Localization and Invigoration of Mei-yu Front Rainfall due to Aerosol-Cloud Interactions: A Preliminary Assessment Based on WRF Simulations and IMFRE 2018 Field Observations

Lin Liu1, Chunguang Cui1, Yi Deng2, Zhimin Zhou1, Yang Hu1, Bin Wang1, Jing Ren1, Zhaoxin Cai1, Yongqing Bai1, Junmei Cai1, and Xiquan Dong2

Abstract Aerosol-cloud interactions remain a major source of uncertainty in our understanding and modeling of the Earth’s hydrological cycle. Based upon a diagnostic and modeling analysis utilizing the latest field measurements from the Integrative Monsoon Frontal Rainfall Experiment (IMFRE) 2018, this paper reports the effects of aerosols on the cloud properties along the Mei-yu front over the Middle Reaches of Yangtze River in China. Numerical experiments with the Weather Research and Forecasting (WRF) model suggest that increasing aerosol number concentration reduces surface precipitation by ~8.8% and delays the onset of rainfall by ~30 min. Furthermore, warm clouds are suppressed but the convective cores are slightly intensified. This corresponds to an overall aerosol effect of “localization and invigoration” of the Mei-yu rainfall and thus an elevated probability of short-term heavy rainfall. The signals of “convective invigoration” with a bulk scheme in this study are relatively weak compared to those simulated by bin microphysics. The increased aerosol concentration strengthens Mei-yu front and changes local morphology of the front, consistent with earlier studies demonstrating positive effects of convective heating on the genesis and maintenance of Mei-yu front via conditional instability of the second kind (CISK) and diabatic generation of potential vorticity. Also discussed are the uncertainties of bulk microphysics in simulating aerosol-cloud interactions, which may shed light on the design of future field campaigns to further understand the impact of aerosol-cloud interactions on weather and climate over China in boreal summer.

1. Introduction

Over the past decade, numerous studies have recognized the roles played by aerosols in changes of climate and weather systems (Bollasina et al., 2011; Li, Guo, et al., 2017; Li et al., 2019; Rosenfeld, 2006). The pathways by which aerosols affect the cloud and precipitation are twofold: aerosol-radiation interactions and aerosol-cloud interactions (ACIs) (Boucher et al., 2013). The buffering mechanisms in aerosol-cloud interactions make quantification of aerosol-induced effects on clouds and precipitation complex and elusive (Stevens & Feingold, 2009). Therefore, many uncertainties remain regarding ACIs, especially in mixed-phase convective clouds (Tao et al., 2012).

Both invigoration and suppression effects of aerosols on different cloud and precipitation types are demonstrated by many previous studies. The inhibition of warm cloud precipitation by aerosols has also been reported in different regions of the world (Ackerman et al., 2003; Wang et al., 2011). For a constant liquid water path, enhancement of aerosol concentration can increase the droplet concentration and decrease the mean size of cloud droplets (Twomey, 1977). The reduction of droplet sizes in polluted clouds inhibits collision and coalescence processes (Rosenfeld et al., 2001), thereby slows the conversion of cloud droplets into rainfall, to the extent of completely suppressing warm rain processes. At the same time, deep convective cloud invigoration by aerosols has been shown in observational (Andreae et al., 2004; Guo et al., 2018) and modeling (Fan et al., 2013; Khain et al., 2005; Lee et al., 2008) studies. The increased aerosol concentrations decrease cloud droplet radius, allowing more water to be carried above the 0°C isotherm, where freezing...
yields additional latent heat, thereby invigorating storms, and producing large ice hydrometeors and higher cloud top heights and occasionally more intense thunderstorms (Andreae et al., 2004). Fan et al. (2018) found that ultrafine aerosol particles that are ingested into deep convective clouds can be activated to form additional cloud droplets and release more latent heat, thus intensifying convective strength over the Amazon. Although previous studies have focused on deep convective clouds, these effects are expected to be significant in warm convective clouds as well (Dagan et al., 2017). Chen et al. (2017) focused on the warm-phase processes within a deep convective system and found that simulations with larger aerosol loading resulted in a larger total cloud mass, a larger cloud fraction in the upper levels, and a larger frequency of strong updrafts and rain rates. Although lots of contributions have been made in understanding the mechanisms of ACIs, the impacts of aerosol on precipitation still remain uncertain for different cloud regimes (Fan et al., 2012; Wang et al., 2013) and environment conditions. Further understanding of the aerosol indirect effects on mix-phase clouds may not only improve our perception of climate change but also improve numerical weather prediction (Jiang et al., 2017).

It has been a great challenge to elucidate the effects of aerosols on mixed-phase clouds accurately, largely due to the effects depending on atmospheric dynamic and thermodynamic conditions. Wind shear (Fan et al., 2009; Lebo & Morrison, 2014), atmospheric instability (Khain et al., 2008; Storer et al., 2010), and updraft strength (Guo et al., 2016) strongly influence ACIs. Many observational studies have shown that relative humidity is an important thermodynamic factor in the relationship between aerosols and precipitation, but it may have opposite effects in continental and marine areas (Khain et al., 2008; Rosenfeld et al., 2008). Li et al. (2011) showed that the precipitation frequency is found to increase with increasing concentration of condensation nuclei for clouds with high water contents but decreases for clouds with low water contents over U.S. Southern Great Plain. In addition, the aerosol composition and size distribution (Zhang et al., 2002), orography conditions (Fan et al., 2015; Nugent et al., 2016) can also influence aerosol indirect effects. Therefore, the relationships between aerosols and precipitation vary significantly on seasonal and spatial scales and depend on different precipitation processes.

The ever-growing population and the rapid industrial and agricultural development in China have led to a rapid and continuous increase in aerosol emissions (Li, Guo, et al., 2017; Liu et al., 2018), which has worsened the air quality in China. Li et al. (2007) considered that the frequent occurrence of heavy air pollution has resulted in middle-eastern China becoming a hot spot for studying the effects of aerosols on the climate system and hydrological cycle. Figure 1 shows the mean aerosol optical depth (AOD) from the Moderate Resolution Imaging Spectroradiometer (MODIS) satellite and daily rainfall during the Mei-yu season (15 June to 15 July) of 2010–2018 in middle-eastern China. The middle reaches of the Yangtze River (MRYR) are one of the most polluted areas during a typical Mei-yu season. In addition, the MRYR is highly impacted by East Asian Monsoon and has a typical humid climate and a high frequency of precipitation during Mei-yu season. This weather and climate background has led to the MRYR becoming a critical region for studying aerosol-cloud interactions, whereas most previous studies focused on the Beijing-Tianjin-Hebei (BTH), Yangtze River Delta (YRD), and Pearl River Delta (PRD) regions.

Mei-yu front cloud system primarily consists of convective and stratiform regions, where significantly different dynamics, thermodynamics and cloud microphysics are observed (Houze, 2014). In convective regions, the growth of ice is dominated by riming, whereas in stratiform regions, deposition and aggregation are the primary mechanisms (Churchill & Houze, 1984). Aerosol indirect effects on Mei-yu front cloud system, which is a complex and mixed-phase cloud system during summer, are worthy of further study. Due to the lack of extensive detailed observations for cloud microphysics, the current understanding of the aerosol-cloud-precipitation buffering system in Mei-yu front is still inadequate at the fundamental level. In this study, a field campaign entitled the “Investigative Monsoon Frontal Rainfall Experiment (IMFRE) 2018” was conducted from 10 June 2018 to 10 July 2018 in the MRYR to gather the meteorological and
aerosol data necessary both to initialize and validate (i.e., in situ cloud property observations and radar remote sensing) the models.

The main question raised here is: “How do aerosol-derived changes affect the properties of complex and mixed-phase clouds systems along the Mei-yu Front in boreal summer?” To address this question, we focus on the macro and microphysical changes of the different cloud regimes along the Mei-yu front in a modeling and diagnostic analysis that utilizes the Weather Research and Forecasting (WRF) model with the two moment aerosol-aware bulk microphysical scheme (Thompson & Eidhammer, 2014) and the latest field measurements from IMFRE 2018. The potential effects of aerosols on the properties of the Mei-yu front are also discussed. The remainder of this paper is organized as follows. Section 2 describes the data and methods. A description of the case study and the numerical experiments design are presented in section 3. The results and discussions are presented in section 4. A summary is provided in section 5.

2. Data and Methods

2.1. Datasets From the 2018 Field Campaign

Various data sets were collected for this study during the IMFRE 2018 field campaign, which was designed to collect comprehensive data to study the mechanisms of mesoscale convective systems (MCSs) during the Mei-yu season (Cui et al., 2015). The aerosol concentrations and MCS structures measured by IMFRE enable the impacts of aerosols on cloud microphysics and precipitation to be examined and model simulations to be evaluated. The red dot in Figure 2a represents the Xianning site used in the field campaign. Convective systems can easily be initiated and develop in this location due to the impact of the mesoscale topography, which is primarily influenced by the Mufu Mountains to the southeast of Xianning (Cui et al., 2015).

The primary instruments used in this study include the China New Generation Weather Radar (CINRAD) and C-band dual-polarization Doppler radar (C-POL). In the study area (Figure 2a), there are 15S-band operational Doppler radars that operate at 10-cm wavelength and provide nine unique elevation scans with different elevation angles every 6 min. The reflectivity data sets underwent critical quality control (Wu, Wan, et al., 2013) and were gridded to 3 km in the horizontal and 0.5 km in the vertical in an attempt to compare with model simulations. Table S1 in the supporting information shows the specifications of measurements from C-POL. The radar data were preprocessed by a group of algorithms (Wang et al., 2015) to provide hydrometeor categories. The algorithms include corrections of system biases of reflectivity and differential reflectivity, suppression of ground clutter and anomalous echoes, and filtering of a differential phase shift.

2.2. Precipitation Data and Reanalysis Products

A precipitation product (Shen et al., 2013), based on the rain-gauge precipitation from automatic weather stations in China and the Climate Prediction Center morphing technique (CMORPH) precipitation (Joyce et al., 2004), were used in this study. This product has 0.1° × 0.1° spatial resolution and hourly temporal resolution. The large-scale air flow and thermodynamic conditions of Mei-yu front are analyzed using the hourly ERA-5 climate reanalysis (Hersbach & Dee, 2016) in a grid of 0.25° × 0.25°, which is the fifth and latest generation of European reanalysis data produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). The ERA-5 reanalysis were also used as model initial and lateral boundary conditions.

2.3. Convective/Stratiform Regime Partitioning Method

A partitioning algorithm, based on Steiner et al. (1995), was used to separate the convective/stratiform regimes for each synthesized radar volume. The partitioning algorithm is based on the 3-km horizontal radar reflectivity gradient, which is a background exceeding technique using a low-level 2-D horizontal reflectivity field (Wu, Dong, et al., 2013). The detailed partitioning steps are as follows:

1. Any grid points having radar reflectivity factor ($Z_c$) greater than 40 dBZ are classified as convective (Huang et al., 2019).
2. $Z_c$ at each radar grid is compared to its background intensity, computed using the linearly averaged reflectivity with a 6-km radius centered on the grid. Any grid point that exceeds its background reflectivity by an intensity dependent value is classified as convective.
3. Each grid point classified as convective by either of the two tests, an intensity-dependent radius (Steiner et al., 1995) of influence is used to assign an area around the point as convective.
4. The rest of the echoes that exceed 10 dBZ are classified as stratiform; the remaining grid points that are not identified as either convective or stratiform grids are assigned as nonclassified grids, which are not included in the statistical range.

This algorithm is also suitable for the WRF-simulated radar reflectivity so that the model results are evaluated at the same metrics as the observations. WRF-simulated radar reflectivity is calculated assuming Rayleigh scattering following the approach of Smith (Smith, 1984). In addition to the convective/stratiform region, warm clouds and convective cores have been further analyzed in different sensitivity experiments. Here “warm clouds” refer to liquid clouds only with the cloud top height below the melting level (Fan et al., 2013). Similar to Luo et al. (2010), a grid column is defined as “convective core” if it satisfies the following criteria: A grid column is within the convective regime as defined above and the vertical velocity is greater than 5 m s\(^{-1}\). The cloud fraction was calculated based on the number of cloudy grid points divided
by the total number of grid points in the domain. Cloudy points were identified by the threshold total condensed water being greater than $10^{-6}$ kg·kg$^{-1}$.

2.4. Intensity and Location of the Mei-Yu Front

An important feature of the Mei-yu front is its strong humidity gradient across the front. Therefore, it is more suitable to describe the intensity of Mei-Yu fronts using the absolute value of the gradient of equivalent potential temperature ($\theta_e$) (Li, Deng, et al., 2017). The composite location of the Meiyu-front is the averaged location of the Meiyu-front within 25°–35°N and 105°–125°E, and the procedure defining the location of Meiyu-front is adopted from Li, Deng, et al. (2017): (1) We first check whether there exists a band at each longitude with $\left| \frac{\partial \theta_e}{\partial y} \right| > 0.04$ K/km at 850 hPa. (2) If it does exist, the center latitude of this band in a specific longitude is then calculated. (3) It can be defined as Meiyu-front when (a) the total grids number of the bands $\left| \frac{\partial \theta_e}{\partial y} \right| > 0.04$ K/km at all longitude exceeds 3,000; (b) the average difference of center latitude

![Figure 3](https://example.com/figure3.png)

**Figure 3.** Longitude-latitude distributions of surface accumulated precipitation between 1800 UTC on 29 June and 1,200 UTC on 30 June 2018 from (a) rain gauge observations and (b) WRF simulations using the control Thompson aerosol-aware scheme, and (c) the hourly surface precipitation rates averaged within the region (108°–117°E, 27°–35°N) from the observations (black solid curve) and the simulations.
between adjacent longitude is less than 1. (i.e., \( \frac{1}{N-1} \times \sum_{i=0}^{N-2} |\text{lat}(i + 1) - \text{lat}(i)| < 1 \)); (c) the standard deviation of the center latitude is lower than 2.0 and the departure of center latitudes from the averaged center latitudes is less than 1.0 (i.e., \( |\text{lat}(i) - \overline{\text{lat}}| < 1 \)).

3. Case Description and Experiments Design

3.1. Case Description

During the IMFRE 2018 field campaign, a Mei-yu front storm was initiated over the MRYR that lasted for about 19 hr from 1800 UTC on 29 June to 1200 UTC on 30 June. The MCSs occurred along the Mei-yu front, causing moderate precipitation in central Hubei Province, with a rain center that exceeded 100 mm adjacent to the Xianning site (Figure 3a). As shown in Figure 3c, rainfall on 30 June occurred mainly between 0000 and 1200 UTC, with the peak precipitation occurring between 0300 and 0600 UTC.

The large-scale meteorological conditions derived from ERA-5 reanalysis data at 0000 UTC on 30 June are presented in Figure S1. The air flow at 700-hPa was characterized by a northeast-southwest oriented shear line formed at 29–30°N. To the south of this shear line, there was a warm and moist southwesterly low-level jet (LLJ) extending from the South China Sea to the JiangHuai Basin. The large pseudo equivalent potential temperature gradient at 850 hPa, which is often referred to as the location of the Mei-yu front, is also located near 30°N. These large-scale weather conditions are typical features of a Mei-yu front rainstorm.

Figure S2 describes the lifecycle of the Mei-yu front rainstorm via the hourly radar reflectivity at 3 km from 0000 to 0800 UTC on 30 June 2018. During the developing stage (1800 UTC 29 June to 0200 UTC 30 June), two small convective cells along the northwest-southeast oriented shear line formed and strengthened when moving eastward. At the mature stage (0300–0500 UTC 30 June), the two cells merged into one convective system that was maintained between 112°E and 114°E. After 0600 UTC on 30 June, the convective cells weakened and gradually dissipated, while the stratiform region remained quite extensive until 0800 UTC on 30 June.

3.2. Model Setup and Experiments Design

To investigate the interactions between aerosols and cloud-precipitation in the Mei-yu front, a series of numerical experiments was performed using the WRF model Version 3.9. The simulation domain consisted of 560 × 530 grids, with a horizontal resolution of 3 km (Figure 2b). There were 51 layers from the surface to 50 hPa. All simulations started at 0000 UTC on 29 June and ran for 36 hr. The model outputs were saved at 6-min intervals. Input and lateral boundary condition were supplied by the ERA-5 reanalysis mentioned above every 1 hr. Spectral nudging was adopted in simulations with Wave Number 6 in both zonal and meridional directions applied to U, V, and T. With the spectral nudging, the WRF model is able to preserve the variability of large-scale weather conditions and therefore regulate the corresponding variabilities of temperature and precipitation (Glisan & Gutowski, 2014).

A two-moment aerosol-aware bulk microphysical scheme, developed and discussed by Thompson and Eidhammer (2014), is used in this study. And this is an update of Thompson et al. (2008) bulk microphysical scheme, which consists of a double-moment bulk microphysical parameterization that explicitly calculates droplet nucleation and ice activation due to aerosols. Besides cloud, rain, snow, graupel, and ice hydrometeor species, the scheme transports two aerosol species (hygroscopic and ice nucleating) adding only about 15% computational cost (Saide et al., 2016). The Köhler activation theory (Köhler, 1936) is used for...
nucleation of aerosols to form cloud droplets by using a lookup table procedure, and a generalized gamma distribution is assumed for cloud droplets. More descriptions can be found in Thompson and Eidhammer (2014). The other physics parameterization schemes used were the Yonsei University (YSU) planetary boundary layer (PBL) scheme (Hong et al., 2006), the Rapid Radiative Transfer Model for General Circulation Models (RRTMG) scheme for longwave radiation calculations (Iacono et al., 2008), the Goddard shortwave scheme for shortwave radiation calculations (Chou & Suarez, 1994), the Noah land surface scheme (Chen & Dudhia, 2001), and no convective parameterization, since the grid was sufficiently high resolution to simulate most clouds explicitly.

Three sensitivity experiments with different initial and boundary aerosol number concentrations were conducted to investigate the response of clouds and precipitation to changes of aerosol number concentrations during a Mei-yu frontal rainstorm. The input monthly-averaged aerosol datasets were obtained from the corresponding regional 7-yr (2001–2007) global simulations (Colarco et al., 2010) of the Goddard Chemistry Aerosol Radiation and Transport (GOCART) model with 0.5° longitude by 1.25° latitude resolution. The aerosol input data include mass mixing ratios of sulfate, dust, black carbon (BC), organic carbon (OC), and sea salt, although black carbon is ignored in the cloud condensation nuclei (CCN) activation. As noted by Thompson and Eidhammer (2014), dust aerosols greater than 0.5 μm are accumulated into the nonhygroscopic, ice-nucleating mode, while all other aerosol species are combined into the hygroscopic, cloud droplet-nucleating mode. The input mass mixing ratios were converted to number concentrations by assuming lognormal distributions and applying the approach of Chin et al. (2002) for characteristic diameters and geometric standard deviations. More details can be found in Thompson and Eidhammer (2014). In our experiments, the model using aerosol-aware scheme was first run with the input and lateral boundary condition data described above representing aerosol conditions that are representative of conditions present in the current era. This simulation will be referred to as “Control.” Two additional simulations were conducted, with the initial hygroscopic aerosol number concentration set to be one tenth and 10 times of that in the Control simulation, referred to as “Clean” and “Polluted,” respectively. Similar to the method of Thompson and Eidhammer (2014), the nonhygroscopic aerosol (dust) remains identical in the three experiments in order to minimize any changes due to ice nucleation in these tests.

### 3.3. Model Evaluation

Compared to the observations, the centers of areas with more than 75-mm precipitation of Control simulation are slightly further to the northwest with a larger scope, and the scope of the 15- to 50-mm rainfall centers is smaller than in the observations (Figure 3). Despite some minor location errors for the rainfall, Control simulation captured the distribution of northwest-southeast oriented rain bands. Figure 3c shows a time series of hourly domain-averaged accumulated precipitation from the observation and simulation. The simulated curves are similar to the observations but underestimate the total mean precipitation by 23–35%, especially for 0000 to 1200 UTC on 30 June. The discrepancies are mainly attributed to the estimation of 15–50 mm precipitation (Figures 3a and 3b).

To evaluate the horizontal and vertical distribution of simulated MCSs, 3-km radar reflectivity and cross sections at developing (0100 UTC), mature (0400 UTC), and dissipate (0700 UTC) stages are presented in Figures S3 and S4. The Control simulations capture the two smaller MCSs, which are merged into one like the radar observations during the mature stage, but the locations are slightly to the northwest compared with the observations. The vertical cross sections of the radar reflectivity show that the large reflectivity (greater than 40 dBZ) within the deep convective cores from the radar extends to about 11 km at the mature stage. The simulated heights of the deep convective cores are similar to the observations but underestimate the radar reflectivity in the stratiform region at heights of 9–13 km.

Moreover, a vertical cross section of hydrometeor categories following the method of Dolan et al. (2013) is derived from C-POL, as shown in Figure 4. The upper regions of the storm are characterized by dry snow and regions of ice crystals. These aggregates then transition to wet snow as they melt between 4 and 6 km, and then form rain below 4 km. Regions of graupel are concentrated at heights from 4 to 8 km, which then transition to rain drops/melting hail. For the vertical distribution of particles, the simulation is closest to the observations, although it simulates fewer ice particles, which indicates a certain confidence in the simulation results.
Figure 5. Profiles of time- and domain-averaged hydrometeor mixing ratios simulated by WRF under the Clean (green solid curve), Control (black solid curve), and Polluted (red solid curve) scenarios over the (a, c, e, g, and i) stratiform and (b, d, f, h, and j) convective regions during the period 0300–0500 UTC on 30 June 2018 within the study domain (108–117°E, 27–35°N). Note the different x axis scales for various species, especially big differences between stratiform and convective region.
4. Results and Discussion

4.1. Aerosol Effects on Cloud Microphysical Properties

Due to the significantly different microphysical and thermodynamic features in convective/stratiform regimes, the simulated clouds were divided into stratiform and convective regimes to analyze. Figure 5 shows the cloud microphysical response to aerosols over the two regimes. With the increased aerosol loading, the cloud water mixing ratios in both stratiform and convective regimes increase (Figures 5a and 5b). The rainwater mixing ratio shows a different trend in stratiform/convective regions from the Clean to Polluted scenario (Figures 5c and 5d). There is a monotonic increase for the rainwater mass in convective regime, while the rainwater mass in stratiform regime is less sensitive to the aerosol concentration, showing a weak decreasing trend. The same quantitative results can also be seen from Table 1, which shows the temporal and spatial mean mixing ratios of cloud, rain, and ice-phase particles under the three scenarios. In terms of number concentration, more cloud particles and less rain particles are present in the dirty environment than in the clean environment in both regimes (Figure S5). For a constant liquid water path, enhancement of aerosol concentration can increase the droplet concentration and decrease the mean size of cloud droplets (Twomey, 1977). The reduction of cloud droplet sizes in polluted conditions inhibits collision and coalescence processes (Rosenfeld et al., 2001), thereby slows the conversion of cloud droplets into rainfall. Below the freezing level, the rainwater mass is increased in the polluted convective regime due to the melting of increased ice-phase particles (Table 1), producing additional rainwater to compensate the loss due to the inhibited collision-coalescence process (Lin et al., 2016).

The ice-phase particles analyzed in this study include graupel, snow, and ice crystal. In the stratiform region, the graupel mixing ratio almost has no changes, whereas an obvious increase is found in the convective region (Figures 5e and 5f). The ice crystal mixing ratio decreases in the two regions (Figures 5g and 5h) and the snow mixing ratio shows a tendency to increase as the added aerosol concentrations in both regions (Figures 5i and 5j). Table 1 indicates that the sum of the three ice-phase particles does not change significantly with increased aerosol concentrations in the stratiform region, whereas the mixing ratio of ice-phase particles are obviously enhanced in the convective region. Figure 6 shows the comparison of the contoured frequency by altitude diagrams (CFADs) of the reflectivity (Yuter & Houze, 1995) measured by the radar versus that simulated by the three aerosol scenarios over different regimes. The discrepancies of CFADs between simulations and observations mainly occur at the upper levels (8–14 km) in both cloud regimes. This suggests that the bulk scheme is not capable of reasonably simulating the ice particles at upper level. The differences of the three scenarios are mainly found at altitudes of 2–7 km with an opposite trend over stratiform and convective regimes. For the stratiform region, the frequency of 15–30 dBZ slightly decreases with the increase of aerosol number concentration below 4 km. Over the convective regime, the reflectivity of the dirty scenario at altitudes of 4–7 km is more concentrated in the 30–45 dBZ region than in the Clean test. The simulated frequency of large reflectivity (i.e., 30–40 dBZ) is also higher than under the clean condition. This implies that the increased aerosol concentrations enhance the convective intensity at altitudes of 4–7 km, which is consistent with the increased ice-phase particles mass in dirty conditions. The microphysics and feedbacks among various hydrometeors is complex (Lim & Hong, 2010). Changes of aerosol loading could indirectly impact other microphysical processes related to ice-phase particles.

4.2. Aerosol Effects on Cloud Macrophysical Properties

In addition to the stratiform/convective regimes, warm clouds and convective core regions were further analyzed to study the cloud macrophysical response to aerosols. Figure 7 illustrates the simulated vertical profiles of time- and domain-averaged cloud fractions over different cloud regimes. The most obvious feature is that the total cloud fractions (Figure 7a) increase with the increased aerosol loading in the middle and upper levels (5–12 km). Further partitioning into the stratiform and convective regimes (Figures 7b and 7c) indicates that the increase of total cloud fraction in the middle and upper levels is due to the increase of clouds in stratiform regime, while the cloud fraction in convective regime decreases with the increase of aerosol concentrations. The underlying mechanism for the increase of stratiform fraction in our study has been...
illustrated by a few studies (Chen et al., 2020; Fan et al., 2013). Specifically, Fan et al. (2013) documented that the microphysical effects induced by aerosols are a fundamental reason for the increases in cloud fraction. The reduction of ice and snow particle size leads to the reduction of their terminal velocities in the stratiform/anvil regime, which greatly reduces the dissipation of stratiform/anvils. Therefore, the stratiform/anvil clouds remain longer, and the strong horizontal advection in the upper troposphere further spreads them out over larger areas.

Figure 6. Contoured frequency by altitude diagrams (CFADs) of radar reflectivity from (a, b) radar observations and WRF simulations at 6-min intervals under (c, d) Clean, (e, f) Control, and (g, h) Polluted scenarios over the stratiform (left column) and convective regions (right column) during the period 0300–0500 UTC on 30 June 2018 within the study domain (108°–117°E, 27°–35°N).
The cloud fraction of warm clouds (Figure 7d) decreases from the Clean to Polluted tests due to the reduction of cloud droplet sizes in polluted conditions inhibiting collision and coalescence processes (Rosenfeld et al., 2001) discussed in section 4.1. For the convective core, the cloud fraction (Figure 7e) increases slightly with the added aerosols at altitudes of 4–8 km. Above 8 km, the cloud fraction of convective core in Polluted scenario reduces compared with the Control scenario. Overall, the increased aerosol concentrations inhibit the formation of warm clouds and increase the formation of stratiform clouds in the middle and upper levels.

4.3. Aerosol Effects on Vertical Motions and Precipitation

Vertical air velocity (w) is one of the key processes driving convective clouds. The intensity, duration and characteristic size of the updrafts determine the convective clouds’ properties. In addition, w affects the distribution of water along the atmospheric column. To investigate how microphysical processes affect vertical
air motion at the grid scale, statistical distributions of updrafts and downdrafts are examined. An updraft or downdraft is defined here as any grid box where $w$ is greater than 1 or less than $−1\text{m s}^{-1}$. This definition is the same as in McFarquhar et al. (2006) and similar to definitions used by Houze (2014).

Figure 8 illustrates CFADs of vertical velocity over the updraft (left column), downdraft (middle column), and convective core (right column) regions from three scenarios. With the increased aerosol concentrations, the updraft speed slightly increases especially for the motions at the altitudes of 4–8 km and the downdraft speed has no obvious changes. The upward speed of convective core has an increase in the dirty conditions. Vertical velocity is related to buoyancy, which is determined in part by latent heating and condensate loading (Luo et al., 2010). With the increase of aerosol loading, the net latent heating of the convective core region increases (Figure S6), which is consistent with the increase of updraft in convective core.
Additional cloud droplets are generated with more latent heating release over convective core in dirty conditions and then the subsequent latent heat release from ice-phased processes, resulting in an increase of updraft in convective core.

Note that the increases of cloud fraction and updraft speed with the added aerosols in convective core are quite moderate, indicating that signals of “convective invigoration” in our analysis with a bulk scheme is relatively weak compared to those that could be detected with bin microphysics in other studies (Chen et al., 2020; Fan et al., 2012, 2013, 2015; Khain et al., 2009, 2015; Lebo & Seinfeld, 2011). This is a typical limitation of bulk microphysics parameterizations and the reasons will be discussed in section 5. Additionally, convective invigoration mainly occurs in the middle and upper levels mentioned above through freezing more cloud water, which is different from the “warm-phase invigoration” documented in Stolz et al. (2015), Fan et al. (2018), and Lebo (2018).

Figure 9 shows 6-min domain average and accumulative surface rainfall rate for three aerosol scenarios. Total rain amount suppression is evident in the Polluted scenario, with the accumulated precipitation reduced by ~8.8% compared to the Clean scenario (Table 2). Tao et al. (2007) simulated two inland deep convective systems, and suggested that precipitation was suppressed by about 24% and 12.7%, respectively, in a high-aerosol-concentration test compared with a low-aerosol-concentration test. Such quantitative differences may have been due to differences in the selection of precipitation events, models, study areas, and durations of the period studied. The zoom-in of rain rate during developing stage is also shown.

Figure 9. (a) The 6-min surface precipitation rate and (b) 6-min accumulative precipitation averaged within the study domain (108°–117°E, 27°–35°N) simulated by WRF under the Clean (green solid curve), Control (black solid curve), and Polluted (red solid curve) scenarios. The zoom-in of rain rate during developing stage is also shown.
delayed the onset of rainfall for about 30 min compared to Clean scenario, which was in good agreement with observations (Rosenfeld, 2000). This is due to increased aerosol concentrations, which reduce the radius of the cloud particles, restricting the collision and coalescence processes and thereby suppressing the warm clouds mentioned above.

Figures 10a and 10b shows the time evolution of rain area fraction from the three scenarios over stratiform and convective regions. The Polluted scenario produces a greater stratiform rain area fraction and smaller convective rain area fraction compared to the Clean scenario (Table 2). Figure 10c shows the rain rate differences between the Clean and Control runs and the differences between the Polluted and Control runs over two cloud regimes. It can be found that the rain rate of stratiform precipitation is weakened with small amplitude in the Polluted scenario. On the contrary, the convective rain rate is enhanced under polluted conditions.

### Table 2
The Time- and Domain-Averaged Accumulated Rainfall, Stratiform Rain Percentage, Convective Rain Percentage, Stratiform Rain Rate, and Convective Rain Rate Under Clean, Control, and Polluted Conditions During 1800 UTC on 29 June to 1200 UTC on 30 June 2018 Within the Region (108°–117°E, 27°–35°N)

<table>
<thead>
<tr>
<th></th>
<th>Clean</th>
<th>Control</th>
<th>Polluted</th>
</tr>
</thead>
<tbody>
<tr>
<td>Accumulated rainfall (mm)</td>
<td>6.60</td>
<td>6.26</td>
<td>6.03</td>
</tr>
<tr>
<td>Stratiform rain percentage (%)</td>
<td>21.63</td>
<td>21.83</td>
<td>22.47</td>
</tr>
<tr>
<td>Convective rain percentage (%)</td>
<td>6.84</td>
<td>6.21</td>
<td>5.53</td>
</tr>
<tr>
<td>Mean stratiform rain rate (mm/hr)</td>
<td>0.80</td>
<td>0.78</td>
<td>0.77</td>
</tr>
<tr>
<td>Mean convective rain rate (mm/hr)</td>
<td>4.98</td>
<td>5.44</td>
<td>5.59</td>
</tr>
</tbody>
</table>

Figure 10. Temporal evolution of the rain area fractions over the (a) stratiform and (b) convective regions under the Clean (green solid curve), Control (black solid curve), and Polluted (red solid curve). (c) Rain rate differences (units: mm hr⁻¹) between the two aerosol scenarios and the control over stratiform and convective regions within the study domain (108°–117°E, 27°–35°N).
4.4. Aerosol Effects on Mei-Yu Front

For the Mei-yu front, a synoptic-scale system, how do aerosols’ thermodynamic and dynamic effects influence the frontal structure and intensity? Figures 11a and 11b show the frontal intensity differences and locations of Mei-yu front at 850 hPa from the three scenarios. With the increased aerosol concentrations, the location of Mei-yu front changes without obvious trends and the frontal intensity slightly increases (values of mean intensity shown at the upper right corner of Figures 11a and 11b). The cross sections of frontal intensity along the black line in Figure 11a were shown in Figures 11c–11e. The height of the front extends from 1,000 to 500 hPa, with the maximum gradient located at 850–750 hPa. In Polluted conditions, the intensity maximum increases significantly and extends further downward in comparison to the clean and control scenario. These results indicate that increased aerosol loading affects local morphology of the Mei-yu front while leaving the main location of the front intact.
Figure 12 shows the temporal evolution of the frontal intensity at 850 hPa and the net latent heating over convective cores and stratiform regimes from the three scenarios. During the analysis period, the frontal intensity increases significantly during 0500–0800 UTC in the dirty conditions. The change in the net latent heating over the convective core regime is characterized by an increase during 0300–0600 UTC in the dirty conditions, consistent with the change in the frontal intensity. The change in the net latent heating over stratiform regime is opposite from that over the convective core. It is clear that the local intensification of the Mei-yu front was mainly attributable to the heating associated with deep convection cores. Figure 13 presents the distributions of low-level (850 hPa) potential vorticity (PV) at 0400 UTC on 30 June under three scenarios. The domain (black box shown in Figure S2e) contains the convective area at 0400 UTC where main condensation occurs. An apparent positive low-level PV anomaly is found in the Polluted scenario (the mean value of PV under three scenarios is 0.76, 0.84 and 0.88 PVU, respectively), which is consistent with higher intensity of the front. Therefore, the enhanced latent heating associated with the organized MCSs, through low-level potential vorticity generation, appears to be the primary mechanism for strengthening the Mei-yu front in dirty conditions. This is consistent with arguments made in earlier studies (e.g., Chen et al., 2006; Cho & Chen, 1994) that conditional instability of the second kind (CISK) is important for the Mei-yu frontogenesis and maintaining, and latent heating from convection along the front feeds back positively to the weak-baroclinicity of a Mei-yu front.

5. Conclusions and Discussions

Aerosol-cloud interaction is one of the uncertain aspects in simulating climate and hydrological cycle due to the buffering mechanisms in aerosol-cloud meteorology. The impacts of aerosol on cloud-precipitation...
system highly depend on different aerosol composition, cloud types and atmospheric environment conditions. The main question raised here is: “How do aerosol-derived changes in the complex cloud systems along Mei-yu Front in China?” To explore this question, we used a WRF model with a two-moment aerosol-aware bulk microphysical scheme to simulate a Mei-yu front rainfall over MRYR during IMFRE 2018. The model results were validated by ground-based observations, such as similarities in the surface rain rate, the vertical structure of radar reflectivity and hydrometeors distribution. The deviations are mainly from the underestimation of total rainfall and upper level radar reflectivity less than 20 dBZ in the stratiform region.

The microphysical and macrophysical properties of clouds were influenced by aerosol concentrations. The reduction of cloud droplet sizes in polluted conditions inhibits collision and coalescence processes, thereby suppresses the warm clouds. With the added aerosol concentrations, the total cloud fractions increase in the middle and upper levels, primarily due to the increase of cloud fraction in stratiform regime. The underlying mechanism has been illustrated by Fan et al. (2013) that the reduction of ice and snow particle size leads to the reduction of their terminal velocities in the stratiform/anvil regime, which greatly reduces the dissipation of stratiform/anvils. The added aerosol loading increases the mixing ratio of ice-phase particles contributing to the enhancement of convective intensity at altitudes of 4–7 km in dirty conditions. Below 8 km, the cloud fraction of convective core is slightly increased, whereas the cloud fraction of convective core in Polluted scenario is reduced compared to the Control scenario above 8 km.

With the increased aerosol concentrations, the updraft speed is slightly enhanced at the altitudes of 4–8 km, especially over the convective core regime in the dirty conditions. Additional cloud droplets are generated with more released latent heat over the convective core in dirty conditions and then the subsequent latent heat release from ice-phased processes, which enhances the updraft in the convective core. Noted that the convective invigoration mainly occurs in the middle and upper levels through freezing more cloud water, which is relatively weak compared with the “warm-phase invigoration” as documented in Stolz et al. (2015), Fan et al. (2018), and Lebo (2018). The freshly added aerosol reduced the total rainfall by ~8.8% and delayed the onset of rainfall for about 30 min due to the suppression of warm clouds in the case study. The Polluted scenario produces a greater stratiform rain area fraction and smaller convective rain area fraction compared to the Clean scenario. Warm clouds are suppressed but the convective cores are slightly intensified induced from increased aerosol concentration, resulting in an overall effect of “localization and invigoration” of Mei-yu rainfall thus an elevated probability of short-term heavy rainfall.

Finally, the aerosol effects on the Mei-yu front were further studied. The added aerosol concentration enhances the frontal intensity and changes the local morphology of the front, although little effects on the mean location of the Mei-yu front can be detected. These findings are consistent with earlier theoretical
work suggesting the positive effects of convective heating in the genesis and maintenance of Mei-yu front via CISK mechanism and diabatic low-level potential vorticity generation. In summary, the aerosol-cloud system is very complex in which microphysical and dynamical processes are tightly linked and modulated by the thermodynamic properties of the environment.

It is worth noting that the signals of “convective invigoration” in our study with Thompson aerosol-ware scheme (bulk scheme) is relatively weak compared to those simulated by bin Microphysics from some earlier studies. The bulk model incorporating a saturation adjustment scheme may be one reason for this weak signal (Lebo & Seinfeld, 2011). And the uncertainties of bulk model in reasonably describing the ice-phase particles at the upper level of cloud mentioned in section 4.1 may also influence the simulation results. In addition, the errors caused by the nonlinear method used in the numerical sensitivity test, the uncertainties of model initial conditions and the limited ability of the model simulations all influence the simulation results. It is concluded that reducing uncertainty in simulations of ACI effect will likely require ensemble methods in addition to continued improvement of model parameterizations (Morrison, 2012). Furthermore, due to the limited number of Mei-yu front rainstorms observed by IMFRE, the ACIs during Mei-yu front rainfall will continue to be observed in future field campaigns and evaluated with ensemble methods.

Data Availability Statement

The authors would like to acknowledge China Meteorological Administration for providing the hourly precipitation data and can be accessed online (http://data.cma.cn). The ERA-5 data are available online (https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5). The input monthly aerosol data sets are available online (https://www2.mmm.ucar.edu/wrf/users/wrfv3.9/mp28_updated.html). The data sets from IMFRE 2018 are available online (http://doi.org/10.5281/zenodo.3633645).

References


