Investigation of the Effects of Anthropogenic Pollution on Typhoon Precipitation and Microphysical Processes Using WRF-Chem

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ABSTRACT

Taking Typhoon Usagi (2013) as an example, this study used the Weather Research and Forecasting Model with Chemistry to investigate the influence of anthropogenic aerosols on typhoons. Three simulations (CTL, CLEAN, EXTREME) were designed according to the emission intensity of the anthropogenic pollution. The results showed that although anthropogenic pollution did not demonstrate clear influence on the track and strength of the typhoon, it clearly changed the precipitation, distribution of water hydrometeors, and microphysical processes. In the CLEAN experiment, the precipitation rate declined because cloud water collected by the rain decreased. Similarly, the precipitation rate decreased in the EXTREME experiment, because the autoconversion of cloud water to rain was restrained. Regarding precipitation type, the rate of stratiform precipitation in both the CLEAN and the EXTREME simulations was suppressed because the ice-phase microphysical processes weakened. Compared with the CTL run, the rate of stratiform precipitation at the periphery of the typhoon was reduced by about 28% in both the CLEAN and the EXTREME simulations. Moreover, the rate of convective precipitation within 140–160 km of the center of the typhoon in the EXTREME experiment was about 33% greater than in the CTL simulation. This increase was triggered by new convection at the periphery in the EXTREME simulation related to cloud water reevaporation. Finally, compared with the CTL experiment, the peaks of both convective and mixed precipitation in the CLEAN and EXTREME experiments shifted 10 km toward the typhoon periphery.

1. Introduction

The annual discharge of anthropogenic aerosols into the atmosphere is considerable, but the effects of those aerosols on weather and climate remain very uncertain (IPCC 2007). Aerosols can absorb and reflect solar radiation, thereby reducing the surface temperature and planetary boundary layer height, but they also act as cloud condensation nuclei (CCN) or ice nuclei, affecting cloud microphysics and subsequent precipitation rates, and increasing cloud coverage, albedo, and lifetime (e.g., Twomey 1977; Rosenfeld et al. 2008). Therefore, the effects of anthropogenic aerosols on typhoons should not be neglected.

The following studies have shown that aerosols might affect cloud microphysical processes, latent heat release,
and the thermodynamic structure and precipitation of typhoons. An increase of CCN concentration can lead to an increase in cloud droplet number concentration (e.g., Zhang et al. 2007; Li et al. 2008). This would increase the cloud water content of a typhoon and decrease the cloud water collision efficiency, thereby inhibiting the formation of warm-cloud precipitation (e.g., Khain et al. 2005; Rosenfeld et al. 2007; Lin et al. 2011). The suppression of precipitation formation by the increase in cloud water content can lead to greater transportation of water to the freezing level, thereby triggering new convection at the typhoon’s periphery (Rosenfeld and Woodley 2003; Rosenfeld et al. 2007). Observational research has shown that increased aerosol loading is associated with taller cloud towers and anvils (Koren et al. 2010). Wang et al. (2014) highlighted that the radiative effect of light-absorbing aerosols causes warming in the lower troposphere, which strengthens lower-level convection and enhances precipitation in the rainband region. Conversely, light-absorbing aerosols in higher atmospheric layers increase stability, thereby diminishing convection, moistening the surface layer, and reducing evaporation and hence precipitation (Ban-Weiss et al. 2012). Furthermore, giant CCN (e.g., sea salt) might enhance warm-cloud microphysical processes and, thus, they could have various effects on typhoons (e.g., Johnson 1982; Feingold et al. 1998; Rosenfeld et al. 2012). Cotton et al. (2007) simulated the effects of aerosols on typhoons using the Regional Atmospheric Modeling System. Their simulation was conducted based on three concentrations of CCN: 100, 1000, and 2000 cm$^{-3}$. The results revealed that a polluted atmosphere weakens typhoons in their mature phase. Xu et al. (2013) used the Fifth-generation Pennsylvania State University–National Center for Atmospheric Research non-hydrostatic mesoscale model with various CCN concentrations to investigate the effects of aerosols on Typhoon Chanchu. Their results indicated that aerosols might have slight effects on typhoon track and intensity and that the response of convective precipitation to varying CCN concentrations was more sensitive than that of stratiform and mixed precipitation. Cotton et al. (2012) simulated Typhoon Nuri and found that during the early stage of pollution ingestion, the tropical storm intensified before weakening later. Herbener et al. (2014) highlighted that enhanced concentrations of aerosols in the eyewall region lead to storm intensification. Carrio and Cotton (2011) indicated that cloud-nucleating aerosols affecting only the outer rainbands would weaken a storm. Hazra et al. (2013) investigated the effects of aerosols using a convection-permitting model with a bulk scheme. They found that an increase in CCN concentration leads to a weakening of the storm but that it has little impact on track. Evan et al. (2011a) found that anthropogenic aerosols loading reduced vertical wind shear and, thus, enhanced convection. They (Evan et al. 2011b) also discovered that anthropogenic aerosols in the Arabian Sea reduce the basinwide vertical wind shear, creating an environment more favorable for tropical cyclone intensification. However, Saharan dust appears to suppress Atlantic tropical cyclone activity by introducing dry, stable air into the storms and enhancing local vertical wind shear (Dunion and Velden 2004). Rosenfeld et al. (2011) showed that CCN aerosols weaken storms by apparently slowing the conversion of cloud droplets into precipitation. Fan et al. (2012) examined the effects of aerosols on the microphysical properties and precipitation of deep convective clouds (associated with thunderstorms) and stratus cloud (associated with a cold front) over the Chinese mainland. They found that the bulk microphysical scheme was unable to simulate the invigoration effect related to CCN aerosols compared with the bin scheme. Furthermore, for a fixed CCN concentration, they found the bulk scheme generated much higher droplet numbers and smaller droplet sizes in a highly polluted environment. Moreover, the effects of aerosols on precipitation are likely nonmonotonic and vary under different meteorological and aerosol conditions (Li et al. 2008). The net effect of aerosols on precipitation depends on the environmental conditions (Khain 2009). Khain et al. (2008) also showed that aerosols could foster lighting formation at the periphery of a tropical cyclone and reduce convection intensity at its center.

Previous research investigating the effects of aerosols on typhoons has mainly defined various concentrations of CCN (e.g., Khain et al. 2005; Zhang et al. 2007; Cotton et al. 2012; Xu et al. 2013) or turned off warm rain formation (Rosenfeld et al. 2007). Generally, the CCN concentration was prescribed or else the simulations were conducted using an idealized tropical storm to examine the influence of aerosols. In reality, however, aerosols reflect and absorb solar radiation and serve as CCN. Moreover, in the atmosphere, aerosols experience chemical reactions, and they undergo nucleation and dry and wet scavenging. Accurate simulation of the complex feedback mechanisms between aerosols and the atmosphere requires fully coupled models, among which the Weather Research and Forecasting Model with Chemistry (WRF-Chem) is a representative online model (Grell et al. 2005) that can provide prognostic CCN size distributions, compositions, and number concentrations. Using WRF-Chem, we conducted three simulations to investigate the effects of anthropogenic aerosols on
the precipitation, thermodynamic structure, and microphysical processes of Typhoon Usagi (2013). This typhoon developed in the northwestern Pacific on 16 September 2013, and it dissipated on 24 September 2013. Its maximum intensity reached 140 knots (kt; 1 kt = 0.51 m s\(^{-1}\)) and it made landfall in the Pearl River Delta of China, which is a region where a considerable amount of anthropogenic pollution is emitted. Therefore, it is instructive to investigate the effects of anthropogenic aerosols on typhoons, particularly as they approach the Pearl River Delta.

2. Model description, configuration, and experimental design

a. Model description

WRF-Chem, which is designed for both operational and research applications, is a fully compressible, Euler nonhydrostatic model with a hydrostatic option available. Arakawa C-grid staggering is employed for the horizontal grid and terrain-following dry hydrostatic pressure is applied for the vertical coordinate. A second-or third-order Runge–Kutta time integration scheme is applied in the model, while smaller time steps are used for acoustic and gravity wave modes. At present, second-to sixth-order advection schemes are available for both vertical and horizontal directions. The available lateral boundary conditions include periodic, open symmetric, and specified options, while the bottom boundary conditions contain physical and free-slip options. Currently, cloud microphysics, cumulus parameterization, surface physics, planetary boundary layer physics, and atmospheric radiation physics are included in the model physics component. A variety of coupled physical and chemical processes such as transportation, deposition/emission, chemical transformation, aerosol interactions, photolysis, and radiation are considered in WRF-Chem (Grell et al. 2005). The model is an online model in which the chemical component is consistent with the meteorological component. The transport scheme of the chemical component is consistent with the meteorological component. Aerosol–cloud interaction is added into the model. Anthropogenic emissions (such as SO\(_2\), NO\(_x\), primary particles, and volatile organic compounds) are transformed into aerosols (secondary aerosols) through physical and chemical processes that include transportation, deposition, gas-phase chemistry, aqueous chemistry, and photolysis. Some of the aerosols can be transported into cloud and then activated as CCN. An activation parameterization of a multiple aerosols model (Abdul-Razzak and Ghan 2000) is used in WRF-Chem. Each aerosol particle consists of an internal mixture of ingredients, and each particle competes against the others for water. The particles start to grow at a critical radius when the surrounding supersaturation is greater than a threshold value for the aerosol particle. A broader description of the model can be found at http://www.wrf-model.org/index.php.

b. Model configuration

Here, the WRF-Chem, version 3.5.1, was used for the typhoon simulations. The experiments were initialized at 0000 UTC 20 September 2013 and integrated for 72 h; the first 12 h were treated as the spinup period. As shown in Fig. 1, three nested domains (D01, D02, and D03) were introduced to the model with horizontal resolutions of 36, 12, and 4 km, respectively. The Lin microphysics scheme (Lin et al. 1983; Ghan et al. 1997; Hong et al. 2004), a two-moment bulk scheme involving five types of hydrometeor (cloud water, rain, cloud ice, snow, and graupel), was used in the simulations. Figure 2 shows the microphysical processes and hydrometeors of the Lin scheme. The formation of warm rain is associated mainly with microphysical processes involving both the autoconversion of cloud water to form rain (RAUT) and the accretion of cloud water by rain (RACW). RAUT is the process of collision and coalescence by cloud droplets to form rain, whereas
RACW represents the process whereby rain continues to grow via the accretion of cloud water. At present, the Lin microphysics scheme is coupled with the chemistry in WRF-Chem (Liu et al. 2005).

The cloud droplet size can be calculated using the method of Lin et al. (1983):

$$\lambda_x = \left( \frac{\pi \rho_x n_{0x}}{\rho_l} \right)^{0.25},$$

where \(\rho_x\) is the density of the hydrometeor (water, snow, or graupel), \(n_{0x}\) is the intercept parameter for the corresponding hydrometeor, \(\rho_l\) is air density, \(l_x\) is the mixing ratios of the hydrometeor, and \(\lambda_x\) is the size distribution of the hydrometeor (Lin et al. 1983). Thus, when the mixing ratio of the hydrometeor is increased, the corresponding hydrometeor size will decrease.

In Lin microphysics scheme, cloud water and cloud ice are allowed to coexist between 0°C and -40°C. The conversion of cloud water to cloud ice within this temperature range (denoted the IDW process) depends on the deposition nucleation of natural ice nuclei and the depositional growth of cloud ice at the expense of cloud...
water (i.e., the Bergeron process). In addition, cloud water also could convert to snow via riming.

Rain is initiated via RAUT and, thus, this process is significant for precipitation. The parameterization of the autoconversion process used in Lin microphysics scheme was provided by Liu and Daum (2004):

$$ P_{\text{raut}} = \alpha N^{-1/3} L^{7/3} H(R-R_e) \quad \text{and} \quad \alpha = \left( \frac{3}{4 \pi d_p^2} \right)^2 \kappa \beta^{2/3} \left( \frac{L}{N} \right)^{2/3}, $$

where $P_{\text{raut}}$ is the autoconversion rate, $H$ is the Heaviside step function, $N$ is the total number concentration of cloud droplets, $R$ is the mean volume radius, $R_e$ is the threshold mean volume radius, $K_3 \approx 1.9 \times 10^{11}$, and $\beta$ depends on the relative dispersion. This parameterization is a Kessler-type autoconversion parameterization. Compared with previous Kessler-type autoconversion parameterizations, this improvement by Liu and Daum (2004) eliminates the assumptions of fixed collection efficiency and fixed terminal velocity. It indicates that the autoconversion rate is highly dependent on liquid water content, cloud water concentration, and relative dispersion. Cloud turbulence has an effect on the dispersion of the cloud droplet size distribution; thus, if the parameterization considered cloud turbulence, it should perform better. A detailed explanation can be found in Liu and Daum (2004).

The Grell–Devenyi ensemble cumulus parameterization scheme (Grell and Dévényi 2002) was used only in domains D01 and D02; no cumulus parameterization scheme was introduced in domain D03. Other major physics schemes involved included the Rapid Radiative Transfer Model for Global climate models (Mlawer et al. 1997; Iacono et al. 2000) shortwave–longwave radiation scheme, the Yonsei University planetary boundary layer scheme (Hong et al. 2006), and the National Centers for Environmental Prediction, Oregon State University, Air Force, and Hydrologic Research Laboratory’s land surface module (Chen and Dudhia 2001; Ek et al. 2003). In addition, a chemical reaction action, which could produce secondary aerosols in WRF-Chem, each mode width $\delta$ is fixed. Furthermore, aerosols are free to interact with radiation and clouds, and the number of aerosols decides the number and sizes of the cloud droplets. In the simulations, the size distributions and geometric widths of the three modes were assumed by MADE/SORGAM.

The initial and boundary conditions of the simulations were obtained from the $1^\circ \times 1^\circ$ National Centers for Environmental Prediction Final (NCEP-FNL) global tropospheric analysis dataset (http://rdap.ucar.edu/datasets/ds083.2). The $0.5^\circ \times 0.5^\circ$ Reanalysis of the Tropospheric chemical composition (http://retro-archive.iek.fz-juelich.de/data/emissions/) and $1^\circ \times 1^\circ$ Emission Database for Global Atmospheric Research (http://www.mnp.nl/edgar/introduction) datasets were used as the emission source of the anthropogenic aerosols. These emission datasets contain energy production and consumption, industrial manufacturing, and agricultural production. These anthropogenic air pollutions [including organic carbon (OC), black carbon (BC), primary PM, and pollution gases] experience diffusion (advection and convection), wet and dry deposition, and chemical reaction, which could produce secondary aerosols in WRF-Chem. Sea salt aerosols were obtained from the Goddard Chemistry Aerosol Radiation and Transport model. Bogus vortices were not used in the simulations.

c. Experimental design

Three experiments that altered only the emission intensity of the anthropogenic pollution were undertaken to study the effects of anthropogenic aerosols on typhoons. The control experiment (CTL) was conducted using the base-level emission intensity. The two sensitivity experiments comprised the clean atmosphere (CLEAN) experiment, which had one-tenth the emission intensity of the CTL, and an extremely polluted (EXTREME) experiment, which had 10 times the emission intensity of the CTL. Furthermore, an experiment with 3 times the emission intensity of the CTL was performed; however, its results revealed similar features to the EXTREME experiment and, therefore, in this paper, the analysis focuses only on the results of the CTL, CLEAN, and EXTREME simulations.
3. Results and analysis

Figure 3 shows the observed and simulated tracks of the typhoon at 3-h intervals. The simulated track, maximum wind speed, and minimum surface pressure data were from domain D02 at the early stages and then from domain D03. The observed typhoon track and intensity data were obtained from the Best Track Dataset for Tropical Cyclones in the Western North Pacific (http://tcdata.typhoon.gov.cn/zjljsjj_zlhq.html). The observed typhoon moved northwestward and made landfall in the Pearl River Delta region at 1240 UTC 22 September 2013. The three simulations produced tracks consistent with the observed track, but their landfall locations were all slightly farther south. This indicated that the anthropogenic aerosols had a weak effect on the typhoon track, as found in previous studies (e.g., Cotton et al. 2012; Hazra et al. 2013). Furthermore, the model was also shown capable of reproducing the main typhoon tendency (Fig. 4). The simulated maximum wind speeds and minimum surface pressures were both lower than observed. This suggested that the intensity of the simulated typhoon was weaker than observed because of the coarse resolution of the NCEP-FNL initial field. The initial and boundary conditions, and other parameters of model, could also affect the results of the simulations. Moreover, typhoon intensity was found to be insensitive to anthropogenic aerosols, consistent with the results of other research (e.g., Lin et al. 2011; Cotton et al. 2012; Hazra et al. 2013; Xu et al. 2013). However, aerosols have been found to have an obvious effect on storm intensity in some earlier studies. Some previous studies have prescribed CCN concentrations (e.g., Khain et al. 2005; Zhang et al. 2007; Cotton et al. 2012) or shut off warm rain formation (Rosenfeld et al. 2007) to represent aerosol effects. Therefore, they might have detected a more obvious effect on storm intensity. In this paper, aerosols could be activated as CCN in WRF-Chem.

Atmospheric aerosols might experience chemical reactions and undergo nucleation and dry and wet scavenging. Thus, only some of the aerosols might be entrained in a typhoon and activated as CCN. However, enhanced concentrations of aerosols in the outer rainband region can lead to storm weakening and, if in the eyewall region, lead to storm intensification (Herbener et al. 2014). As aerosols could be advected into both the eyewall and the outer rainband regions in WRF-Chem, the two effects would be competing with each other. Thus, this could be another reason why storm intensity was found insensitive to the concentrations of aerosols in this study.

Figure 5 shows the 24-h accumulated precipitation from the simulations and observations from 0000 UTC 22 September to 0000 UTC 23 September 2013. The observed precipitation was obtained from an automatic meteorological station in Guangdong Province. It can be seen that the spatial distributions and magnitudes of the simulated precipitation are consistent with the observations. Overall, the model reproduced the track, intensity, and distribution of the precipitation of Typhoon Usagi reasonably well.

During the early stages of the simulations, the tracks and intensities in the three experiments showed little variation because of low levels of anthropogenic aerosols. In this scenario, the anthropogenic aerosols were
emitted mainly from land, especially around the Pearl River Delta, following which they became entrained into the typhoon. Over the ocean, far from the continent, the anthropogenic aerosols subjected to wet and dry deposition affected the typhoon only marginally. However, as the typhoon approached the Pearl River Delta, greater quantities of anthropogenic aerosols were entrained into the typhoon and activated as CCN. This led to a corresponding increase in the quantity of cloud water and a decrease in the cloud droplet size, with less effective rainfall. To understand better the impact of anthropogenic aerosols on typhoons, we focused on the period from 1800 UTC September 21 to 0600 UTC 22 September 2013. During this period, the typhoon was near the Pearl River Delta and, thus, was affected by a considerable amount of anthropogenic aerosols, but not having made landfall, its structure was undisturbed.

Although there was little evidence that the track and intensity of the typhoon were affected by the anthropogenic aerosols, the distributions of the precipitation types and the microphysical processes could have been altered. First, precipitation was separated into convective, mixed, and stratiform types, based on the method of Sui et al. (2007). In this method, vertically integrated mixing ratios of cloud hydrometeors (cloud water, cloud ice, rain, snow, and graupel) were calculated at the beginning. Then, the cloud ratio, defined as the ice water path divided by the liquid water path (IWP/LWP), was used to evaluate the relative importance of the ice and water hydrometeors. Here, IWP was taken as the sum of the vertically integrated mixing ratios of ice hydrometeors and LWP was calculated as the sum of the vertically integrated mixing ratios of water hydrometeors. For a cloud ratio of less than 0.2 or a value of IWP > 2.55 mm, the precipitation was designated convective. Precipitation was treated as stratiform when the corresponding cloud ratio was greater than 1 and the value of IWP was less than 2.55 mm; the remainder was regarded as mixed precipitation. However, the mixed type contained precipitation that was more convective than stratiform.

The spiral distributions of precipitation for each simulation are shown in Fig. 6. The region around the eyewall constituted mainly convective precipitation, while most of the stratiform precipitation was located at the periphery of the typhoon. We noticed that the grid

![Fig. 5. The 24-h accumulated precipitation (mm) on land from Typhoon Usagi comparing observation and the three simulations. The time period is from 0000 UTC 22 Sep to 0000 UTC 23 Sep 2013.](unauthenticated)
number percentages of stratiform-type precipitation in the EXTREME and CLEAN experiments were markedly smaller than in the CTL experiment. However, the grid number percentage of convective-type precipitation in the EXTREME simulation exceeded that in the CTL (Table 1).

Anthropogenic aerosols also changed the typhoon’s precipitation rate. The mean precipitation rates (Table 1) of the CTL, CLEAN, and EXTREME experiments were 10.01, 9.55, and 9.72 mm, respectively, showing that the mean precipitation rates of the EXTREME and CLEAN simulations were suppressed. The mean precipitation rate in the EXTREME (CLEAN) simulation was about 5% (3%) smaller than in the CTL simulation. These differences in the hydrometeor content, microphysical processes, and precipitation rates (Tables 1 and 2) between the three simulations might appear small; however, it must be noted they are domain and temporal averages.

The horizontal distributions and horizontal changes in the hydrometeors and microphysical processes due to the aerosol perturbations were inhomogeneous. In the EXTREME simulation, rain was reduced in the eyewall region and increased at the periphery of the typhoon. In the CLEAN simulation, rain was reduced throughout the typhoon except for within a distance of 100–120 km from the storm center. Therefore, the overall difference in the mean value of rain (shown in Table 1) among the three simulations was small. The variations of other hydrometeor contents and microphysical processes were the same as for rain.

To interpret the mechanism behind the redistribution of precipitation type, we examined the changes of cloud hydrometeors. Figure 7 shows that the CCN (at supersaturation of 0.02%) concentration of the EXTREME simulation, which was configured with 10 times the normal emission intensity, was only about 3 times the CCN concentration of the CTL simulation. WRF-Chem is a fully online coupled model, in which the feedback mechanism between the meteorology and chemistry fields was included and where only a fraction of the aerosols could be activated as CCN. Therefore, we designed the three anthropogenic emission intensities to obtain clearer results. Compared with the CTL simulation, the cloud droplet number mixing ratio was increased in the EXTREME simulation and decreased in the CLEAN simulation. Figure 8 shows the greater cloud water mixing ratio in the EXTREME experiment, resulting from the abundant CCN (Fig. 7c). This increased cloud water mixing ratio was associated with smaller cloud water radii and a greater cloud droplet number mixing ratio than in the CTL experiment. In the EXTREME (CLEAN) simulation, the production rate of RACW (Table 2) was 47.29 (33.53) and the production rate of RAUT was 55.44
Table 1. Domain average within a 210-km radius of typhoon and time average from 1800 UTC 21 Sep to 0600 UTC 22 Sep 2013 of important variables in the three simulations. [CCN] is the vertical integration of CCN (at supersaturation of 0.02%); $\rho_v$ is precipitation rate; stratiform, convective, and mixed are representative of the proportion of grids of each precipitation type to all precipitation grids. IWP, LWP, [qc], [qg], [qr], and [qs] are vertical integrations of ice water path, liquid water path, cloud water, cloud ice, rain, snow, and graupel, respectively.

<table>
<thead>
<tr>
<th></th>
<th>CTL</th>
<th>CLEAN</th>
<th>EXTREME</th>
</tr>
</thead>
<tbody>
<tr>
<td>[CCN] (10^6 cm^{-2})</td>
<td>2.86</td>
<td>2.07</td>
<td>7.89</td>
</tr>
<tr>
<td>$\rho_v$ (mm h^{-1})</td>
<td>10.01</td>
<td>9.55</td>
<td>9.72</td>
</tr>
<tr>
<td>Stratiform (%)</td>
<td>20.33</td>
<td>18.82</td>
<td>17.96</td>
</tr>
<tr>
<td>Convective (%)</td>
<td>46.94</td>
<td>47.04</td>
<td>47.67</td>
</tr>
<tr>
<td>Mixed (%)</td>
<td>29.69</td>
<td>29.57</td>
<td>31.02</td>
</tr>
<tr>
<td>IWP (mm)</td>
<td>0.99</td>
<td>0.95</td>
<td>0.96</td>
</tr>
<tr>
<td>LWP (mm)</td>
<td>2.65</td>
<td>2.48</td>
<td>2.70</td>
</tr>
<tr>
<td>[qc] (mm)</td>
<td>0.48</td>
<td>0.40</td>
<td>0.57</td>
</tr>
<tr>
<td>[qr] (mm)</td>
<td>0.12</td>
<td>0.10</td>
<td>0.11</td>
</tr>
<tr>
<td>[qs] (mm)</td>
<td>2.17</td>
<td>2.08</td>
<td>2.13</td>
</tr>
<tr>
<td>[qs] (mm)</td>
<td>0.17</td>
<td>0.16</td>
<td>0.17</td>
</tr>
<tr>
<td>[qs] (mm)</td>
<td>0.71</td>
<td>0.68</td>
<td>0.67</td>
</tr>
</tbody>
</table>

In the EXTREME experiment, RACW was about 20% larger, whereas RAUT was about 23% smaller than in the CTL simulation. In the CLEAN experiment, RACW was about 16% smaller, whereas RAUT was about 5% larger than in the CTL simulation. Thus, although the RACW intensified, which could enhance the formation of rain, the precipitation rate in the EXTREME simulation was reduced by the suppression of RAUT. The cloud water mixing ratio in the CLEAN experiment was smaller because of the lack of CCN. Consequently, less vapor mass condensed into cloud water, reducing the precipitation rate and weakening the RACW process. In other words, warm rain microphysical processes, which were associated mainly with RAUT and RACW, were suppressed in both the EXTREME and the CLEAN simulations. In addition, cloud ice, snow, rain, and graupel contents (Fig. 8) were generally reduced in the CLEAN experiment except at the range of 100–130 km. Note that the rain content in the EXTREME simulation increased significantly at the range of 140–160 km from the typhoon center.

In the EXTREME simulation, there was greater vapor condensation because of the increase in the CCN (Fig. 9c) at the periphery. Thus, the cloud water mixing ratio (Fig. 8c) increased in association with a greater cloud droplet number mixing ratio and smaller cloud water radii. This was conducive to the evaporation of cloud water. It blocked vapor transportation to the typhoon’s center, causing greater stagnation of vapor content at the periphery, consistent with Rosenfeld et al. (2007). This also meant a smaller amount of vapor mass was transported to the typhoon center in the EXTREME simulation. The vapor content change (Fig. 9c) in the EXTREME simulation was complex. At low levels, the vapor content was reduced at the eyewall and increased at the periphery. Thus, a smaller amount of vapor content was transported to the upper level from the lower- and midlevel peripheral regions via updrafts and the radial wind. Thus, a belt of increased vapor content occurred in the upper level.

We also noted that cloud ice (Fig. 8c) increased at the upper levels outside the eyewall in the EXTREME experiment. The formation process of cloud ice includes the deposition growth of cloud ice (IDEP), homogeneous freezing of cloud water to form cloud ice (IHOM), and the Bergeron process (IDW). There were many processes for the sources and sinks of each hydrometeor (Fig. 2). The most important formation processes of cloud ice were IHOM (figure not shown) and IDEP (Fig. 10). IHOM was enhanced at the eyewall region and IDEP was enhanced throughout the typhoon near 8-km altitude in the EXTREME simulation. Thus, cloud ice was increased at the eyewall but decreased at the periphery even though IDEP was enhanced. We could infer that the increased peripheral cloud ice due to IDEP was converted to snow or graupel. Thus, the snow and graupel mixing ratio was increased at the periphery (near the range of 140–160 km). Finally, the increased levels of snow and graupel might have converted to rain...
within the range of 140–160 km. Based on the method of Sui et al. (2007), the increased rate of precipitation at the range of 140–160 km was calculated as convective precipitation.

Anthropogenic aerosols also changed the thermodynamic structure of the typhoon. First, according to Fig. 11, the temperature decreased at low levels at the periphery and outside the eyewall in the EXTREME simulation relative to the CTL simulation. This decrease resulted from the additional cloud water associated with the smaller droplet size. This suppressed cloud water collisions and enhanced cloud water evaporation (figure not shown), which absorbed heat. However, we also noted a warmer area in the midlevel peripheral region because of enhanced release of latent heat of condensation. In contrast, despite the smaller release of latent heat of condensation in the CLEAN simulation, the temperature increased at lower and midlevels. Possible reasons for this were the decrease in cloud water...
evaporation and smaller transport of latent heat to the upper levels because of the weaker vertical velocity (Fig. 11e).

In the CLEAN experiment, vapor condensation (Fig. 9e) was suppressed throughout most of the region, especially at the eyewall, because of the lack of CCN. Many factors control the vapor content. In the CLEAN simulation, the low-level radial wind was reduced, which led to less entrainment of vapor into the typhoon system. However, vapor condensation was suppressed, especially in the eyewall region and, thus, less vapor content was converted to cloud water. Therefore, the overall vapor content in the eyewall region was

**Fig. 8.** (a)–(c) The azimuthally averaged and time-averaged cloud water mixing ratio (shaded) and cloud ice mixing ratio (contours; dashed lines represent negative values). (d)–(f) Azimuthally averaged and time-averaged rainwater mixing ratio (shaded) and sum of snow and graupel mixing ratio (contours; dashed lines represent negative values). (a),(d) Outputs of CTL experiment. (b),(e) Differences of CLEAN and CTL simulations. (c),(f) Differences of EXTREME and CTL simulations. The time period is from 1800 UTC 21 Sep to 0600 UTC 22 Sep 2013. The red lines in (a) and (d) are the 0° isotherm.
increased. Figure 8e shows that rain, cloud ice, snow, and graupel contents were all increased at a range of around 100–130 km in the CLEAN simulation. The content of cloud ice was increased slightly at around 100–130 km, whereas the snow and graupel contents had obvious increases at around this range. First, we noted that vapor condensation increased significantly at an altitude of 2–8 km at around the range of 100–130 km. However, cloud water decreased in the corresponding area. Thus, we could infer that the increased cloud water in this area was converted to snow and graupel, and that this would eventually be converted to rain. These processes were associated with the mixed precipitation rate. Figure 12 shows that the mixed precipitation rate at around 100–130 km in the CLEAN simulation was increased compared with the CTL simulation. Based on the method of Sui et al. (2007), when IWP was greater than 2.55 mm, the precipitation was treated as convective precipitation. The cloud ice, snow, and graupel mixing ratios in the CLEAN simulation increased at around 100–130 km. Thus, the corresponding precipitation could be treated as convective precipitation. Figure 12
also shows that the area of convective precipitation in the CLEAN simulation was larger than in the CTL simulation at around 100–130 km. However, at this range, the convective precipitation rate in the CLEAN simulation was about 15 mm h\(^{-1}\), whereas it was about 14 mm h\(^{-1}\) in the CTL simulation. The convective precipitation in the CLEAN simulation was about 7% greater than in the CTL simulation at around 100–130 km.

Figure 10 shows that the radial wind speed in the EXTREME and CLEAN runs was reduced at low levels and increased at upper levels. This could have impeded vapor mass entrainment into the typhoon system at low levels but could also have been conducive to vapor mass detrainment at upper levels, leading to reduced vapor content at each of these levels. Therefore, compared with the CTL experiment, the integrated cloud ice, cloud snow, and graupel (Table 1) and ice phase microphysical processes (Table 2) decreased in the EXTREME and CLEAN runs.

Changes of ice phase microphysical processes significantly affected the structure of each precipitation type.
The rate of stratiform precipitation in the CTL simulation was the highest of all three runs (Fig. 12). The rates of stratiform precipitation in the CTL, CLEAN, and EXTREME simulations at the periphery of the typhoon were about 1.4, 1.0, and 1.0 mm h$^{-1}$, respectively; that is, the rate of stratiform precipitation at the periphery in the CLEAN and EXTREME simulations was about 28% smaller than in the CTL simulation. Stratiform precipitation is associated with ice phase microphysical processes. There was less vapor content at the upper levels in the CLEAN and EXTREME experiments and, therefore, their stratiform precipitation decreased because of the suppression of integrated ice phase microphysical processes such as IDEP and the depositional growth of snow. In addition, the peaks of the convective and mixed precipitation rates in the CLEAN and EXTREME simulations shifted 10 km toward the typhoon periphery. This could be attributed to the lower
transport of vapor content toward the typhoon center in these two experiments. Finally, we noticed that in the EXTREME simulation, convective precipitation rates increased within the range of 140–160 km from the center of the typhoon. Based on the proposed rainfall separation method, when IWP was greater than 2.55 mm, the precipitation was categorized as convective. This situation was always associated with convection in the upper levels. In the EXTREME experiment, the additional amount of cloud water was reevaporated, which then enhanced the formation of cloud ice by IDEP (Fig. 10f) at the periphery. This increase in cloud ice was consumed by snow and, ultimately, converted to rain. Therefore, the precipitation increased at the range of 140–160 km from the center of the typhoon and the additional precipitation was classified as convective. In other words, anthropogenic aerosols triggered new convection in the peripheral upper levels. In the EXTREME and CTL simulations, the rate of convective precipitation was about 6 and 4.5 mm h$^{-1}$, respectively, in the range of 140–160 km. The convective precipitation rate in the EXTREME simulation at this range was about 33% greater than the CTL simulation. In the CLEAN simulation, the mixed precipitation rate was slightly greater than the CTL run at around 100–130 km.

4. Summary

The fully coupled WRF-Chem Model was used to investigate the effects of anthropogenic aerosols on the precipitation type distribution, rainfall, and microphysical processes of Typhoon Usagi based on three experiments with differing intensities of anthropogenic emissions: low (CLEAN), normal (CTL), and high (EXTREME). The results showed that the track and intensity of the typhoon were nonsensitive to anthropogenic aerosol concentrations. However, both the CLEAN and the EXTREME experiments exhibited decreased rainfall rates. An increase of the cloud water mixing ratio associated with smaller cloud water radii occurred in the EXTREME simulation, which is detrimental to the collision of cloud water. In the CLEAN simulation, a smaller quantity of vapor was transformed to cloud water because of the suppression of vapor condensation due to the lack of cloud condensation nuclei (CCN). Compared with the CTL simulation, the autoconversion of cloud water to form rain (RAUT) decreased by 23% in the EXTREME experiment and the accretion of cloud water by rain (RACW) decreased by 16% in the CLEAN experiment. In the CLEAN simulation, the precipitation rate declined mainly because of the suppression of RACW. A decrease in

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**Fig. 12.** Azimuthally averaged and time-averaged precipitation rate by type. The time period is from 1800 UTC 21 Sep to 0600 UTC 22 Sep 2013. Solid line is CTL, dashed line is CLEAN, and dotted line is EXTREME.
rainfall rate in the EXTREME experiment was caused mainly by the inhibition of RAUT.

It was found that anthropogenic aerosols also modified the distribution of precipitation type in the typhoon. Initially, precipitation in the simulated typhoon was separated into stratiform, convective, and mixed components, based on the method of Sui et al. (2007). Both the CLEAN and the EXTREME experiments showed a decrease in the percentage of stratiform-type precipitation. The percentage of convective precipitation increased in the EXTREME experiment. We found that radial velocity decreased at lower levels and increased at upper levels. This caused reduced vapor mass ingestion into the typhoon at lower levels and greater vapor mass output at upper levels. Thus, the weakened ice microphysics of integrated deposition growth of cloud ice (IDEP) and depositional growth of snow in the CLEAN and EXTREME experiments reduced cloud ice and suppressed stratiform precipitation. Compared with the CTL run, the peripheral stratiform precipitation rate was decreased by about 28% in both the CLEAN and the EXTREME experiments. Less moisture transport to the typhoon center in these two experiments was used to explain the 10-km shift in their rainfall peaks toward the periphery. Finally, in the EXTREME simulation, the enhancement of vapor condensation produced additional small cloud water with low collision rates at the periphery, and a greater cloud water mixing ratio favored cloud water evaporation. This prevented vapor advection to the typhoon’s center, resulting in the stagnation of vapor content at the periphery. This cloud water from reevaporation triggered new convection at a range of 140–160 km from the typhoon center. Additionally, IDEP was enhanced in the EXTREME simulation, and the production of cloud ice by IDEP ultimately transformed into rain, which led to the decrease in cloud ice at the periphery of the typhoon. Compared with the CTL simulation, the rate of convective precipitation in the EXTREME simulation was 33% greater at the range of 140–160 km, and the rate of mixed precipitation was increased slightly in the CLEAN simulation at around 100–130 km.

The mechanism by which anthropogenic aerosols affect typhoons is complex and the effects vary with typhoon stage. This investigation only considered the influence of anthropogenic aerosols on a mature typhoon using WRF-Chem, and the aerosols were not regarded as ice nuclei in the Lin microphysics scheme. Furthermore, light-absorbing aerosols that cause warming in the lower troposphere, further strengthening lower-level convection and enhancing precipitation in the rainband region were not considered here (Wang et al. 2014). We focused solely on the variations of precipitation and microphysical processes caused by different concentrations of aerosols. Further simulations will be conducted to examine the radiative effects of aerosols on typhoons. Furthermore, different aerosol size distributions have different effects on typhoons. For example, sea spray (sea salt), which could act as giant CCN, are ingested into the cloud base and accrete cloud water efficiently—hence, greatly accelerating the conversion of cloud droplets into rain. Thus, sea spray might enhance the warm rain process (Feingold et al. 1998; Rosenfeld et al. 2012). Conversely, air pollution aerosols suppress the warm rain process. We focused on the effects of anthropogenic pollution on typhoon precipitation and microphysical processes in this study; however, the effects of sea salt, with particular reference to typhoon energy, will be examined in future studies using WRF-Chem.

WRF-Chem contains feedback mechanisms between the chemistry and meteorology and it predicts CCN concentrations. Using the WRF-Chem fully coupled model, this work provided detailed analyses and discussions on the changes in typhoon structure and microphysical processes due to process-level aerosol perturbations. Furthermore, we also provided an explanation of how aerosols alter and redistribute the convective and stratiform precipitation of a typhoon.

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