosol-induced convection invigoration has been reported in many observational and numerical studies (Khain 2009; Rosenfeld et al., 2008; Tao et al., 2012).

There is a substantial difference in the treatment of aerosol effects in the SBM used in the study and in bulk-parameterization schemes. In SBM, the size distribution of aerosols serving as cloud condensational nuclei (CCN) is defined on a mass grid containing a few tens of doubling mass bins (in the current study, 33 bins were used). The largest CCN size corresponds to a 2- μm radius droplet. These aerosols are advected like other hydrometeors. Using the supersaturation calculated in each grid point and at each time-step, a critical radius of CCN is calculated in the grid points using the Köhler theory. CCNs with radii exceeding the critical radius are converted to droplets, so new droplets are nucleated. The corresponding bins in the CCN size distribution at that grid location become empty. To activate the next portion of the CCN spectrum, supersaturation should exceed the values that a cloud volume experienced in the past. The calculation of the size of newly nucleated droplets is performed according the Köhler theory, with some specific treatment of large CCN, as described in detail in Khain et al. (2000). Such aerosol treatment allows for the utilization of different initial spatial distributions of CCN.

In some bulk-parameterization schemes, the effects of aerosols on cloud microphysics are taken into account by different choices of DSD shapes and different rates of autoconversion (Lin et al. (1983). In most bulk-parameterization schemes, the concentration of nucleated droplets is calculated using the Twomey formula $N = N_o S^k$, where N_o is the concentration of activated CCN (newly nucleated droplets) at supersaturation of 1%, and k is a slope parameter. In most schemes, N_o and k are prescribed parameters that do not change with time or space (Thompson et al., 2008; Morrison, 2005, 2009). The schemes by Thompson and Eidhammer (2014) and by Mansell and Ziegler (2013) contain an explicit prediction of aerosol concentration. In Thompson and Eidhammer's (2014) scheme, the droplet concentration is determined by applying lookup tables created using the Köhler activation theory and assuming a lognormal CCN distribution.

Mansell and Ziegler (2013) apply a parameterization procedure that allows the calculation of the fraction of available CCN to be activated to droplets depending on vertical velocity at cloud base. A decrease in the aerosol concentration by droplet nucleation is taken into account in both Thompson and Eidhammer's (2014) and Mansell and Ziegler's (2013) schemes. Similar to the SBM, droplet nucleation in the scheme by Thompson and Eidhammer (2014) can occur both at cloud base and within the cloud.

4. The experimental design

The simulations used WRF (version 3.4. for SBM simulations and 3.6.1 for some simulations with bulk schemes) with three nested grids. The grid spacing of the outer grid was 9 X 9 km, where there were 700 X 700 grid elements. The intermediate moving nest had 3 X 3 km grid spacing and 300 X 300 grid elements. The finest 1 X 1 km domain contained 500 X 500 grid elements. In each of the simulations there were 30 vertical layers, with the finest resolution in the boundary layer. Note that the same SBM scheme was used on all grid meshes.

The simulations began at 00 UTC 21 August 2011 and finished at 00 UTC 29 August 2011. The initial position of the storm was at 15N, 59W (the extreme northeastern Caribbean Sea), and the final position was 44.N, 72.1W 192 hours later when it became extratropical over the northeastern US. Each of the simulations used the GFS analysis fields as boundary conditions, which included grid nudging of the temperature, winds, and specific humidity above the boundary layer in the outer 9-km grid. The 2.5 order closure BL scheme "MYNN2" was used for the description of the boundary layer (BL) (Nakanishi, 2000; Nakanishi and Niino, 2009). This model configuration is used in all simulations with the different microphysical schemes.

Comparatively high initial values of sea surface temperature (SST) were associated with a very low depth of the oceanic mixing layer of just 15 m (<http://maracoos.org/irene/?p=305>). The mixing of the surface ocean water layer during the storm's movement led to a decrease of 4-5 °C in SST (Ginis, personal communication, 2013). Ocean coupling leading to the decrease in SST

is simulated using a one-dimensional ocean model (Davis et al., 2008). A comparison with observed sea surface temperatures showed that the model simulated quite well the SST cooling. Still, the use of a 1-D ocean model led to some underestimation of the SST decrease below the TC center because of lack of upwelling.

The simulations were carried out with both SBM and several bulk-parameterization schemes listed in Table 1, as previously noted.

The effects of aerosols on cloud microphysics were taken into account according to various assumptions made in the specification of aerosols or cloud drop concentration within the different microphysical schemes. In simulations with SBM, two initial spatial distributions of aerosols were used, as shown in **Figure 2**. The simulations *SBM_MAR* had only maritime aerosols (Fig. 2 left). In the simulation referred to as *SBM_COMP*¹, there were continental aerosols over the land and maritime over the clean sea, while the SAL band covered part of the sea and land (Fig. 2, right).

The initial size distribution of dry aerosol particles serving as CCN was calculated using the empirical dependence of CCN concentration N on supersaturation as $N = N_o S_w^k$. The CCN of maritime conditions set $N_o = 100 \text{ cm}^{-3}$ at 1% saturation, and $k = 0.7$ over the entire computational area. Continental refers to $N_o = 4000 \text{ cm}^{-3}$ and $k = 0.9$. The dust concentration was evaluated using CALIOP lidar data (Lynn et al., 2015). The maximum droplet concentrations in the COMP and in maritime aerosols were several hundred and several tens per cm^3 , respectively.

The SAL band was comprised of a mixture of “maritime” and “continental” aerosols. Here, CALIOP lidar data was used to determine that a constant aerosol “mixed” layer extended from about the surface through a 3-km depth stretching across the length of the

¹The abbreviation “COMP” stands for the word complete – as compared to a simulation with either only continental aerosols over land (or land and sea)

outer domain from 10° latitude to 30° latitude. Outside of the dust band, the CCN concentration was assumed as in MAR.

Concentration of aerosols was assumed constant in the lower 2-km layer and decreased exponentially with height above the 2-km altitude with a characteristic spatial scale of 3 km. Above the SAL band, the aerosol concentrations were initially zero.

There are no explicit aerosol sources in the model within the computational area, with the exception of aerosols at the lateral boundaries, where air flow was directed inside. Over the 8 days of simulation, the aerosols initialized in the large region advect into the region of interest, thus emulating continuous sources. Aerosols are not radiatively active in these simulations. There is only one sink of aerosols, namely, nucleation scavenging. Two advanced bulk schemes (Thompson “Aerosol-Aware” option and *NSSL-2MOMCCN*) calculate the drop mass and number concentration based on aerosol concentration. The set-up was the same as in Fig. 1, but an intermediate aerosol concentration was added where N_o was set to 2000 cm^{-3} . The simulations were referred to as *THO_A-Mar*, *THO_A-Int*, and *THO_A-COMP*, as well as *NSSL-COMP* and *NSSL-MAR* ($N_o=2000 \text{ cm}^{-3}$ and $N_o=100 \text{ cm}^{-3}$), respectively. In addition to *NSSL-COMP* and *NSSL-MAR*, two supplemental sensitivity simulations were performed with *NSSL-2MOMCCN*. The goal was to test the sensitivity of this scheme, which does not use saturation adjustment, to treatment with aerosols. The first one, referred to as *NSSL-N*, does not include a decrease in CCN concentration during the process of droplet nucleation. In the second one, referred to as *NSSL-2MOMCCN-NBC*, the concentration of aerosols at the outermost boundary of the largest computational domain was assumed to be zero. This resulted in the concentration of CCN decreasing with time in the computational area, beginning with the boundaries. When Irene approached the coastline of the USA, the CCN concentration decreased substantially over the land in this simulation. Thus, a comparison of results obtained

in simulations *NSSL-2MOMCCN* and *NSSL-2MOMCCN-NBC* indicates the sensitivity of TC intensity and structure to continental aerosols in simulations using this bulk scheme.

Another (and older) version of the Thompson microphysics and the Morrison scheme provide the user with the ability to specify cloud drop concentration. Based on the suggestions within each module, droplet concentrations of 100 cm^{-3} in maritime (Thompson-MAR) and 400 cm^{-3} in continental (Thompson_Con) were specified throughout each domain. In simulations with the "Morrison" scheme, the droplet concentrations were set equal to 250 cm^{-3} in Morrison_MAR and to 1000 cm^{-3} in Morrison_CON. Accordingly, we performed the simulations "Thompson-MAR", "Thompson-Con", "Morrison_MAR" and "Morrison-Con," where the drop concentrations in clouds were assumed constant throughout the domain. In other schemes of Table 1, the parameters were used as given in the WRF definitions, so no distinction between maritime and continental cases was made.

5. Results

5.1 Simulation of TC intensity and structure using SBM: effects of aerosols

Figure 3 shows time dependencies of minimum pressure and maximum wind speed in the simulations *SBM_MAR* and *SBM_COMP*. These two simulations differ only with respect to concentration and spatial distributions of aerosols. One can see that aerosols in *SBM_COMP* led to a weakening of the maximum wind speed after 90 h, in agreement with the observations. So, while the model using *SBM_MAR* predicts the lowest minimum pressure and the highest maximum wind at the same time instance, the *SBM_COMP* managed to predict the observed time shift between the highest maximum wind and the lowest minimum pressure.

Figures 4 to 6 illustrate from simulations *SBM_COMP* and *SBM_MAR* the physical mechanisms by means of which aerosols in situ might have led to the discrepancy in the time of maximum wind and minimum pressure. **Figure 4** shows the fields of rain water content (RWC)

near the oceanic surface in simulations *SBM_MAR* and in *SBM_COMP* at time instances 120h, 138h and 144 h. One can see that during this time period, the radius of the eyewall in the *SBM-MAR* simulation increased only minimally. In contrast, a *partial* secondary eyewall arose in *SBM_COMP* with a radius much larger than the radius of original eyewall. The competition between the two eyewalls results in the suppression of the inner one, which led to a significant increase of the radius of the TC inner zone.

We attribute the changes in the structure of the inner core of Irene to the effects of aerosols, which intensified the rainbands at the outer periphery of the inner TC zone, fostering intensification of the outer eyewall. **Figure 5** shows the fields of both cloud droplet and aerosol concentration in *SBM_COMP* during the time period 120-144h. One can see that aerosols penetrated the internal TC area and intensified rainbands along the trajectories of aerosol particles. The microphysical effect of aerosols is seen by the increased concentration of cloud droplets along the aerosol tracks. These invigorated rain bands formed the secondary eyewall.

Figure 6 shows the time dependence of the inner radius of eyewall in *SBM_COMP* and *SBM_MAR*. One can see that the radius of maximum wind increased until ~100 km in *SBM_COMP*. In contrast, the radius of the hurricane winds in *SBM_MAR* varies between 30 km and 60 km.

The changes in the eyewall radius simulated in *SBM_COMP* accords well with the observed changes in Irene's eyewall radius (http://www.nhc.noaa.gov/data/tcr/AL092011_Irene.pdf).

Also of interest is that the impact of aerosols depended on the stage of Irene's development. Initially, the eyewall radius was larger and the simulated Irene was weaker during the time period 50-70 h in *SBM-MAR* than in *SBM_COMP*. Later on, the eyewall radius was smaller and the minimum pressure was lower in *SBM-MAR* (Figs. 3, 6). Note that the change in the structure of the inner core in the *SBM_COMP* began at about 100 h was characterized by the

formation of a pronounced secondary eyewall, in contrast to *SBM-MAR*, where a single eyewall remained.

The aerosol effects on the TC structure are illustrated in more detail in **Figure 7**. This figure shows the radial dependencies of azimuthally averaged maximum wind and the minimum surface pressure in the *SBM_COMP* and *SBM_MAR* at various times in the simulations (or stages in the simulated development of Irene). During Irene's penetration into the Saharan dust belt (50 h), aerosols penetrated the eyewall and led to an intensification of convection in the central TC zone.

Figure 8 compares the vertical velocity fields in *SBM-MAR* and *SBM_COMP* at $t=60$ h at 9 km altitude. One can see that aerosols penetrated the eyewall substantially and led to an increase in vertical velocity. This is evidence of aerosol-induced convection invigoration, when aerosols lead to an increase in the latent heat release and the buoyancy in cloud cores. The intensification of an idealized TC caused by the penetration of CCNs into the eyewall was also reported by Herbener et al. (2014). As a result, Irene in the *SBM_COMP* was stronger than in the *SBM-MAR* (see Fig. 3 and 6). Later on, rain bands at the TC periphery in *SBM_COMP* approached the inner core of Irene and increased the eyewall radius ($t=70$ h). The influx of aerosols from the band and land continued to support convection within the secondary eyewall leading to its intensification. Toward 100 h, a second wind maximum clearly appeared in the azimuthally profile of maximum wind. The changes in the internal TC structure led to a decrease in the radial pressure gradient, that is, to a decrease in the maximum wind speed (as seen in Fig. 3). At the same time, the total latent heat within the central TC area increased, which lead to a decrease in pressure at 100 h, as compared to 70 h.

Hence, changes in the structure and the formation of the secondary eyewall and a weakening of the old eyewall in the *SBM_COMP* experiment accounts for the simulated time shift between maximum wind speed and minimum pressure. The last panel in Fig. 7 (~ 150 h) shows that the

internal eyewall fully disappeared in the simulation *SBM_COMP*. As noted, this structural transformation of the inner TC core and the overall lack of contraction of the new eyewall was reported in the NHC summary. In contrast, in the simulation *SBM-MAR*, no such transformation of the inner TC structure took place. There was no division of the eyewall, and Irene in the *SBM-MAR* was stronger at this stage than that in the *SBM_COMP* (Fig. 7).

5.2 Simulation of TC intensity and structure using bulk-parameterization schemes

We begin with the results obtained using bulk schemes with the Morrison, Thompson and NSSL schemes, which turned out to best predict the intensity of Irene.

Figure 9 shows the time dependencies of the minimum pressure and the maximum wind velocity in simulations carried out with the Morrison, Thompson and NSSL schemes, which allow for changing aerosol and/or droplet concentrations. Results are also presented for comparison with the SBM, as well as the simplest warm rain Kessler parameterization.

Analysis of these dependencies shows that the minimum pressure (fig 9a) is predicted reasonably well by all three parameterization schemes, regardless of whether or not aerosols are specified. An exception is the *NSSL-MAR* simulation that predicts too low minimum pressure. In simulations with an *a priori* prescription of droplet concentrations (Fig 9b) the predicted wind speed is larger than the observed wind speed. The Thompson scheme predicts lower wind speed than that of Morrison, but the difference is not large. The effects of droplet concentrations are very small for these schemes, much weaker than those in SBM simulations. It is interesting that the simple warm rain Kessler scheme predicts maximum wind reasonably well. A discussion of this scheme is presented below.

Figure 9c shows that the *Tho_A-COMP* predicts the maximum wind better than the schemes by Thompson and Morrison. The sign of the response of this scheme to an increase in aerosols is similar to that of SBM: maximum wind during the time period $t \sim 120$ -130 decreases

with an increase in the aerosol concentration. At the same time, the response of the bulk-parameterization scheme to aerosols is substantially weaker than that of SBM. For instance, the maximum wind in *Tho_A_MAR* and *Tho_A-COMP* was only 5-7 kts. As a result, even in the *Tho_A-COMP*, the maximum wind at t=130 h remains larger than that at t=90 h, so the observed time shift between the wind maximum and the pressure minimum was not reproduced. It is remarkable that the NSSL scheme's sensitivity to aerosols turned out to be much higher than that of the *Tho_Aeros-Aware*, and is very close to the sensitivity of SBM. The high sensitivity of the NSSL scheme to aerosols was evident already in Fig 9a. The difference in the minimum pressures between *NSSL_COMP* and *NSSL_MAR* exceeds 10 mb at t=110 h. The difference between maximum wind speeds in *NSSL_MAR* and *NSSL_COMP* was about 20 kt, which is close to the difference between *SBM_MAR* and *SBM-COMP*. Moreover, the *NSSL-COMP* predicted the times of reaching maximum wind speed at ~110 h, in contrast to ~130-140 h, as in other bulk schemes. The results of *NSSL_COMP* turned out to be the closest to those of SBM. However, the NSSL scheme did not reproduce well the observed time shift between the wind maximum and the pressure minimum. A comparison of time dependencies of maximum wind simulated in *NSSL_COMP* and *NSSL_COMP-NBC* shows that a decrease in the concentration of continental aerosols led to a shift in the maximum wind in *NSSL_COMP-NBC* to t=140 h. Such a shift resembles that which occurred in *SBM-MAR*, and indicates that a penetration of continental aerosols is responsible for the earlier TC weakening.

To illustrate the role of aerosols in the *Tho_Aerosol Aware* scheme we present **Figures 10 and 11**, which are analogs of Figures 3 and 6 plotted for the SBM runs.

Comparison of Figure 10 and Figure 4 shows that: a) the areas of strong precipitation in *Tho_Aerosol_Aware* are lower than in SBM; b) there is an important difference in the structure of RWC fields, which reflects the difference in the structure of the inner core. Indeed, in SBM simulations there is a clear difference between *SBM_COMP* and *SBM_MAR*, with a single

eyewall in the *SBM-MAR* and with formation of two eyewalls in *SBM_COMP* and the dissipation of the inner one. In contrast, the TC structures in *Tho_A-COMP* and *Tho_A-MAR* are quite similar: in both simulations, two eyewalls arise and remain until the end of the time period considered. At the same time, the innermost eyewall in *Tho_A-COMP* seems to be more suppressed, that is, aerosols tends to act similarly in the SBM and the *Tho_Aerosol Aware* schemes, but their impact on TC structure is not as strong as SBM. Figure 6 also shows the time dependence of the inner radius of the eyewall in the Thompson and Morrison simulations. Among them, the variation of hurricane winds with a radius in the simulation *THO_A-COMP* was the most similar to that of the *SBM_COMP*. However, the sharp increase in the radius in the *THO_A-COMP* begins only at $t=130-140$ h, that is, about 40 h later than in the *SBM_COMP*. An increase in the inner radius at $t=140$ h also takes place in the *THO_A-MAR*, but this also takes place with a relative time delay, and the maximum radius reached is less than in *THO_A-COMP*, but more than in *SBM_MAR*.

As shown in **Figure 6**, substantial changes in the TC structure take place at 140 h both in *THO_A-MAR* and *Tho_A-COMP*. The increase in the radii, which may imply a weakening of the inner eyewall, leads to a very sharp decrease in the maximum wind, seen in Fig. 9c at $t=140$ h.

Figure 11 shows the radial dependencies of azimuthally averaged maximum wind in *THO-MAR*, *THO_A-INT* and *Tho_A-COMP*. The comparison of radial profiles in Fig. 11 with those in Fig. 7 shows again that the bulk scheme's simulation of pressure and wind speed is substantially less sensitive to changes in aerosol concentrations than SBM. Aerosols, however, render the secondary eyewall more pronounced (see the panel, $t=120$ h in Fig. 11) in both the *THO_A-INT* (and *THO_A-COMP*) and SBM. The later changes in the radius of eyewall in the *THO_A-COMP* as compared to the *SBM_COMP* (Figure 6) were associated with delayed weakening in the maximum wind. This delay does not allow for a reproduction of the observed shift between the wind maximum and the pressure minimum.

492 As regards to the simulations with the non-aerosol aware versions of the Thompson and
493 Morrison schemes, the results were almost insensitive to the concentration of aerosols or droplets.
494 In the case of Morrison_CON, the radius of the inner eyewall reached ~80 km, vs 60 km in
495 Morrison_MAR (Fig. 6). Thus, although the minimum surface pressure was similar to the SBM
496 and the *THO_Aerosol Aware* simulations, the sensitivity of pressure and maximum wind to
497 aerosols was much lower.

498 Figure 12 shows fields of RWC in *NSSL-MAR*, *NSSL-COMP* and *NSSL-COMP_NBC*. This
499 figure is an analog of Figures 4 and 10. Results of the supplemental simulation *NSSL-*
500 *COMP_NBC* indicate that a lack of continental aerosols resulted in the radius of the eyewall
501 remaining small (and maximum wind remaining high) till at least $t=144$ h. In *NSSL-MAR* the
502 radius of the eyewall does not change much with time. There is a general increase in the area of
503 the internal TC zone without intensification of convection and precipitation at the TC periphery. In
504 contrast, the RWC fields in *NSSL-COMP* indicate a larger eyewall radius, which agrees with the
505 results of SBM.

506 Figure 13 is the analog of Figures 7 and 11. One can see that the aerosol-induced differences in
507 the radial profiles of azimuthally averaged surface pressure and maximum wind in the *NSSL-*
508 *COMP* is closer to those in SBM than other schemes. As in SBM, *NSSL-MAR* does not show the
509 formation of a secondary eyewall and a dissipation of the original one. The area of the inner core
510 monotonically increases, which leads to a decrease in the radial pressure gradient and maximum
511 wind. The results obtained in *NSSL-COMP-NBC* indicate the existence of a small eyewall radius
512 and high TC intensity till $t=150$ h. These profiles show, once again, the significant sensitivity of
513 the TC structure to aerosols simulated by *NSSL-COMP*.

514 At this point, we will consider the results of the simulations of Irene's intensity using all the
515 bulk schemes. To show how droplet concentration can affect microphysical structure and TC
516 intensity, we added a supplemental experiment *NSSL-N*, which is similar to *NSSL-COMP*, but

the concentration of CCN was not decreased after droplet nucleation, so droplet concentration increased and reached very high values. **Figure 14** shows the time dependence of minimum surface pressure and maximum wind velocity for simulations with different bulk-parameterization schemes. The various simulations of Irene exhibited high variability in the predicted values of minimum pressure and maximum wind. The values of the minimum pressure varied from 960 mb (CAM) to 920 mb (*NSSL -N* and *SBU_YLIN*) when the observed value was 942 mb. The maximum wind velocity in the simulations varied from 80 kt (CAM) to 140 kt (*NSSL -N*) with the observed value of 105 kt.

The CAM scheme simulated the weakest TC, possibly because this scheme was designed for GCMs with large grid spacing. The CAM has only two prognostic variables: cloud droplets and ice crystals. Moreover, the advection of precipitating particles by the background wind is not taken into account. This computationally fast set of assumptions is apparently unsuitable for simulations of hurricanes with high wind speeds, especially in the case of small grid spacing. The result seems to be trivial, but useful. It shows that the time dependencies of minimum pressure and maximum wind presented in the study are not random, and that the better schemes predict TC intensity more accurately. The results also show that simplifications of a microphysical scheme that are acceptable for large grid spacing become unsuitable for simulations with comparatively high resolution. We believe the results to be of interest because the CAM scheme is used in some NWF models with continuously decreasing grid spacing.

Figure 15 shows the time dependencies of the internal eyewall radius for different simulations of Irene. We see again that the time variation of the radial distances simulated using the Thompson and Morrison schemes, and especially the *Tho_A-COMP*, were most similar to that in the *SBM-COMP*. The rest of the bulk schemes produce substantially different time variations in the radial distances. A typical feature of these bulk schemes is that the eyewall radius remained nearly constant until $t \sim 150$ h, after which the simulated Irene weakened and the eyewall disappeared.

The high variability of the intensity of the model TC to the microphysical schemes shows that TC intensity and structure are highly sensitive to microphysical processes; that is, to the rates and vertical profiles of latent heat produced by different schemes. The simulations also suggest that it might be possible to improve bulk-parameterization schemes, as was done by Thompson and Morrison through their calibration against observations or possibly the results from SBM-type models. For instance, it is necessary to calibrate the rate of autoconversion of cloud droplets to rain under low CCN concentrations to increase the rate of rain production at lower levels and to prevent the penetration of a large mass of cloud droplets to the upper levels.

5.3. Comparison of microphysical fields produced by different schemes

We stressed above that the dynamical response of the simulated Irene to the changes in the CCN concentration (and/or drop concentration) was substantially weaker in the simulations with bulk-parameterizations than in the SBM simulations. Since the dynamical response is just a reflection of thermodynamic changes caused by cloud microphysics, one can expect that the microphysical structures produced by bulk schemes are themselves not sufficiently sensitive to aerosols. The next three figures (**Figs. 16-18**) show different microphysical fields at the time instance of 144 h, which corresponds (approximately) to the maximum intensity of the modeled TCs. While we used azimuthal averaged values of pressure and wind speed (e.g., Fig 7) to investigate differences in the radial TC structure in different experiments, the time averaging of cloud microphysical fields is undesirable because of both high spatial and "age" variability of clouds. As a result of such averaging, clouds at the developing stage will be "averaged" with decaying clouds of quite different structures. In our view, the comparison of instantaneous fields of microphysical values better reveals the difference in cloud characteristics produced by different schemes. Figures 16 and 17 show CWC within cross-sections with lengths of 500 km, i.e., large enough to obtain a statistically significant conclusion.

Note that the characteristic features of cloud fields produced by different schemes remain the same at different time instances.

The cloud-water mixing ratio of cloud water content (CWC), in and of itself acutely sensitive to aerosol concentration, most significantly affects the rates of both warm and cold processes. To compare the sensitivity of different microphysical schemes to aerosols, we present **Figure 16** showing vertical east-west cross-sections in the fields of CWC obtained in simulations in which sensitivity to aerosols or droplet concentrations have been claimed. Note first the similarity of the maximum values of the CWC in the different simulations, which is in line with the fact that these schemes simulated similar minimum pressures and maximum winds.

Still, in the simulations with higher aerosol concentrations, CWC is typically larger than in the simulations with low CCN concentration. At the same time, the difference in the fields of CWC in the bulk schemes with high and low aerosol (or droplet) concentration is much less pronounced than in the simulations using the SBM. Indeed, a) the ratio of the maximum values of CWC in the SBM_COMP and the SBM_MAR exceeds 2. At the same time, said ratio in the bulk schemes is typically below 1.5; b) in the SBM_MAR, no CWC occurs above an altitude of about 5 km, yet in all simulations with bulk schemes, significant supercooled CWC reached 9 km, and even the level of homogeneous freezing (~10 km). Actually, the cross-sections in the CWC fields in low-CCN bulk simulations are much more similar to the cross-section SBM_COMP than to the SBM_MAR. So, Figure 16 shows a much lower sensitivity of bulk schemes to aerosols than SBM. This suggests possible errors in the simulation of the vertical distributions of CWC and other hydrometeors using bulk schemes. The low sensitivity of bulk-parameterization schemes to aerosols accords with their simulated lack of sensitivity of maximum wind/minimum pressure to changes in aerosol concentrations/preset droplet concentration. The exception, perhaps, is the *NSSL* scheme, in which the CWC values in the *NSSL-MAR* are lower than in *NSSL_CON*. While even in this version the CWC in the *NSSL-MAR* reach the 9-km level, the magnitude of the CWC at this level is low. The higher sensitivity

of the NSSL microphysical scheme to aerosols can be at least partially attributed to that fact that this scheme does not use a saturation adjustment process. The differences in microphysical structures produced by NSSL-MAR and NSSL_CON is immediately reflected in the difference in the TC intensity (see Figure 9)

Figure 17 shows the CWC cross-sections plotted for the simulations with WSM6, WDM6 bulk schemes, where these schemes were used "as-is," without a particular reference to maritime or continental conditions. This figure also shows CWC in NSSL-N with a very high droplet concentration. Note again that in all these fields, supercooled CWC reaches the level of homogeneous freezing, which is to be expected in the presence of high CCN concentrations. The CWC field in WSM6 is quite similar to the corresponding fields calculated using other bulk parameterizations shown in Fig. 16. At the same time, the cross-sections of CWC in WDM6 and, especially, NSSL-N indicate much higher CWC maxima than in other microphysical schemes. Overestimation of CWC affects ice microphysics (as shown below), and it was an indicator of very strong convection that in the context of eyewall dynamics led to comparatively low minimum pressure and high maximum wind speeds (Figure 14).

The differences in warm microphysics (e.g., CWC) lead to the subsequent development of differences in cold (ice) microphysics. **Figure 18** presents fields of aggregates (snow) mixing ratios calculated at $t=144\text{h}$ by different microphysical schemes at an altitude of 7.5 km. One can see that the different structures of the snow maps reflect specific dynamical features of the above-discussed simulated circulations. For instance, the *SBM_COMP* snow field shows that the radius of the eyewall is larger than in *SBM_MAR*. Note that in *SBM_MAR*, the magnitude of snow mixing ratios is larger than in the *SBM_COMP*. This can be attributed to the lower supercooled CWC in *SBM_MAR* at higher levels that decreases the rate of riming and conversion of snow (i.e. aggregates) to graupel. Since in the bulk schemes the differences in the amount of supercooled CWC in polluted and clean cases are small, the snow mass amounts in the bulk

schemes are not very sensitive to aerosols. Among the bulk schemes, only *NSSL* shows a clear tendency to increase in snow mass content at the upper levels in *NSSL-MAR*. At the same time, the maximum values of snow mass content are twice as large as compared to *SBM*.

The structure of the snow shield reflects the structure of the vertical velocities of the TC inner core. One can see that while in *SBM_COMP*, the internal eyewall disappeared toward 144 h, the *THO_A* schemes indicates only partial replacement, and the Morrison scheme shows the existence of a double eyewall structure. As discussed above, these differences in the TC structure determine the differences in the maximum wind speed. Remarkably, in all schemes producing similar TC intensities, the structures and maximum values of snow mixing ratios were of the same order of magnitude. At the same time, in the schemes producing TC intensities very different from those in *SBM_COMP*, the snow content was much lower in *WDM6* and *WSM6*, and was negligible in *NSSL-N*. This can be partially attributed to significant supercooled water content in the *NSSL-N* scheme (Fig. 17), which leads to intense riming and intense graupel production.

6. Conclusions

In the present study, Hurricane Irene (2011) was simulated with WRF using 1-km grid-spacing with ocean coupling with different microphysical schemes. The goals of the simulations were to better understand the role of aerosols in the evolution of Irene, and thus possibly explain the 40-h time shift between the time instances of maximum wind and minimum pressure during Irene's evolution, as well as to estimate the sensitivity of the hurricane intensity to model microphysics.

Simulations show that the single *SBM* scheme with a realistic spatial distribution of aerosols able to simulate the observed time shift was the one in which the effects of continental aerosols penetrating the inner core of Irene were taken into account. It was shown that aerosols from the SAL that penetrated the eyewall when Irene crossed the band of Saharan aerosol first

tended to intensify Irene's development, when compared to the simulation with only maritime aerosols. Later, aerosols intensified the rainbands at the TC periphery and fostered formation of the secondary eyewall further from the TC center. As a result, the size of Irene's simulated inner core increased, leading to a decrease in both the radial pressure gradient and the maximum wind speed. Thus it was the case that the magnitude of the minimum pressure which continued to decrease for another 40 h—and maximum wind speed changed in opposite directions. Hence, this study demonstrates that aerosols not only decrease the intensity of tropical storms, as was found in previous studies, but also affect TC size and dynamical structure. *The simulations using the SBM show that effects of aerosols on convection should be considered in cloud-resolving TC forecast models.*

Thompson and Eidhammer's (2014) bulk-parameterization scheme, which takes into account the spatial variability of aerosol concentration and the interrelations between the parameters of gamma distributions used for the approximation of particle size distributions in clouds, yielded results most similar to those found in the observations and in the SBM. At the same time, it was shown that the sensitivity to aerosols of all bulk-parameterization schemes, including this new one, is relatively low. This degree of sensitivity did not allow the bulk-parameterization schemes to simulate the observed evolution of Irene's intensity and structure. Analysis of the microphysical structure of clouds produced by different bulk-parameterization schemes reveals that most of them simulate structures more typical of continental conditions with high values of CWC up until the level of homogeneous freezing. The bulk-parameterization schemes were not able to simulate the structures typical of maritime convective clouds with rapid formation of warm rain and the disappearance of supercooled CWC at the altitudes of 5-6 km.

We suggest several physical causes for the insensitivity of the bulk schemes tested to explicit or implicit changes in aerosol concentrations. The application of saturation adjustment makes the schemes insensible to aerosols at the stage of diffusional growth, as was discussed in

Section 3. Errors in size sorting during sedimentation dramatically decrease the sensitivity of the bulk schemes to aerosols at the precipitating stages of cloud evolution. The application of different autoconversion schemes cannot increase the sensitivity to aerosols, especially in cases when one-moment bulk schemes (which are unable to simulate size sorting) are used (Planche et al., 2014, Grabowski , 2014). It is remarkable that the 2-moment NSSL-2MOMCNN (Mansell et al., 2010; Mansell and Ziegler, 2013) scheme, which does not use the saturation adjustment assumption, revealed a high sensitivity to aerosols. In this scheme, the changes in aerosol conditions led to changes in minimum pressure and maximum winds by 15 mb and 15 kts, respectively. These results show that avoiding the saturation adjustment procedure is extremely important for improvement of bulk-parameterization schemes. This topic is elaborated in detail by Khain et al. (2015).

Another possible reason for the "continentality" and the insensitivity of bulk schemes to aerosols is that in some schemes (e.g., the Morrison scheme, and possibly some others) the process of autoconversion or collection does not reduce the droplet concentration. So, the decrease in CWC reduces in this case to a decrease in droplet size; that is, to the appearance of "continental" properties.

The microphysical fields produced by those bulk schemes that predict unrealistic TC intensities differed dramatically from the fields produced by microphysical schemes producing realistic TC intensities. On a positive note, when simulated bulk microphysical fields were similar to those of SBM, they produced minimum pressure and maximum winds that were similar to those produced by SBM.

The differences in the minimum pressure of 40 mb obtained by using different bulk schemes indicate the extremely important role of microphysics on the evolution of tropical storms as well as for accurate TC prediction. The differences in the vertical profiles of supercooled CWC clearly imply differences in the vertical profiles of latent heat release in different microphysical

schemes. The profiles of convective heating produced by different schemes seem to differ substantially from the profile that can be obtained using a simplified assumption about the moist adiabatic process (see saturation adjustment, Section 3) in the TC eyewall clouds. More accurate simulations of TC intensity using novel bulk schemes represent substantial progress in the parameterization of microphysical processes. However, further steps should be taken to sensitize the schemes to droplet concentration and aerosols. We believe that comparisons of microphysical fields obtained by bulk-parameterization schemes with those obtained by SBM will foster further progress in bulk-parameterization modeling.

In this study we used some averaged characteristics of continental and maritime aerosols. In reality aerosols are of different chemical composition with different nucleation activity. More detailed studies of aerosol effects on TC intensity that take in to account the aerosol types are of interest. The attempt of such analysis was performed, for instance, by Rosenfeld et al. (2011). However, it is a difficult problem because of many uncertainties.

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Name	Reference	Description of the scheme
Kessler	Kessler, 1969	One-moment, warm rain bulk parameterization, Hydrometeor types: Water droplets and raindrops.
Lin	Lin et al. (1983)	One moment. Hydrometeor classes: cloud droplets, rain drops, ice crystals and hail treated as a high density hydrometeor.
WSM6	Hong and Lim (2006)	One-moment, WRF Single-Moment 6-class scheme. Hydrometeor classes: droplets, raindrops, ice, snow, graupel. A gradual melting of snow is allowed. This scheme is efficient at grid spacing between the mesoscale and cloud-resolving grid
Goddard	Goddard microphysics scheme. Tao et al. (1989)	One-moment, based on Lin et al. (1983) and Rutledge and Hobbs (1984) schemes. A new procedure of ice-water saturation adjustment is included. Hydrometeor classes: droplets, raindrops, ice, snow and graupel
Thompson	Thompson et al., 2004 (as it is in WRF)	One-moment for all hydrometeors, except ice. Ice is described using a two moment scheme. Hydrometeor classes: droplets, raindrops, ice, snow and graupel. Empirical relation between parameters of gamma distributions, and between the parameters and environmental conditions.. A scheme is build for high-resolution simulations.
Thompson	Thompson et al., 2008, (as it is in WRF)	As Thompson et al. (2004), but two-moment approach is used for cloud ice and rain. It includes rain and ice number concentrations.
Tho-Aerosol_Aware	Thompson and Eidhammer, 2014 (as it is in WRF)	Two-moment scheme for cloud water, rain, ice and one-moment for snow and graupel. Prediction of the aerosol number concentration.
Morrison	Morrison et al., 2005, 2009	Double-moment, hydrometeor classes: droplets, rain drops, ice, snow, and graupel. Improved calculation of supersaturation is included. Concentration of CCN is prescribed.
SBU-YLin	Lin and Colle, 2011 Stony Brook University Y. Lin scheme	One-moment, a 5-class scheme. Hydrometeor classes: Droplets, raindrops, ice crystals, precipitating ice. The projected area, mass, and fall velocity of precipitating ice depends on riming intensity and temperature.
WDM6	Lim and Hong (2010)	Double-moment for droplets and rain, one moment for other hydrometeors. A prognostic treatment of cloud condensation nuclei (CCN) is introduced. Hyrdrometeor classes as in WSM6.
NSSL-2MOMCCN	Mansell et al., 2010; Mansell and Ziegler, 2013	The name in WRF: NSSL-2MOMCCN, double-moment, hydrometeor classes: droplets, raindrops, ice, snow, graupel, hail. Graupel density is recalculated. Conversion of graupel to hail takes place when wet growth begins. No saturation adjustment is assumed. Aerosols are a bulk CCN count, and are advected like any scalar number concentration. The CCN concentration is reduced by the amount of nucleated droplets. Activation stops when the CCN count goes to zero, so CCN are assumed to be activated at fairly low supersaturations.
CAM	Morrison and Gettelman (2008)	Double moment, used in the NCAR Community Atmosphere Model. Developed for GCM. Explicitly predicts mass and number mixing ratios of cloud droplets and ice crystals. Precipitation is calculated diagnostically in neglecting horizontal advection of precipitating particles

897 **Table 2. The main features of representation of different microphysical processes in SBM**
898 **and bulk schemes**

Microphysical process	SBM	Bulk-parameterization schemes
Droplet nucleation	Calculation of size distribution of nucleated droplets using size distribution of CCN and supersaturation; droplet nucleation depends on supersaturation history and is allowed both at cloud base and within clouds. Droplet nucleation affects size distribution of aerosols via nucleation scavenging.	Utilization of CCN activity spectrum to calculate droplet concentration at cloud base. Droplet nucleation typically does not affect the CCN activity spectrum.
Diffusional droplet growth	DSD is calculated by solving the equation for diffusional growth.	Saturation adjustment is typically used to calculate growth of the liquid water mixing ratio. In advanced schemes, the increase in the liquid water mixing ratio and in droplet concentration allows for an evaluation of the mean drop radius
Drop collisions and raindrop production	DSD of drops of all sizes is calculated by solving the stochastic collection equation. Size distribution of raindrops is calculated automatically as the distribution of drops with sizes exceeding a particular threshold.	The mixing ratio of raindrops is calculated using expressions for the rates of autoconversion. The expressions are either empirical or derived from solutions of the stochastic collection equation. The shape of the raindrop size distribution is prescribed.
Collision in mixed-phase and ice clouds	PSDs are calculated by solving stochastic collection equations, where all possible collisions between cloud particles are calculated.	Mixing ratios (and concentrations) of hydrometeors of different classes are calculated using different parameterization expressions for the conversion of hydrometeors of one class to another.
Sedimentation	Sedimentation is calculated for particles belonging to different bins having different fall velocities. Fall velocities depend on particle mass and hydrometeor class. They also depend on the rate of riming, liquid water fraction, and other factors.	Sedimentation is calculated with fall velocity averaged over all particle sizes. Mass-averaged fall velocity is larger than concentration-averaged fall velocity, so the mass and the concentration sediment with different velocities.
Freezing and melting	The rates of freezing and melting depend on particle size. Thus, freezing and melting lead to change in PSDs.	The average rates of freezing and melting are calculated by averaging over particles belonging to the same hydrometeor class.

Figures:

Figure 1. Combined MODIS Aqua and Terra aerosol optical depth at 550 nm 28 August 2011. The location of Irene’s center is indicated by the red TC symbol (D. Rosenfeld, personal communication, 2014).

Figure 2. Spatial initial configuration of aerosols used in the SBM and the *Tho_Aerosols_Aware*. The left panel shows the configuration used in SBM_MAR and in THO_A-Mar. The right panel shows the spatial distribution of aerosols in SBM_COMP, THO_A-Int, and THO_A-Comp. The asterisk shows the location of Irene. Initially, Irene was put with the area of clean air shown by the small square. This was done to take into account that Irene brought in clean air when she penetrated the SAL band.

Figure 3. Time dependencies of minimum pressure and maximum wind speed in simulations SBM_MAR and SBM_COMP. Observed time dependencies are also shown. The arrow shows a decrease in the maximum wind caused by aerosols.

Figure 4. Fields of RWC (in $g\ kg^{-1}$) near the oceanic surface at 120, 138 and 144 h in SBM_COMP (upper row) and SBM_MAR (lower row).

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963 Figure 5. Fields of droplet concentration (cm^{-3}) at z=800 m (upper row) and aerosol
 964 concentration at z=600 m in SBM_COMP at t=120, 138 and 144 h.

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Figure 6. Time dependence of the inner radius of eyewall in several simulations with SBM, and schemes by Thompson and Morrison. The inner radius is determined by the wind of 64 kt.

Figure 7. Radial dependencies of azimuthally averaged maximum wind and minimum surface pressure at different time instances as calculated with SBM_MAR (blue) and SBM_COMP (red).

Figure 8. Horizontal cross-sections in the field of vertical velocity at $z=9$ km in MAR (left) and SBM_COMP simulations. The fields are shown for the time instance of 60 h. One can see aerosol -induced intensification of convection in the eyewall in SBM_COMP .

Figure 9. Time dependencies of the minimum pressure (left) and maximum wind velocity (middle and right) in simulations carried out with schemes by Morrison, Thompson and NSSL allowing for changes of aerosol or droplet concentrations. The results obtained using the SBM and the simplest warm rain Kessler parameterization, as well as observed dependencies are presented for comparison. A description of the schemes is presented in Table 1.

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1028 Fig. 10. The same as in Fig. 4, but for *Tho_A-COMP* (upper row) and *Tho_A-MAR*(bottom)

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1041 Fig. 11. The same as in Fig. 7, but for THO_A-COMP and THO_A-MAR.

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Figure 12. The same as in figs. 4 and 10 but for simulations *NSSL-COMP_NBC* (panels a,b), *NSSL-COMP* (panels c and d) and *NSSL-MAR* (panels e and f). Rain water content is in g kg^{-1} .

Figure 13. The same as in Figures 7 and 11, but for NSSL-COMP, NSSL-MAR and NSSL-COMP-NBC.

Figure 14. Time dependences of minimum surface pressure (left) and maximum wind velocity (right) in simulations with different bulk-parameterization schemes. The observed dependencies are also presented (black curves). Description of schemes is presented in Table 1.

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1086 Figure 15. The same as in Fig. 6, but for different microphysical schemes.

Figure 16. Vertical east-west cross-sections in the fields of cloud water mixing ratio (in $g\ kg^{-1}$) in simulations with microphysical schemes sensitive to aerosols. Panels in the left column correspond to simulations with higher CCN concentrations. The cross-sections in the right column show results of simulations with low (maritime) aerosol concentrations. Color scales in different panels are different.

Figure 17. The same as in Fig. 16, but for the WSD6, WDM6 and NSSL-N bulk-parameterization schemes. The cloud water mixing ratio is in $g\ kg^{-1}$. Color scales in different panels are different.

Fig. 18. Snow mixing ratio at altitude of 7.5 km in simulations with different microphysical schemes. Color scales in different panels are different.