



RESEARCH ARTICLE

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Investigation of the Sensitivity of Tropical Cyclogenesis to Aerosol Intervention

Key Points:

- Aerosols of different sizes and concentrations have contrasting effects on microphysics and wind intensities of a developing vortex
- Coarse aerosols initially dampen vortex acceleration; fine/ultrafine aerosols boost it first but later weaken it more than coarse aerosols
- All aerosols hold promise for intervention, but technological limits impede the generation of the necessary volume of coarse aerosols

Supporting Information:

Supporting Information may be found in the online version of this article.

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Abstract As risks from tropical cyclones (TCs) are fueled by climate change escalation, there is an urgent need for transformational solutions to complement traditional approaches. Seeding TCs using aerosols can be a promising method to reduce cyclone intensity, supported by theoretical understanding of the microphysical effects of aerosols on TC clouds. The ideal time to intervene effectively in TCs is likely during their initial stage, before TC wind speeds reach their peak. However, studies exploring potential aerosol effects on TC formation remain scarce. This study investigates how a TC embryo responds to the addition of aerosols of varying sizes using the Weather Research & Forecasting (WRF) model coupled with a spectral-bin microphysics model. We found that aerosols of different sizes and concentrations distinctively affect the pre-TC vortex's microstructure and dynamics. Fine and ultrafine aerosols enhance the latent heat of condensation, freezing, deposition, and riming, initially intensifying the vortex. However, this results in enhancement of the cold pool, thereby reducing inflow and surface fluxes, subsequently weakening the vortex. Coarse aerosols produce the opposite effect to that of fine and ultrafine aerosols. Coarse aerosols lead to a slower initial acceleration owing to enhanced warm rain. However, the resulting weaker cold pool is insufficient to effectively reduce the strength of the vortex at the later stage. This study provides critical insights into how aerosols of varying sizes and concentrations modulate the energy cascade and impact the evolution of a TC embryo, laying the groundwork for further research on TC risk management through aerosol intervention.

Plain Language Summary Tropical cyclone (TC) risks are increasing as climate change intensifies, requiring new strategies to reduce risks alongside traditional methods. One potential approach is using aerosols (small airborne particles) to influence TCs during their early development stage to effectively prevent their spin-up. This idea is based on scientific understanding of how aerosols interact with clouds. However, there has not been much research exploring this possibility. In this study, we used computer models to investigate how adding aerosols of different sizes affects the early development of a TC. We found that fine (0.05–1 μm in diameter) and ultrafine ($<0.05 \mu\text{m}$) aerosols initially strengthen the storm by increasing heat released from water vapor condensing, freezing, and ice growth in the clouds. However, this also creates a stronger “cold pool” (a region of cooled air near the surface), which reduces energy inflow into the storm, eventually weakening it. In contrast, larger (coarse, 1–4 μm) aerosols initially slow down the storm by enhancing rainfall, but later create a weaker cold pool that hardly weakens the storm. This study helps us understand how aerosols of different sizes and concentrations influence TC formation, providing a foundation for future studies on using aerosols to manage TC risks.

1. Introduction

Tropical cyclone (TC) risks are growing with global warming. It is expected that there will be an increase in the frequency of intense, deluge-triggering storms in conjunction with poleward movement and further penetration inland, along with more rapid intensification (e.g., Bhatia et al., 2022; Bruyère et al., 2019, 2022; Knutson et al., 2020; Kossin et al., 2016; Tran et al., 2022; Tran, Ritchie, et al., 2024). As these risks grow, additional tools are required to protect people, infrastructure, and the environment from TCs. Intervening in the TC itself could complement traditional TC risk measures which aim to reduce exposure and vulnerability. Miller et al. (2023) recently evaluated TC intervention approaches. Based on scientific evidence and governance criteria, they

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demonstrated the potential of using aerosol injection for TC intervention. Aerosols can affect TCs through two primary mechanisms: aerosol-radiation interactions (ARI) and aerosol-cloud interactions (ACI). Through ARI, aerosols scatter and absorb sunlight, modifying the amount of solar radiation reaching the surface (e.g., Carlslaw, 2022), which in turn alters key environmental conditions critical to TC formation and development, including sea surface temperature (SST) and atmospheric instability. Through ACI, aerosols act as cloud and ice condensation nuclei altering TC cloud microphysics and dynamics (e.g., Fan & Li, 2022; Rosenfeld et al., 2012).

Over the last two decades, considerable effort has been focused on identifying mechanisms behind aerosol influences on TC microphysical and dynamical structures. This is of intrinsic scientific importance to TC modification. The hypothesis that invigorating TC peripheral convection using small hygroscopic particles could diminish TC intensity was first proposed by Rosenfeld et al. (2007) and Cotton et al. (2007). This hypothesis was reinforced by subsequent numerical and observational studies (Carrio & Cotton, 2011; Khain et al., 2010; Krall & Cotton, 2012; Liang et al., 2021; Rosenfeld et al., 2011; Wang et al., 2014; H. Zhang et al., 2007, 2009). Scientific explanations, as summarized in Rosenfeld et al. (2012), suggest that increasing the concentration of small aerosol particles at the periphery of the TC could enhance condensation, resulting in an abundance of small cloud droplets. These droplets are less likely to coalesce and are instead uplifted to supercooled altitudes where cloud ice hydrometeor formation is favored. Latent heat released both from low-level condensation and upper-level freezing invigorates the outer rainbands, enhancing ice precipitation. Enhanced cold rain resulting from melting cloud ice, along with subsequent evaporation in the lower atmosphere, creates a stronger cold pool that reduces the inflow of warm and moist air toward the TC core. Convection in the TC eyewall is suppressed, leading to a weakened TC.

The conceptual mechanism described by Rosenfeld et al. (2012) relates to TCs in the well-developed stage and aerosols generally smaller than 1 μm in diameter. However, aerosol-cloud interactions are sensitive to both TC lifecycle stage and aerosol size. Once TCs are fully fledged, strong dynamics within the TC circulation are likely to modulate aerosol effects (Cotton et al., 2007; Rosenfeld et al., 2007). Further modifications can be caused by sea spray generated by hurricane force winds (i.e., wind speeds ≥ 64 kts) in mature TCs (Hoarau et al., 2018; Jiang et al., 2019a, 2019b; Shpund et al., 2019a, 2019b). Shpund et al. (2019a, 2019b) found contrasting microphysical effects on the TC eyewall clouds, caused by wind-driven sea sprays with diameters in range of approximately 0.02–500 μm . Coarse sea salt particles act as small raindrops at the cloud base, actively gathering smaller droplets, producing large raindrops; however, a high concentration of fine sea spray below and above the freezing level fosters eyewall convection leading to the strengthening of the TC (Shpund et al., 2019b). Focusing on effects of ultrafine aerosols (diameter $<0.05 \mu\text{m}$) in a convective system, Fan et al. (2018) showed that these tiny particles which are generally considered too small for droplet nucleation, can significantly intensify convection under suitable conditions (i.e., high supersaturation).

The ideal time to alter TCs is likely during their initial or developing stage, before TC wind speeds reach their peak, to prevent them from evolving into intense systems (Rosenfeld et al., 2007, 2011). However, understanding of TC-aerosol interactions during these formative periods remains limited. There may be considerable differences in the impacts of aerosol on nascent versus mature TCs. In fully developed systems, aerosol effects primarily influence the outer cloud bands, since wind-induced sea spray aerosols dominate the eyewall clouds. In contrast, during early TC stages, aerosols can directly affect the vortex core. Although Rosenfeld et al. (2012) suggested that weak, poorly organized TCs might be intensified by the presence of pollutants—due to the invasion of aerosols into the circulation center in the absence of spiral rainbands, cold pools and downdrafts—this hypothesis has never been tested. These effects are expected to vary significantly depending on aerosol size and concentration, presenting an intriguing scientific challenge yet to be explored. Recently, Tran, Prinsley, et al. (2024) reviewed cyclogenesis theories and aerosol effects, suggesting two conceptual approaches for mitigating tropical cyclogenesis through targeted application of coarse or fine aerosols to interrupt convective development. These concepts require validation and further exploration through numerical modeling studies.

This paper quantitatively examines the impacts of aerosol size and concentration on a developing TC embryo at a cloud-resolving scale (1 km in resolution) using the Weather Research & Forecasting (WRF) model with a Spectral Bin Microphysics model (Fan et al., 2012; Khain et al., 2004). By focusing on a pre-TC weak vortex, we isolate the effects of injected aerosols from the influences of strong dynamic conditions and wind-induced sea sprays typically associated with fully fledged TCs. The specific objectives of this study are as follows: (a) advance understanding of how aerosols of varying sizes and concentrations impact the acceleration of a pre-TC vortex; (b)

establish the microphysical and dynamical driving mechanisms of aerosol effects on TC genesis; and (c) identify potential mitigation strategies of tropical cyclogenesis using a range of aerosol treatments.

This study lays the groundwork for developing effective aerosol intervention strategies for TCs, offering insights for risk reduction across different stages of their lifecycle—from early formation through maturity. Research methods are described in Section 2, and simulation results for aerosol effects on the development of the pre-TC vortex are discussed in Section 3. Section 4 presents the underlying mechanisms for aerosol impacts on the vortex microstructure and dynamics. In Section 5, conclusions and discussion are presented.

2. Research Methods

2.1. Modeling System

The WRF version 4.0.3 is used in this study (Skamarock et al., 2019) with microphysics parameterized using a modified version of the Fast Spectral Bin Model (FSBM; Khain et al., 2004). The superiority of FSBM to bulk schemes lies in its improved physical representation of key cloud-aerosol interaction processes such as activation and condensation (e.g., Fan et al., 2012, 2016; Y. Zhang et al., 2021). The plausibility of the simulated TC intensity and structure with FSBM has been demonstrated (Khain & Lynn, 2011; Khain et al., 2016). In FSBM, four size distributions are explicitly calculated for aerosols, liquid droplets and raindrops, low-density ice (snow and ice crystal), and high-density ice (hail and graupel). Each size distribution consists of 33 mass-doubling bins. In this study, the largest aerosol size in the 33rd bin is set to 4 μm in diameter. Aerosols over the 33 bins are prognostic with advection, activation, and regeneration considered.

A particular feature of the FSBM used in this study is the ability to account for aerosol regeneration from liquid hydrometeor (cloud droplets and raindrops) evaporation, following Fan et al. (2009). Aerosol scavenging by precipitation is not considered. The liquid hydrometeor distribution is represented by 33 bins; cloud and rain droplets are not distinguished in microphysics calculations but they are separated in the output of FSBM with a size threshold of 100 μm in radius. In addition, droplet homogeneous freezing is treated as an instant freezing of all droplets at temperatures below -38°C . This is different from the original SBM developed in Khain et al. (2004), which used the formulation of Bigg (1953) to represent both heterogeneous and homogenous freezing (Fan et al., 2010). The effective radius of droplet, ice and snow calculated in FSBM is used for shortwave and longwave radiation calculations. The graupel option is used for the high-density ice category. The version of FSBM used here also corrects bugs identified in previous versions such as in the parameterization of collision and melting. Other physical settings in the WRF simulations include the Rapid Radiative Transfer Model Shortwave and Longwave Schemes (RRTMG; Iacono et al., 2008), the Yonsei University Scheme for planetary boundary layer parameterization (YSU; Hong et al., 2006). The Kain-Fritsch cumulus parameterization (Kain, 2004) is employed only for the outer domain, which has a gridspacing of 5 km (see the section below).

2.2. Experimental Design

Previous studies report that subsequent changes in the meteorological conditions induced by aerosols modulate their primary effects on convective clouds (e.g., Fan & Li, 2022; Zang et al., 2023). Dust aerosol intrusions can also influence the radiative balance, hydrological cycle, and large-scale environments, affecting TC formation (e.g., Pan et al., 2018, 2024). To disentangle aerosol effects, the idealized TC case experiment in WRF is utilized, excluding the impacts of large-scale meteorological conditions and regional dust influences. A weak and slowly developing vortex is configured to exclude influences of strong dynamical processes and sea spray generated by intense winds commonly associated with mature TCs.

Although air-sea interaction significantly influences TC intensity through surface cooling (“cold wake”) from enhanced vertical mixing (e.g., Bender & Ginis, 2000; Black et al., 2007; D’Asaro et al., 2007; I.-I. Lin et al., 2013), recent work shows ocean feedback only substantially modifies aerosol effects when wind speeds exceed 40 m s^{-1} (Y. Lin et al., 2023). Since TCs in their genesis stage rarely reach the intensities sufficient to cause significant cold-wake effects, ocean coupling is not included in this study. This simplification avoids the intricacy and computational burden of modeling minor ocean feedbacks. The ultimate goal is to isolate the pure aerosol effects on TC initiation, providing a foundation for future research on TC mitigation through the development of effective seeding strategies.

Table 1
Model Simulations and Corresponding Aerosol Statistics

Experiment	Acronym	Particle diameter (D — μm)	Concentration (cm^{-3}) Ratio to Control run
Marine background 1—pristine condition	<i>MarBG1</i>	$D \leq 4$	195 0.67
Marine background 2—pristine condition (Control run)	<i>MarBG2</i>	$D \leq 4$	290 1
Marine background 2 with coarse aerosol perturbation	<i>Coarse</i>	$1 < D \leq 4$	617 2.13
Marine background 2 with fine aerosol perturbation	<i>Fine</i>	$0.05 < D \leq 1$	3585 12.36
Marine background 2 with ultrafine aerosol perturbation	<i>UFine</i>	$D < 0.05$	5596 19.30

To achieve the configuration that best fits the purpose of this study, we carried out sensitivity tests to evaluate the effects of domain size, grid spacing, environmental sounding, and structure and intensity of the initial vortex with the single-moment 6-class microphysics scheme (WSM6; Hong & Lim, 2006) because the FSBM scheme is highly computationally expensive. These sensitivity analyses showed that the following settings produce a vortex reaching a quasi-steady state after around 10 days—thereby satisfying the aforementioned criteria: the initial maximum winds of 6 m s^{-1} at the radius of 40 km; the initial outer radius of 250 km; and the vortex depth of 10 km. The Coriolis parameter is set up to allow simulations of a Northern Hemisphere vortex at 20.05°N and the SST is uniformly fixed at 28°C . For atmospheric background conditions, the widely used Dunion input sounding (Dunion, 2011), which simulates an external atmospheric environment that is strongly conducive to TC development (e.g., Bruyère et al., 2019), is utilized in this study.

Based on a compromise between the required resolution and computational cost, and to target a small-scale vortex, the two nesting grid domains (i.e., inner and outer grids) were both set up with 401×401 points with the corresponding grid spacings of 5 and 1 km, respectively. The 1 km grid spacing allows deep convective clouds within TCs as well as aerosol-cloud interaction processes to be resolved (e.g., Khain et al., 2010; Lynn et al., 2016). There are 56 vertical levels in the model, and the top level is set at an altitude of 20 km. The Dunion sounding provides weak wind shear conditions leading to a more realistic idealized simulation. However, the vortex is slightly steered away by environmental winds. The vortex following technique is, therefore, applied to maintain the vortex in the middle of the nesting domain.

Five aerosol size distributions (SDs) are used in simulations to examine the impacts of aerosols of different sizes and concentrations on TC initiation. Details on simulations and corresponding aerosol SDs are provided in Table 1 and Figure 1. Two SDs are included to simulate a realistic pristine aerosol condition. These are: *MarBG1*, which is described as the typical marine aerosol concentration in Ghan et al. (2011); and *MarBG2*, which is adjusted from *MarBG1* with an increase in concentration of aerosols with diameter $>0.01 \mu\text{m}$. The preliminary simulation results show that *MarBG1* produces quite low droplet concentrations (i.e., less than 20 cm^{-3} in general at the cloud base), whereas cloud droplet concentration in *MarBG2* is approximately 50 cm^{-3} which is closer to the clean background condition (i.e., $30\text{--}50 \text{ cm}^{-3}$) observed in the tropics (Rosenfeld et al., 2019). Therefore, *MarBG2* is chosen as the control run and other perturbations are built on this pristine condition. The sensitivity of the vortex to aerosols in *MarBG1* and *MarBG2* will be discussed in Section 3.

The *UFine* and *Fine* simulations are used to test the effect of adding ultrafine and fine particles into a clean atmosphere (characterized by the *MarBG2* aerosol concentration). These designs were based on typical ship emission aerosol concentrations, using actual measurements to determine a practical reference number (i.e., the Star Livorno plume size distribution in Figure 5 in Hobbs et al., 2000). The simulation with a coarse aerosol distribution—*Coarse*, is adjusted from the SD of 30 m s^{-1} wind-induced sea spray produced in Shpund et al. (2019a, 2019b). Note that the aerosol type in all simulations is sea salt particles and the differences among simulations lie in the sizes and concentrations of the perturbed aerosols. Because of the weak dynamics in the TC initiation stage, there is insignificant differentiation in aerosols between the periphery and the vortex center. Hence, aerosol distribution across the domain is set up uniformly at the initialization.

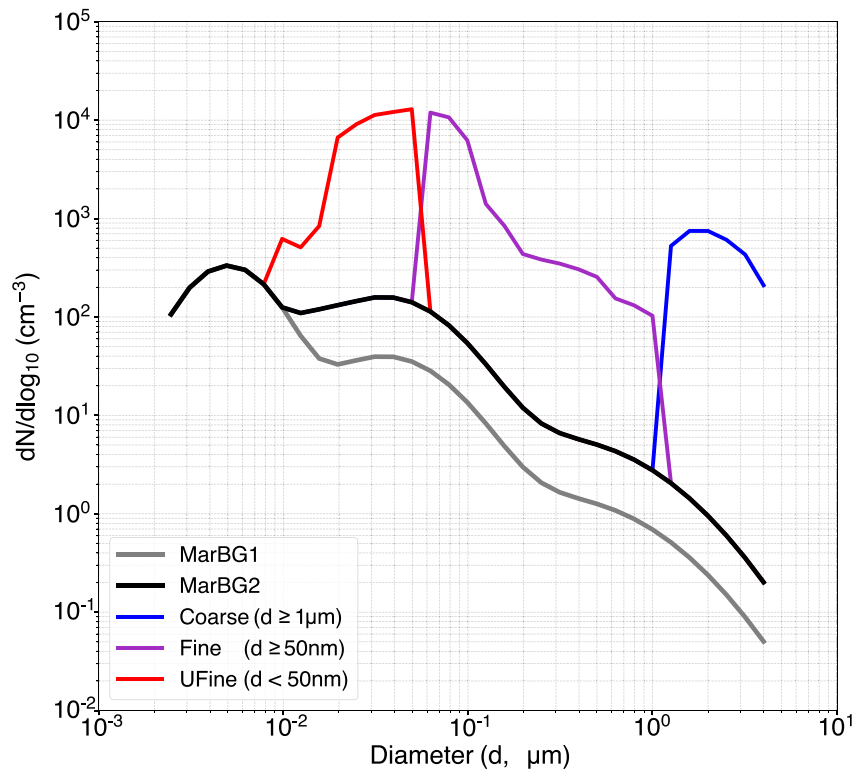


Figure 1. Aerosol size distributions in the background (*MarBG1* and *MarBG2*) and perturbed conditions using varying sizes and concentrations (*Coarse*, *Fine*, and *UFine*), examined in this study. Each aerosol perturbation includes the artificial aerosols added over the background aerosol condition in *MarBG2*.

2.3. Analysis Method

The model simulation outputs were post-processed to calculate a number of variables used to investigate the microphysical effects of aerosols, viz: supersaturation (S), water vapor path (WVP), cloud water path (CWP), rain water path (RWP), snow water path (SWP), graupel water path (GWP), and cloud ice path (CIP). Total ice path (denoted as TIP) is calculated using the sum of cloud ice, graupel and snow quantities. Cloud properties (e.g., mixing ratios of cloud and precipitation particle types, supersaturation, and vertical velocity) in strong updrafts are investigated following Fan et al. (2018), in which strong updrafts are defined as top 10th percentiles of vertical velocities greater than 2 m s^{-1} .

To assess aerosol-induced changes in TC destructive potential, integrated kinetic energy (IKE ; Powell & Reinhold, 2007), which takes into account the entire surface wind field of the TC circulation, is calculated after Kreussler et al. (2021):

$$IKE = \int_v \frac{\rho U^2}{2} dV \quad (1)$$

where U is the wind speed at the model lowest level of grid points within 200 km radius from the vortex center, V is the volume calculated from the domain area and the thickness of the first layer. Additionally, IKE_{volume} denoting the total energy of the vortex is calculated by integrating IKE across 56 model vertical levels. In the subsequent section, the development of the vortex under different aerosol treatments is presented.

3. Vortex Development Under Various Aerosol Perturbations

3.1. Impact of Aerosol Regeneration on Supersaturation

Aerosol regeneration could be an important process affecting the influence of aerosols on TCs. Therefore, we first evaluate the effect of aerosol regeneration on aerosol condensation nuclei (CCN) number, cloud droplet number

concentration, and supersaturation, which are important quantities for aerosol-cloud interactions. The level mean *CCN* concentrations (i.e., averaged for each level and plotted as a time series of vertical profiles) in the conditions of clear-sky (total mixing ratio of all cloud and precipitation particle types $< 1 \cdot 10^{-7} \text{ kg kg}^{-1}$) and cloudy-sky (total mixing ratio of all cloud and precipitation particle types $> 1 \cdot 10^{-5} \text{ kg kg}^{-1}$) between the simulations of fine aerosol perturbation with and without recycled aerosols are compared (Figures 2a–2d). As the activation efficiency is highest in the fine aerosol perturbation, due to its concentration and size (discussed in detail in Section 4), we selected fine aerosols to evaluate the impacts of aerosol regeneration. The comparison results show that aerosol regeneration leads to the reproduction of a substantially high concentration of *CCNs* at all levels up to 5 km. This is due to droplet evaporation since clouds start to form throughout the simulation in the cloudy condition (i.e., typical increases of approximately $150\text{--}550 \text{ cm}^{-3}$, except some large increases of $800\text{--}900 \text{ cm}^{-3}$ at 1.5–2.5 km altitudes from $t = 27\text{--}36 \text{ hr}$; Figure 2d). The aerosol regeneration process also significantly replenishes ambient *CCN* concentrations as seen in Figure 2b where typical increases of approximately $250\text{--}850 \text{ cm}^{-3}$ are simulated.

The increase in *CCN* concentration enhances droplet nucleation. Comparing the mean cloud droplet concentration ($\bar{N}_d$), defined as the average of cloud droplet number within strong updrafts and averaged for each level, shows an increase of approximately $100\text{--}800 \text{ cm}^{-3}$ below 6.5 km altitude (Figures 2g and 2h). The increased cloud droplets due to aerosol regeneration favor condensation on these nucleated droplets, hence within clouds, supersaturation is significantly reduced for all levels up to 8.5 km (Figures 2e and 2f). These results demonstrate a more realistic representation of cloud processes, indicating better-resolved cloud microphysics in the model.

3.2. Evolution of TC Intensity, Energy, and Rainfall

In both simulations for background aerosol conditions (*MarBG1* and *MarBG2*), the evolution of the vortex consists of an intensifying phase and a decaying phase with some re-intensification occurring near the end of the simulation period (Figures 3a and 3b). The vortex with a higher background aerosol concentration (i.e., *MarBG2*) is slightly stronger and is accompanied by slightly higher destructive potential and total energy compared to *MarBG1* (Figures 3a, 3c, and 3d). Recall that the destructive potential (i.e., near-surface energy) and the total energy of the vortex are defined as the areas under the respective *IKE* and *IKE_{volume}* curves in Figures 3c and 3d. The rain rate and total rain amount do not differ greatly between *MarBG1* and *MarBG2* (Figures 3e and 3f). This indicates that a small addition of aerosols, in all size ranges, into very clean background air slightly enhances the vortex strength.

The results of perturbations with aerosols of varying sizes and concentrations reveal that the evolving characteristics of the vortex are sensitive to aerosol size and concentration. Simulations where the vortex is stronger in the intensifying phase exhibit a more rapid decay later (e.g., after 27 hr). Specifically, as seen in Figures 3a and 3b, Figure S1 in Supporting Information S1, and Table 2, *Fine* (purple) and *UFine* (red) show the largest strengthening and weakening of the vortex in the corresponding (intensifying and decaying) phases with the peak maximum wind speeds of 13.7 and 13.5 m s^{-1} , respectively. After 33 hr, stronger maximum winds (approximately $9\text{--}11 \text{ m s}^{-1}$) are seen for both *Fine* and *UFine* compared to other simulations (Figure 3b), but it is important to note that these are localized extreme wind speeds. The overall wind field of the vortex in these simulations remains relatively weak as illustrated in Figures S1d1–S1d4, and S1e1–S1e4 in Supporting Information S1. The addition of coarse (blue) aerosols prevents the vortex circulation from rapidly accelerating and decreases the vortex intensity during the intensifying stage (i.e., peak at 10.9 m s^{-1} at around 24 hr), but the mitigation efficiency in the later stage is less than simulated for the fine and ultrafine aerosol perturbations (Figures 3a and 3b; Figure S1 in Supporting Information S1; Table 2).

The time series of *IKE* shows the most significant decrease in the destructive potential of the vortex in the perturbation with fine aerosols—approximately 25% reduction compared to *MarBG2* (Table 2). This is also true for the total energy of the vortex (i.e., *IKE_{volume}*) with the largest decrease of about 17% in *Fine* (Figure 3d; Table 2). Similar levels of near-surface energy reduction are simulated for *Coarse* and *UFine* (i.e., approximately 20%–21% in *IKE*). However, coarse aerosols reduce the total energy of the vortex by a greater amount than ultrafine aerosols (i.e., 16% and 11%, respectively, Table 2). The lower mitigation efficiency in *IKE_{volume}* in *UFine* is due to the more substantial increases in wind speeds across vertical levels near the TC center and from 2.5 to 9 km altitude in the TC outer region compared to the *Coarse* simulation (Figures S2b and S2d in Supporting Information S1).

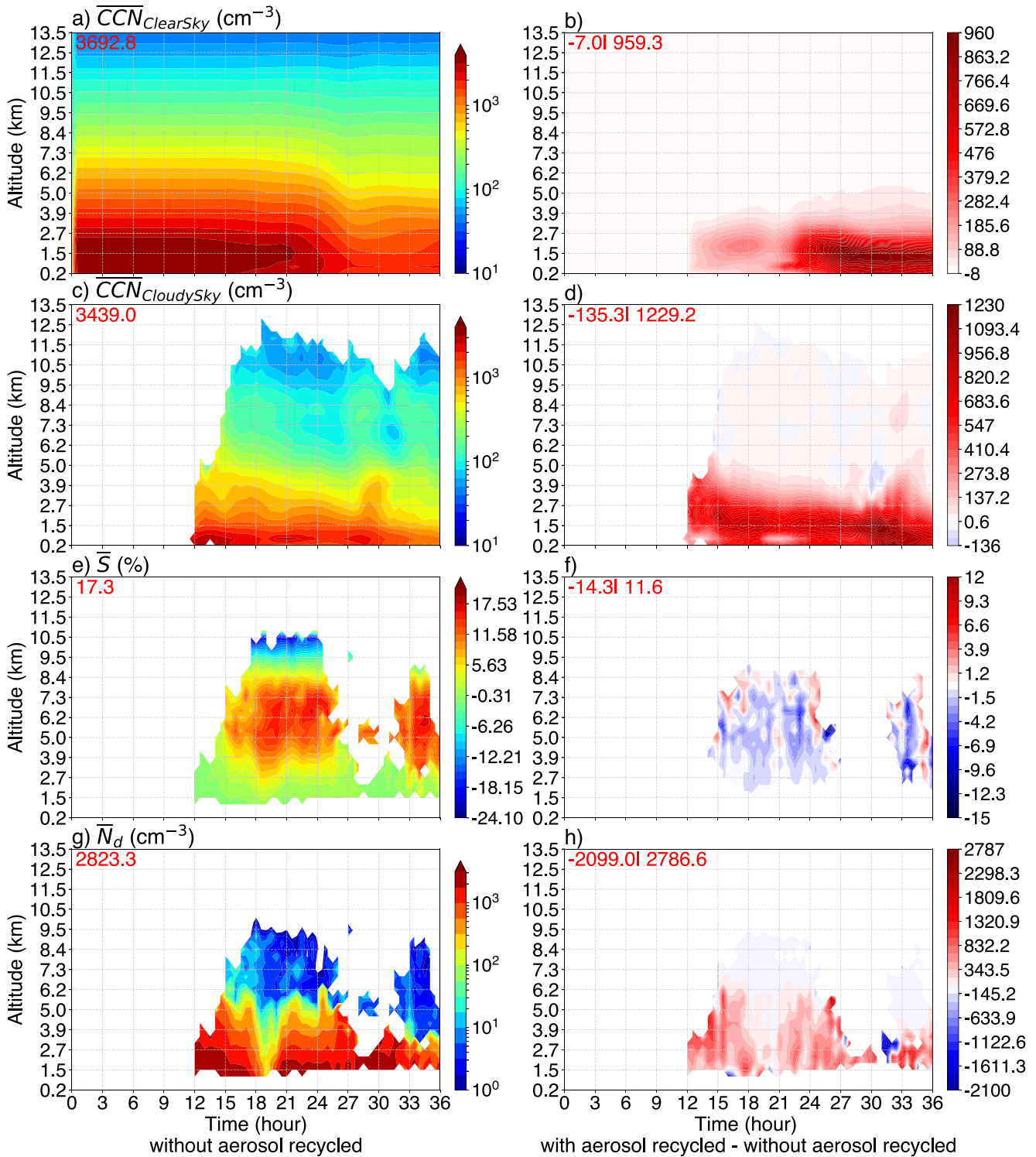


Figure 2. Time series of vertical profiles of mean: (a) aerosol concentration in the clear-sky condition ($\overline{CCN}_{ClearSky}$); (c) aerosol concentration in the cloudy-sky condition ($\overline{CCN}_{CloudySky}$); (e) supersaturation ($\overline{S}$); and (g) cloud droplet concentration ($\overline{N_d}$) for the perturbation simulation of fine aerosols without aerosol regeneration [left column]; and the difference in the corresponding parameters between the simulations with and without aerosol regeneration [right column] in (b), (d), (f), and (h). $\overline{S}$ and $\overline{N_d}$ are calculated for the top 10th percentile of updrafts exceeding 2 m s^{-1} within the inner domain of corresponding simulations. The red text on the top left of each panel in the left column indicates the maximum value, whereas the red text on the panels in the right column shows the minimum and maximum values.

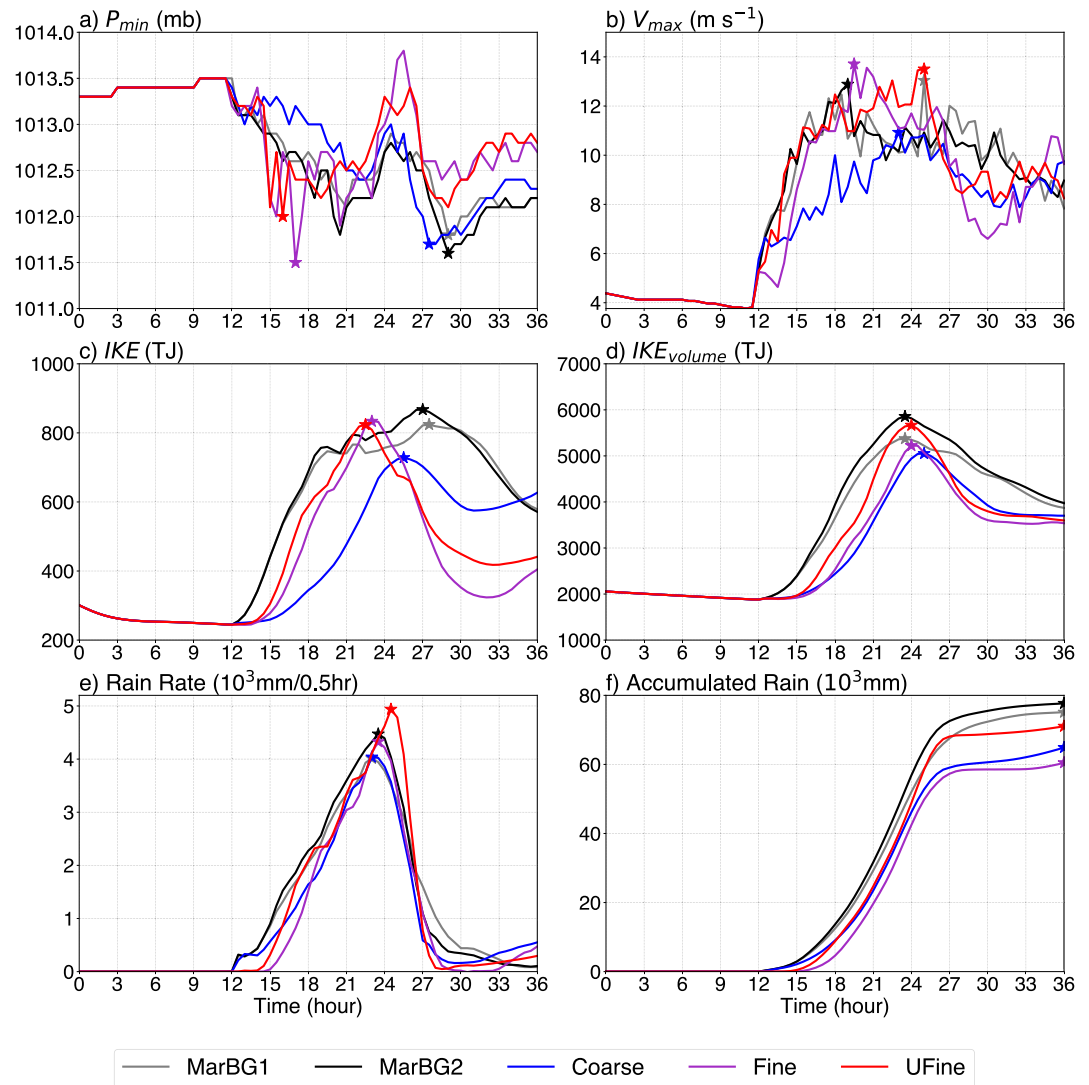


Figure 3. Time series of: (a) center pressure (P_{min}); (b) maximum winds at surface (V_{max}); (c) near-surface integrated kinetic energy (IKE); (d) integrated kinetic energy volume (IKE_{volume}); (e) integrated rain rate; and (f) accumulated rainfall of corresponding simulations. The stars in each panel indicate the most extreme values. Integrated rain rate and accumulated rainfall are calculated by adding all corresponding rain rate and accumulated rain values within 200 km (horizontally) of the TC center at every 30-min time frequency.

Table 2

Extreme and Integrated Values of Intensity, Integrated Kinetic Energy, and Rainfall Among Simulations

Experiment	P_{min} (mb)	V_{max} (m s ⁻¹)	Rain rate _{max} (mm 0.5 hr ⁻¹)	IKE_{max} (TJ)	^a IKE (TJ) [^b ΔIKE (%)]	^c IKE_{volume} (TJ) [^b ΔIKE_{volume} (%)]	^d Total rainfall (mm) [^b Δ rain (%)]
MarBG1	1011.8	13	4015.1	824.1	38268.9	245571.6	75118.9
MarBG2	1011.6	12.9	4472.1	867.2	38923.3	255361	77668.7
Coarse	1011.7	10.9	4038.4	728.2	30676.4 [−21.2%]	215058.5 [−15.8%]	64891.7 [−16.5%]
Fine	1011.5	13.7	4326.7	833.6	29360.8 [−24.6%]	213072.1 [−16.6%]	60486.7 [−22.1%]
UFine	1012	13.5	4938.3	823.8	31309.7 [−19.6%]	226421.2 [−11.3%]	71047.4 [−8.5%]

^a IKE indicates the total integrated kinetic energy near the surface (i.e., the areas under the curves in Figure 3c). ^bPercentage changes in IKE , IKE_{volume} , total rainfall are calculated based on a comparison to the MarBG2. ^c IKE_{volume} indicates the total integrated kinetic energy across the volume of the vortex (i.e., the areas under the curves in Figure 3d). ^dTotal rainfall denotes the accumulated rainfall in each simulation.

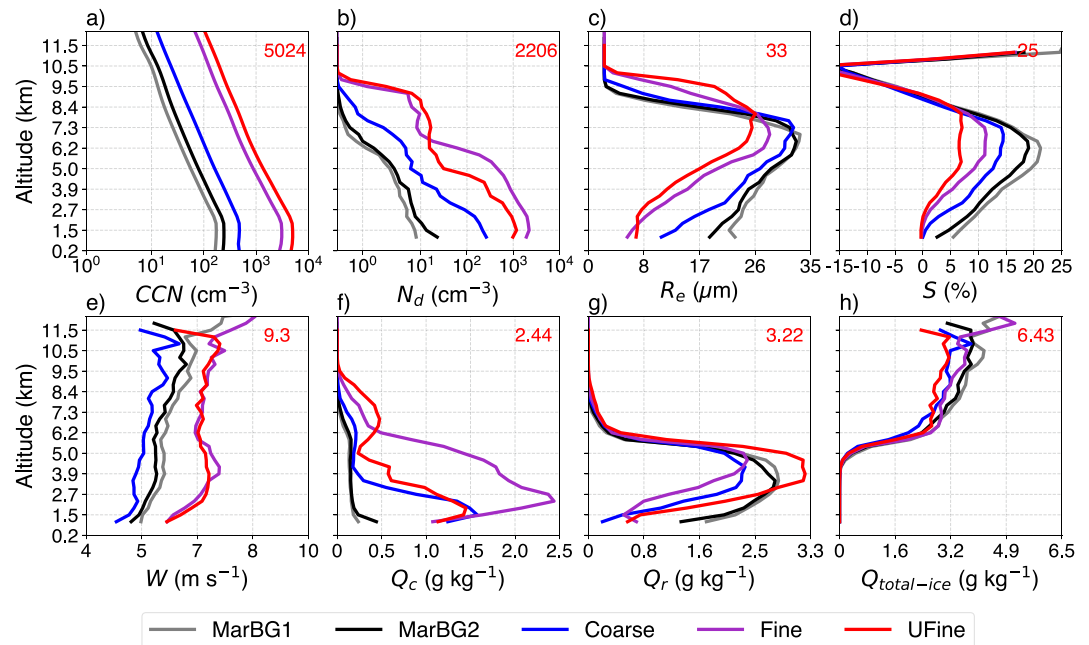


Figure 4. Temporally averaged vertical profiles of: (a) cloud condensation nuclei concentration (CCN); (b) cloud droplet number (N_d); (c) cloud droplet effective radius (R_e); (d) supersaturation (S) in the top row; and (e) updrafts (W); (f) cloud water mixing ratio (Q_c); (g) rain water mixing ratio (Q_r); and (h) total solid hydrometeor mixing ratio (the sum of cloud ice, graupel, snow quantities; $Q_{total-ice}$) in the bottom row. Data are calculated for the top 10th percentile of updrafts exceeding 2 m s^{-1} within the inner domain of corresponding simulations. The red text on the top right of each panel denotes the maximum value.

Without considering aerosol impacts, rain rate generally increases with increasing TC intensity (Rogers et al., 2009). The reduced rain rate and rain accumulation with the injection of coarse aerosols (relative to *MarBG2*) is well correlated with the weaker intensity of the vortex in this perturbation (Figures 3a, 3b, 3e, and 3f). However, despite similar vortex intensities for the *UFine* and *Fine* simulations, there are differences in rainfall amounts. This highlights the different influences that aerosol size and concentration have on the correlation between TC intensity and TC rain. Specifically, higher rain rates in *UFine* compared with *Fine*, lead to a larger total rainfall accumulation as shown in Figures 3e and 3f. As a result, the addition of ultrafine aerosols reduces TC rain by approximately 9%, whereas the effect of adding fine aerosols is more than double this rate, at about 22% (Table 2). The underlying mechanisms by which aerosols of different sizes and concentrations regulate the vortex microstructure and dynamics, which lead to these evolving behaviors of TC rainfall, intensity, and energy, are explored in the following section.

4. Underlying Mechanisms for Aerosol Impacts on the Vortex Microstructure and Dynamics

In this section, we will first discuss the effects of aerosol size and concentration on each microphysical regime, that is, warm phase, mixed- and ice- phases, and the corresponding response of the vortex's dynamics. These impacts will be discussed in parallel to highlight the differences among aerosol perturbations. A summary is provided in Section 4.4 to synthesize the results and portray a comprehensive picture of the mechanisms relating to microphysics and dynamical feedbacks associated with each aerosol perturbation.

4.1. Warm-Phase Microphysics

Figures 4a, 4b, and 4d show that the small increase in concentration of aerosol particles with $D > 0.01 \mu\text{m}$ from *MarBG1* to *MarBG2* increases condensational efficiency. A more plausible value of temporally averaged cloud droplet concentration at the cloud base (approximately 600 m; Figure 4b), associated with the pristine marine atmosphere, is found for *MarBG2* rather than *MarBG1*. This demonstrates that *MarBG2* provides a more suitable background aerosol concentration for use in the control run.

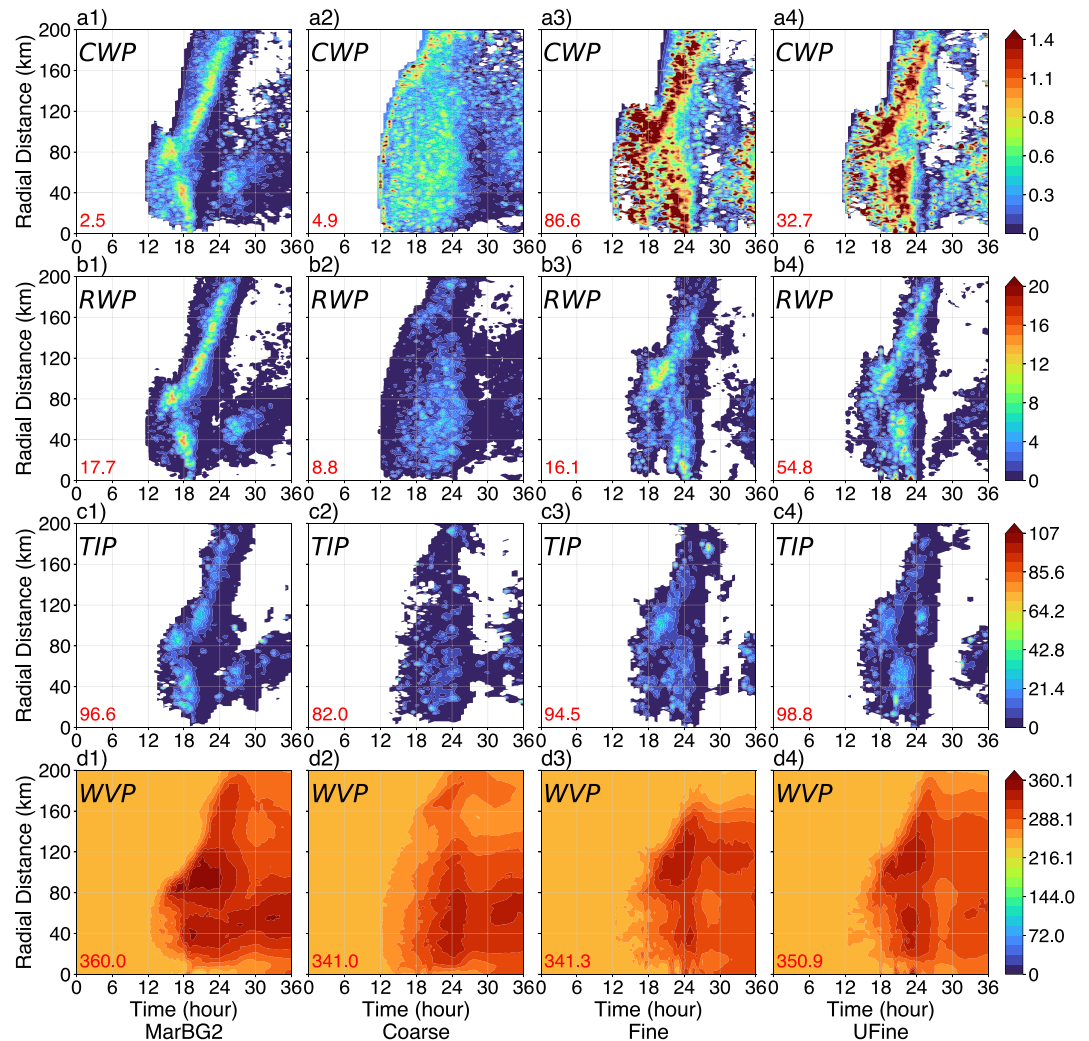


Figure 5. Time-radius composite of azimuthally averaged: (a1–a4) cloud water path (CWP; kg m⁻²); (b1–b4) rain water path (RWP; kg m⁻²); (c1–c4) total ice path (TIP; kg m⁻²); and (d1–d4) water vapor path (WVP; kg m⁻²) of corresponding simulations. The red text on the bottom left of each panel denotes the maximum value.

Compared to the perturbation simulations, the *MarBG2* control run has the lowest aerosol concentration—hence CCN loading, resulting in the fewest cloud droplets, the largest droplet effective radius, and the highest supersaturation (Figures 4a–4d; Figures S3a1–S3a4, S3b1–S3b4 in Supporting Information S1). Coalescence and collision between cloud droplets promote fast warm rain below 2.7 km from 12 hr onwards (Figure 3e; Figure S4b2 in Supporting Information S1). Rain intensity in the control run is strongest both near the vortex center and at further radii during 16–25 hr when the vortex reaches and passes its peak intensity (Figures 3e, 5b1, and 6al).

In the aerosol perturbation simulations, the addition of aerosols increases droplet number concentration (Figures 4a and 4b; Figures S3a1–S3a4 in Supporting Information S1) and significantly enhances condensation, indicated by the substantial decreases in supersaturation below 8.4 km (Figure 4d; Figures S3b1–S3b4 in Supporting Information S1), decreases in water vapor content (Figures 5d1–5d4), and increased cloud water content (Figures 4f and 5a1–5a4; Figures S12a1–S12a4 in Supporting Information S1). The injected large-sized (coarse) aerosols are easily activated leading to the most widespread distribution of cloud water across the vortex circulation (Figure 5a2). Because of the lower droplet concentration in *Coarse* compared to other perturbations, cloud droplet effective radius is the second largest after *MarBG2* (Figures 4b and 4c). Larger drops coalesce and collide, significantly increasing the likelihood of rapid warm rain formation once clouds form (Figures 3e and 5b2). As a result, supersaturation is still high in this case (Figure 4d; Figure S3b2 in Supporting Information S1),

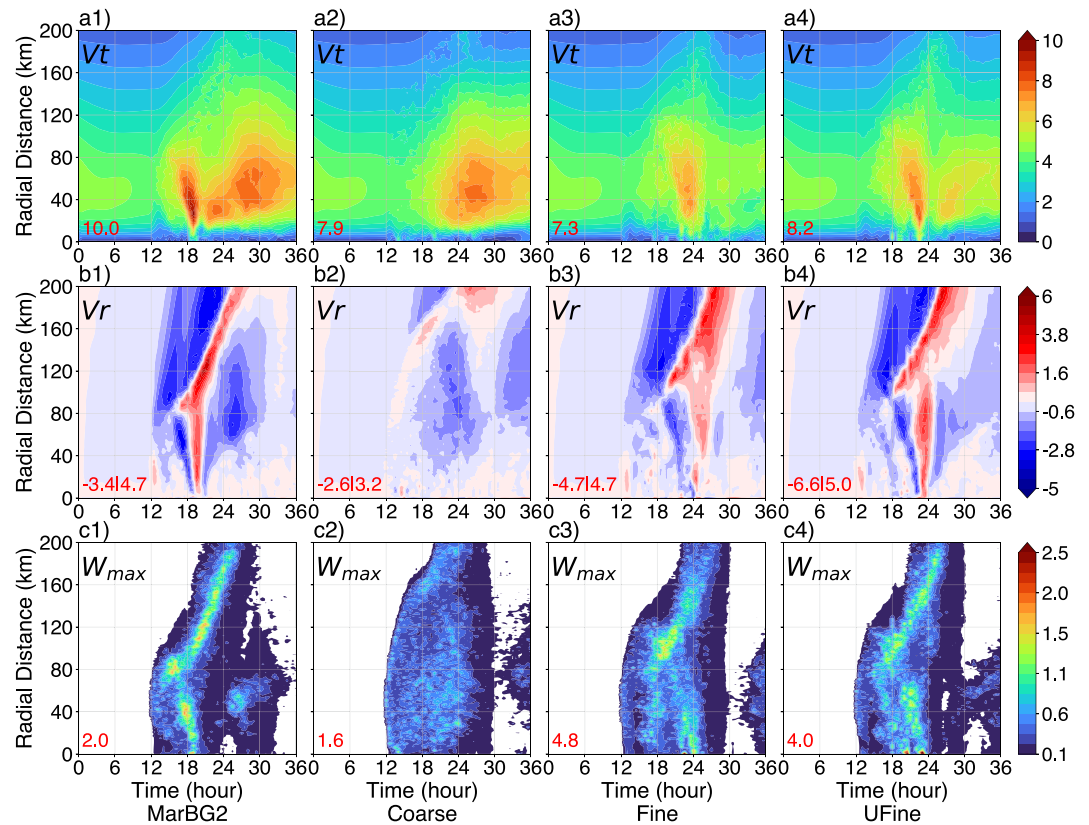


Figure 6. Time-radius composite of azimuthally averaged: (a1–a4) tangential winds (V_t ; m s^{-1}); (b1–b4) radial winds (V_r ; m s^{-1}) at 200 m; and (c1–c4) maximum updrafts (W_{max} ; m s^{-1}) of corresponding simulations. Blue and red shaded colors in (b1–b4) represent inflow and outflow, respectively. The red text on the bottom left of (a1–a4, c1–c4) denotes the maximum value, whereas that on (b1–b4) shows the minimum and maximum values.

convection is suppressed (Figures 4e and 6c2); and *Coarse* manifests the most extensive rain area spreading from the center to the outer radii of the vortex (Figure 5b2).

Ultrafine particles, due to their extremely small size, require a high critical supersaturation to activate and form droplets (i.e., Fan et al., 2018). However, their activation and subsequent condensation on newly formed cloud droplets can profoundly reduce supersaturation, primarily driven by high aerosol concentration. This is exactly what we see for the *UFine* simulation (Figures 4b and 4d; Figure S3b4 in Supporting Information S1). Compared to *Fine*, *UFine* generally has the highest cloud droplet concentration (N_d) at altitudes above approximately 6.8 km, due to large activation of ultrafine particles that are transported aloft (Figure 4b; Figures S3a3–S3a4 in Supporting Information S1). The lower N_d below 6.8 km on average (Figure 4b) is mainly because at the initial and later stages fewer ultrafine aerosols can be activated (Figures S4a2, S5a2, S6a2; S4a4, S5a4, S6a4; S8a2, S9a2, S10a2; S8a4, S9a4, a10a4 in Supporting Information S1) as supersaturation is less than the critical supersaturation required for activation. We note the complexity of temporal evolution of convective systems and aerosol effects. Specifically, due to fewer cloud droplets during the first few hours after cloud formation relative to *Fine*, *UFine* has a larger cloud droplet effective radius (R_e ; Figures S4a2–S6a2, and S4a3–S6a3 in Supporting Information S1). This initially large R_e supports coalescence, leading to fast warm rain above 2.5 km at 13 hr in *UFine* (Figure 3e; Figure S5b2 in Supporting Information S1). This warm rain causes a reduction in the integrated droplet surface area available for condensation (Fan et al., 2018). As a result, more water vapor is available in the clouds, increasing supersaturation. Increases in supersaturation, in turn, allow secondary nucleation to occur at higher altitudes, around 4.5 and 5.5 km, where ultrafine particles activate at 14 hr (Figure S6a2 in Supporting Information S1). Owing to the supply of ultrafine particles and moisture from low levels by updrafts, new cloud droplets continuously form and then precipitate, creating a positive loop that increases both cloud water and

rainwater contents in subsequent timesteps (Figures S7b1 and S7b2 in Supporting Information S1; Figures 5a4 and 5b4).

Supersaturation in *Fine* is not as low as *UFine* (Figure 4d; Figures S3b3 and S3b4 in Supporting Information S1) because its aerosol concentration is 1.5 times lower (Figures 1 and 4a; Table 1). Larger aerosol particle size in *Fine* (Figure 1) increases the rate of aerosol activation, leading to the highest concentration of cloud droplets at low levels (Figures 4b and 5a3; Figures S3a1–S3a4 and S12a1–S12a4 in Supporting Information S1). Up to 14 hr, the competition for water vapor by cloud drops leads to the smallest effective radius and supersaturation in *Fine* compared to other simulations (Figures S4–S6a3–S6a4 in Supporting Information S1). The less efficient coalescence and collision compared with *UFine* and *Coarse*, caused by the markedly small effective radius of droplets delays warm rain until 15 hr (Figures 3e and 5b3; Figures S4–S7b2 in Supporting Information S1). Cloud droplets form successively over time but do not rain out because of their small sizes (Figures 5a3–5b3; Figures S4–S7a2–S7a3, S7b2 in Supporting Information S1). Pan et al. (2020) explained the “aerosol-hindered warm precipitation” by the generation of a large number of uniformly sized small droplets impeding collision. Here, the fine particles are shown to be most effective in suppressing warm rain compared to other aerosol sizes examined in this study (Figure 3f).

4.2. Mixed-Phase and Ice-Phase Microphysics

The terminologies for solid hydrometeors in this study include cloud ice, graupel and snow. The total content of solid hydrometeors in the background simulation (*MarBG2*) is largest during the intensifying stage (i.e., 16–25 hr; Figures 5c1 and 6a1) contributed most by snowflakes, followed by graupel (Figures S11, S13a1–S13c1 and Table S1 in Supporting Information S1). This results from the transportation by strong updrafts (Figure 6c1) of large-sized cloud droplets (Figure 4c) to levels above freezing level (i.e., roughly at around 5.5 km in this study as shown in Figure S12d1 in Supporting Information S1). The cloud properties found in *MarBG2* with warm rain at lower levels and high snow content in the upper levels align with unique features of TC clouds in a pristine maritime environment (e.g., Shpund et al., 2019b).

In *Coarse*, fewer cloud droplets are lifted to higher levels due to their relatively large size (Figure 4c), faster formation of warm rain (Figure 5b2), and weaker updraft conditions (Figure 6c2). The altitude that they can reach is also restricted, hence the process of undergoing cooling and transforming into solid hydrometeors is limited. Therefore, the amount of each solid hydrometeor type, and the total quantity of solid hydrometeors, in *Coarse* is smallest among simulations (Figure 5c2; Figures S13a2–S13c2 and Table S1 in Supporting Information S1).

Compared to *Coarse*, cloud droplets in the *UFine* simulation are much smaller (Figure 4c), increasing the likelihood of being lifted to higher levels, fostering the formation and growth of graupel and cloud ice (Figures S11a, S11b, S12c4, and S12d4 in Supporting Information S1). This process is supported by stronger updrafts in *UFine* (Figure 4e). An increase in effective radius with height is seen below 7.5 km, whereas this decreases above this altitude, indicating a preferential freezing of larger drops below 7.5 km and smaller droplets above this height because smaller droplets are transported to higher altitudes (Figure 4c). The notable presence of cloud ice during the intensifying stage of the vortex is at altitudes above 9.5 km (Figures S11a and S12c4 in Supporting Information S1). At lower levels, graupel forms on existing cloud ice, initiating at around 5.5 km (Figures S11b and S12d4 in Supporting Information S1), and subsequently gathers the newly nucleated droplets. The melting of graupel (Figure S15c1 in Supporting Information S1) contributes to downdraft cold rain generation, which, combined with substantial warm rain at lower levels (Figure S12b4 in Supporting Information S1), leads to the most intensive rain rate and largest rainfall accumulation in *UFine* among perturbations (Figures 3e and 3f). This is consistent with invigoration effects on TC precipitation caused by ultrafine particles from anthropogenic sources reported in previous studies (e.g., Pan et al., 2020; Qu et al., 2017; Wang et al., 2014).

Only a small fraction of cloud water precipitates as rain in *Fine*—referred to as “aerosol-hindered warm rain” in Section 4.1 (Figures 4f and 4g). Instead, the predominant cloud water content in *Fine* among the aerosol perturbations consists of a substantial amount of small-sized cloud droplets retained below the 0°C isotherm level (Figure 4f). In addition, a portion of cloud droplets is transported to higher elevations, where they convert to supercooled liquid water, facilitating riming growth raising cloud ice quantity (Figures S11a, S13a3, S14d1, and S14d2 in Supporting Information S1). Particularly, the supercooled droplets collide and freeze onto the formed cloud ice crystals, which then aggregate into snowflakes (Figures S11c and S13c3 in Supporting Information S1). Some of these snowflakes subsequently become coated by supercooled water droplets, generating graupel

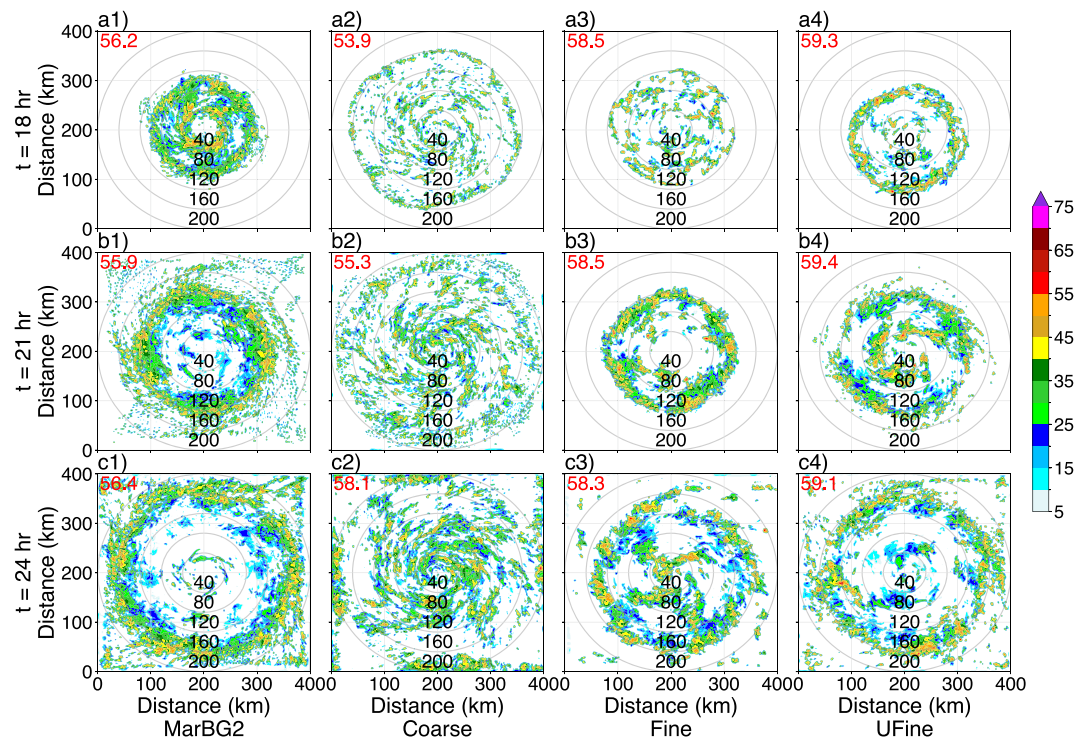


Figure 7. Horizontal cross-section of maximum radar reflectivity (unit in dBZ) of corresponding simulations at: (a1–a4) 18 hr; (b1–b4) 21 hr; and (c1–c4) 24 hr. The red text on the top left of each panel denotes the maximum value.

(Figures S11b and S13b3 in Supporting Information S1). As a result, the quantities of each solid hydrometeor species—including cloud ice, graupel, and snow particles—are largest, leading to the highest total mass of solid hydrometeors in *Fine* among the perturbations (Figures S13a3–S13b3 and Table S1 in Supporting Information S1).

4.3. Dynamical Feedback as a Result of Aerosol-Cloud Interactions

In the background condition (*MarBG2*), heavy rain during the period of 18–25 hr (Figure 3e) contributes to strong convective downdrafts via enhancing latent cooling. Downward motions enhance outflow in the boundary layer (Figure 6b1), driving the propagation of the vortex's squall line outward (Figures 7a1, 7b1, and 7c1). This is also a proposed mechanism for the formation and propagation of TC spiral rainbands (Li & Wang, 2012; Sawada & Iwasaki, 2010). Downdraft-induced outflow delays the development of the vortex (Figures 3a and 6a1). This is consistent with the response of the vortex to latent cooling described in Frisius and Hasselbeck (2009). When precipitation decreases after 25 hr, the low-level inward flux strengthens again (Figure 6b1), resulting in the re-intensification of the vortex after 27 hr (Figures 3a and 6a1).

Downdraft-driven outflow in *Coarse* is insignificant (Figure 6b2), attributed to the presence of weak convection (Figure 6c2) and low rain intensity (Figure 3e). However, widespread warm rain (Figure 5b2) cools down a broad surface area (Figure 8a1) diminishing upward fluxes that draw energy from the sea surface for the vortex. Another source of energy for TC development is the latent heat released from in-cloud microphysical processes. Our results for the evolving TC embryo indicate that condensation below the freezing level contributes the most to latent heating, significantly more than freezing, riming and depositional growth on cloud ice at higher levels (Figures S14a2, S14b2, S14c2, S14d2 in Supporting Information S1). This aligns with the findings of Fan et al. (2018) for tropical convection. Importantly, all the latent heating rates—and thus the total heating rate—are significantly reduced in *Coarse* compared to *MarBG2*, making it the lowest among simulations (Figure 9b; Figures S14a2–S14d2 in Supporting Information S1). This reduction is attributed to the small number of large-sized aerosols added in *Coarse* (Figure 1), which hasten warm rain (Figure 5b2). Fast warm rain reduces the condensational area and limits the transportation of large-sized cloud drops to higher levels, where mixed-phase

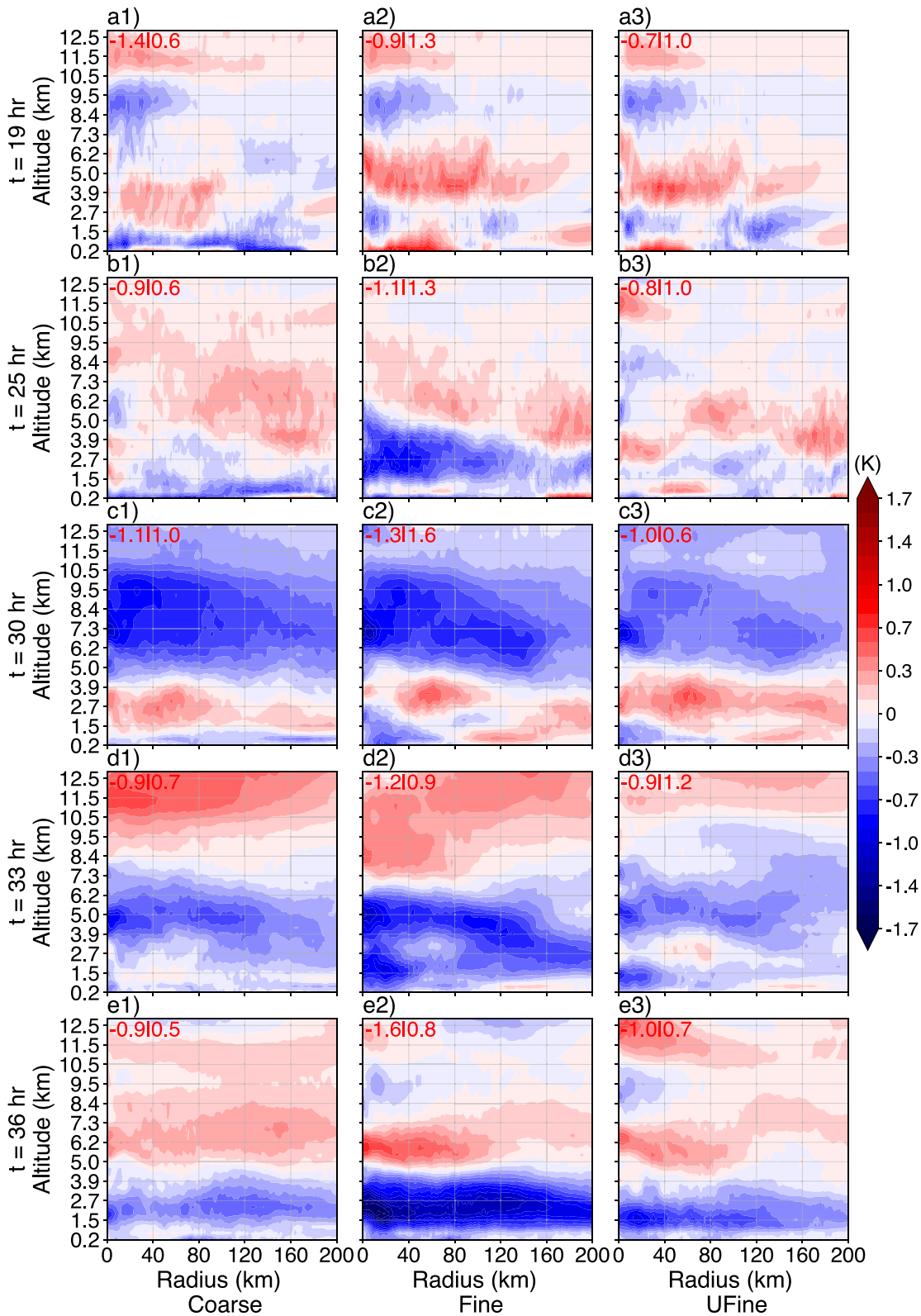


Figure 8. Radius-height composite of differences in azimuthally averaged temperature between each perturbation simulation (i.e., *Coarse*, *Fine*, and *UFine*) and the control run (*MarBG2*) at: (a1–a3) 19 hr; (b1–b3) 25 hr; (c1–c3) 30 hr; (d1–d3) 33 hr; and (e1–e3) 36 hr. The red text on the top left of each panel denotes the minimum and maximum values.

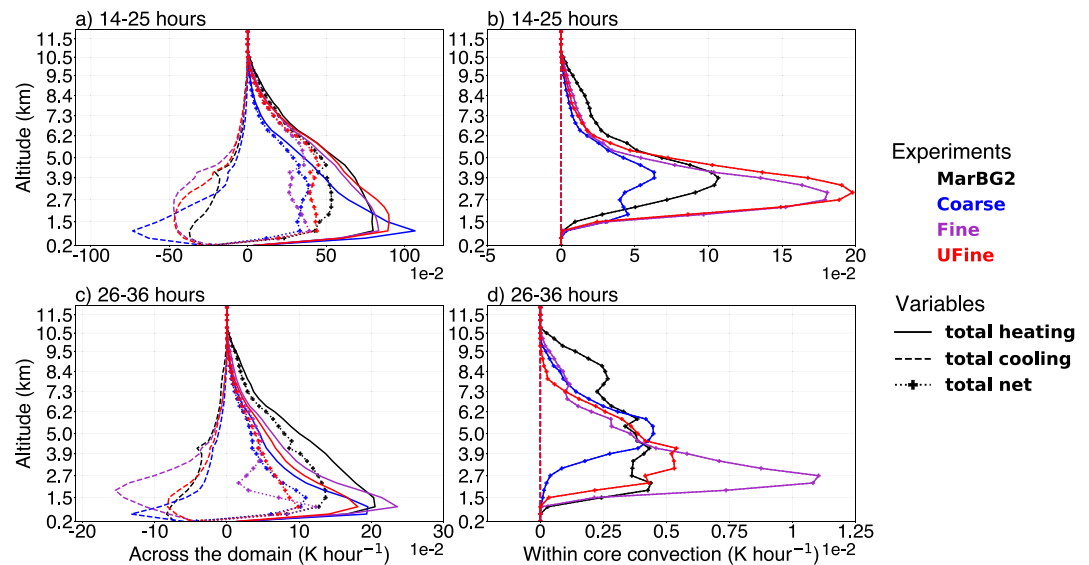


Figure 9. Total diabatic heating and cooling rates from all microphysical processes, along with the net latent heat rates, averaged over the following time periods: (a and b) 14–25 hr and (c and d) 26–36 hr across the simulations. Data are averaged: (a1–d1) over the entire domain; and (a2–d2) over the top 10th percentile of strong updrafts.

processes occur. Consequently, this results in lower efficiencies of condensation on liquid droplets, ice freezing, ice riming and depositional development (Figures S14a2–S14d2 in Supporting Information S1). Furthermore, in accordance with thermodynamic principles, raindrops fall increasing the mass loading at lower levels, resulting in a decrease in buoyancy and vertical velocity (Figures 4e and 6c2). Therefore, the vortex manifests its slow intensification during the first phase with a poor structure (Figures 3b, 7a2, 7b2, and 7c2).

As the vortex evolves in *Coarse*, a cold pool is formed mainly due to widespread warm rain and evaporation of liquid droplets at lower levels (Figure 5b2, Figure S14a1 in Supporting Information S1). However, latent cooling from melting and sublimation are the lowest among simulations due to the low quantity of solid hydrometeors in *Coarse* (Figures S14b1, S14c1 and S15b1, S15c1, Table S1 in Supporting Information S1). Consequently, the cold pool is quickly diminished by upward heat fluxes from the sea surface (Figure 8e1). Therefore, it is unable to degrade the vortex later on and impede the vortex's re-intensification at the end of the simulation as efficiently as in other perturbations (Figures 3b and 6a2–6a4).

In *UFine*, both positive and negative factors influencing the development of the vortex are enhanced as a result of the microphysical changes induced by ultrafine particles. On the one hand, more efficient cloud droplet nucleation in low and middle levels supports both condensational, freezing, riming, and depositional processes releasing a substantial amount of latent heat (Figure 9b; Figures S14a2–S14d2 in Supporting Information S1), which invigorates convection and energizes the vortex. Enhanced convection generates strong updrafts (Figure 4e), which strengthen the vortex secondary circulation. This intensified “in-up-out” vertical circulation leads to rapid spin-up of the near-surface vortex circulation (Figures 3b and 6a4). On the other hand, intense rainfall between 20 and 24 hr occurs simultaneously (Figure 3e) alongside droplet evaporation, ice sublimation, and melting, enhancing latent cooling (Figure 9a; Figures S14a1–S14c1 in Supporting Information S1). The cooler air becomes more dense than its surroundings, supporting strong downdrafts that inhibit convection. These downdrafts also generate low-level outflows (Figure 6b4), accompanied with a low-level cold pool, which is indicated by the negative latent heat net below the cloud base (i.e., approximately 600 m; Figure 9a). The cold pool slows vortex intensification. The vortex attains peak intensity at 25 hr and then weakens (Figures 3b and 6a4).

During the 25–28 hr period in *UFine*, rainfall decreases (Figure 3e); however, strong outflows persist (Figure 6b4), sustaining weak updrafts (Figure 6c4), that are unfavorable for convection. This results in significant reductions in latent heating both within the core convection and across the domain (Figures 9c and 9d). In *UFine*, sublimation and melting are more pronounced than in *Coarse* (Figures S15b1–S15c1 in Supporting Information S1). Evaporation occurs not only in the boundary layer but also at mid-levels (up to 6 km; Figure S15a1 in Supporting Information S1), driven by the presence of small supercooled water droplets transported to these

heights as well as the evaporation of cold rain from melted solid hydrometeors. It is worth noting that intense rainfall persists longer than in the other simulations, lasting up to 27 hr (Figure 3e), which collectively contributes to latent cooling supporting a stronger cold pool in *UFine* compared to *Coarse* (Figures 8d1–8e1 and 8d3–8e3). This cold pool associated with outflows blocks inward warm and moist fluxes (Figure 6b4), which are critical for TC development (Frisius & Hasselbeck, 2009; Wang, 2002), resulting in the reduced re-intensification of the vortex at the end of the simulation (Figures 3b and 6a4).

The most robust and extensive cold pool is found in the *Fine* simulation (Figures 8c2–8e2). Its larger quantity of solid hydrometeors (Table S1) leads to large melting rates compared to *UFine* and *Coarse* (Figure S15c1 in Supporting Information S1). Most importantly, strong evaporation of both freshly melted liquid water and the substantial amount of cloud water below the freezing level significantly enhances latent cooling (Figure S15a1 in Supporting Information S1; Figure 9c), resulting in pronounced cooling of the lower and mid-levels (Figure 8d2). Ultimately, an intense cold pool spreads across the vortex circulation and remains relatively strong for up to 36 hr, as indicated in Figure 8e2. This cold pool, accompanied with intensified outflows near the surface (Figure 6b3), effectively degrades the vortex (Figures 3b and 6a3). Previous studies emphasize the integral role of cold pools in suppressing well-developed TCs (Carrio & Cotton, 2011; Krall & Cotton, 2012). Here, we show that the substantial cold pool created by injecting fine aerosols can significantly slow down tropical cyclogenesis.

4.4. Summary of Underlying Mechanisms for Effects of Different-Sized Aerosols

Comparing the influences of different aerosol sizes and concentrations demonstrates the opposing invigoration effects caused by coarse and fine aerosols. The schematic diagram in Figure 10 summarizes the main mechanisms accounting for the effects found in our simulations.

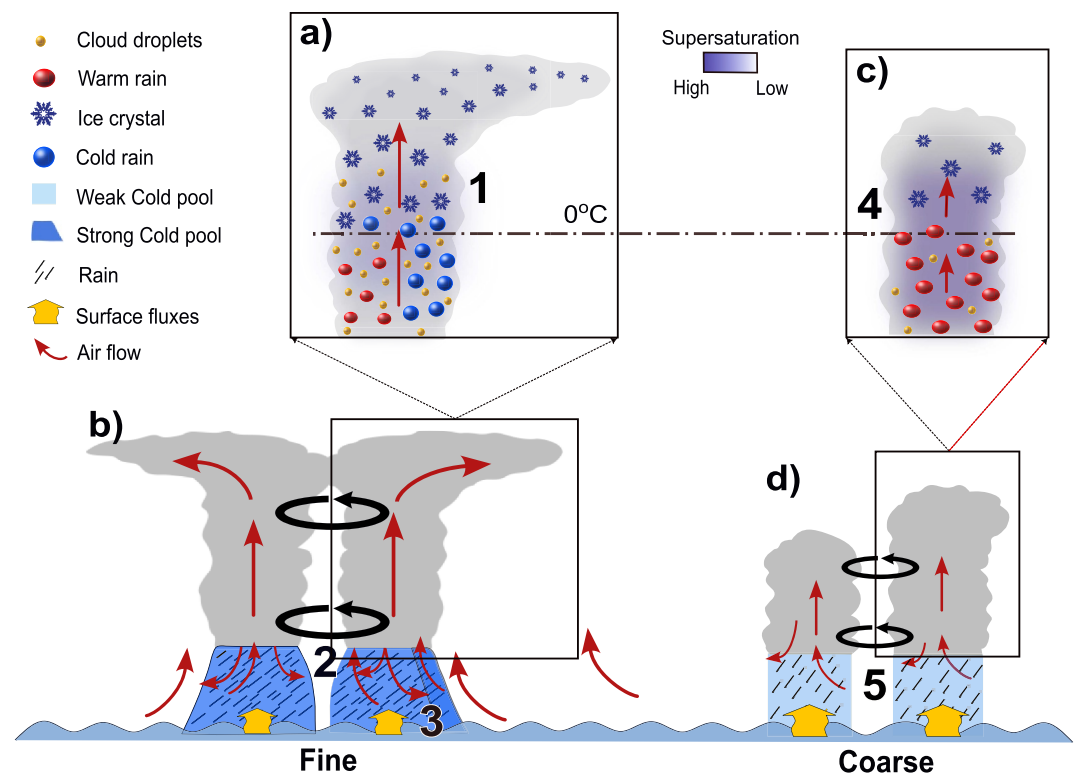


Figure 10. Schematic diagram showing the effects of coarse and fine aerosols on: (a, c) microstructure; and (b, d) dynamics of the vortex. (a) and (c) are a subset of (b) and (d), respectively. (1) Adding fine aerosols reduces supersaturation enhancing condensation and droplet freezing efficiency, convection is invigorated via latent heating. (2) The aerosol-invigorated convection initially enhances the low-level cyclonic circulation. (3) The aerosol-enhanced cold pool expands outward, which enhances the outward propagating squall line and separates the inflow of warm air from the periphery of the vortex leading to its weakening. (4) Adding coarse aerosols has the opposite effect of adding fine and ultrafine aerosols. Warm rain is fastened at the expense of the ice processes leading to a slower acceleration of the vortex initially. (5) The weak generated cold pool is quickly eroded by surface fluxes, so coarse aerosols do not weaken the vortex as much as fine aerosols.

For *fine aerosols*, the addition of a large number of small-sized *CCN* particles substantially reduces supersaturation and enhances condensation efficiency. The small effective radius of cloud droplets inhibits warm rain in exchange for the creation of a large quantity of solid hydrometeors at upper levels. Updrafts are accelerated by the enhanced energy released via latent heating from condensation on liquid water droplets, ice freezing, and ice depositional and riming growth, significantly invigorating the convection (Number 1 in Figure 10a). The low-level vortex circulation is, therefore, energized to accelerate rapidly during the intensifying stage (Number 2 in Figure 10b). Later, a strong downdraft-driven cold pool is contributed by diabatic cooling from intense rain, sublimation, ice-melt, and especially the strong evaporation of liquid droplets below the 0°C isotherm. This cold pool expands outward, enhancing the outward propagating squall line and diminishes both the inward and surface moist and heat fluxes. The resulting outflows in the atmospheric boundary layer lead to the weakening of the vortex (Number 3 in Figure 10b). A recovery occurs when the first squall line has completely propagated outward and is replaced by new rain bands.

In contrast, adding a small number of *coarse aerosols* boosts fast warm rain at the expense of the ice state. A relatively small amount of latent heat is released, as indicated by considerably high supersaturation, and therefore updrafts and resultant convection are not supported, leading to a slow initial acceleration of the vortex (Number 4 in Figure 10c). Warm rain cools down the lower troposphere producing a weak cold pool; however, it is quickly eroded by surface heat fluxes (Number 5 in Figure 10d). Therefore, the vortex is not diminished as effectively as compared to fine aerosols. The vortex then intensifies more significantly at the end of the simulation.

The effects of *ultrafine aerosols* are similar to fine aerosols, but warm rain is not suppressed. The activation of a large number of ultrafine particles and the condensation on newly formed cloud droplets across vertical levels in *UFine* lead to a higher condensation rate, which significantly reduces supersaturation compared to *Fine*. In the early time period following cloud formation, the smaller size of ultrafine particles requires higher supersaturation for activation, leading to a lower droplet number concentration relative to *Fine*. At this stage, owing to less competition for water vapor, cloud droplets in *UFine* have a larger effective radius compared to *Fine*. This allows them to effectively coalesce and generate warm rain. The formation of warm rain reduces condensational surface area, which facilitates the secondary activation of ultrafine particles at higher altitudes. As a consequence, both the vortex intensity and rainfall are increased during the intensifying stage of the vortex in *UFine*. Intense rainfall contributes significantly to cooling in addition to evaporation, sublimation, and melting, maintaining a strong cold pool, which effectively weakens vortex circulation at a later on stage. As the cold pool gradually warms by converging surface heat fluxes, the vortex slowly re-intensifies.

4.5. Practical Assessment Based on the Required Aerosol Mass

Based on the results, fine, ultrafine and coarse aerosol perturbations all offer theoretical promise for intervention in TC initiation, quantitatively validating the conceptual approaches suggested by Tran, Prinsley, et al. (2024). However, the use of coarse aerosols may be challenging due to the limited technologies currently available for their mass production and seeding into a developing TC vortex. Using fine aerosols, or a combination of fine and ultrafine aerosols, may be more feasible due to their smaller mass and ease of production and application. One may question whether the aerosol settings in this study are practical for interventions. In reality, it is unnecessary to saturate the whole domain with uniform aerosol concentrations as currently configured. For practical applications, we envision a more targeted approach where seeding is implemented along a circular path centered on the vortex, positioned beyond the vortex's cloud bands within the boundary layer. Low-level convergent flow would then transport these aerosols toward the cloud bands, where updrafts would lift them to microphysically and dynamically affect the TC embryo, eventually mitigating its intensity. The convective clouds that ingest the boundary layer air containing the seeded aerosols will receive nearly the same amount of aerosols as they would if the entire domain were saturated.

In that context, appropriate technology could be used to disperse hygroscopic aerosol plumes (e.g., small ammonium sulfate particles, which are lighter than sea salts) at a 200-km distance from the vortex center (outside the cloud bands; Figure 7) at 300-m height. We assume an average inflow of 5 m s^{-1} (consistent with the high values in Figures 6b1–6b4), which, along with updrafts in clouds, is capable of filling the 0–600 m lower layer, feeding aerosols to all convective clouds within this region. The required seeding mass flux using the particle size distributions studied (Figure 1) is computed and presented in Table 3. Clearly, the *UFine* configuration, having the smallest particle size, requires the lowest seeding mass flux at approximately 79 tons hr^{-1} . After subtracting

Table 3
Mass Flux Required for Targeted Seedings Using the Aerosol Size Distributions Studied

Experiment	Mass flux required for targeted seedings	
	(kg s ⁻¹)	(tons hr ⁻¹)
<i>MarBG2</i> (to be subtracted from the seeded aerosols)	20.83	74.99
<i>Coarse</i>	16,674.77	60,029.17
<i>Fine</i>	124.05	446.59
<i>UFine</i>	21.83	78.58

the aerosol amount in the natural condition (*MarBG2*), this is reduced to only 4 tons hr⁻¹. This amount can be feasibly achieved by operating several aircraft over several hours to disperse aerosols, with the goal of inhibiting the spin-up of the TC embryo. For targeted intervention using fine aerosols, the required seeding mass flux is around 372 tons hr⁻¹ (above the background aerosols mass flux)—a quantity transportable by four to ten large cargo aircraft each with 100+ tons payloads, depending on the duration of the operation.

5. Conclusion and Discussion

This study investigates, for the first time, the impacts of aerosols of varied sizes and concentrations—coarse [$1 < D \leq 4 \mu\text{m}$, 617 cm^{-3}], fine [$0.05 < D \leq 1 \mu\text{m}$, 3585 cm^{-3}], ultrafine [$D < 0.05 \mu\text{m}$, $5,596 \text{ cm}^{-3}$ —on the formation of a TC. A simulation of an idealized vortex, with intensity never exceeding the tropical depression strength ($V_{\text{max}} < 34 \text{ m s}^{-1}$), is configured to isolate the effects of aerosols from the large-scale environment, and from strong TC dynamics as well as ocean feedbacks (as seen in mature TCs).

We found that aerosols of different sizes and concentrations uniquely affect the vortex's microstructure, dynamics, energy cascades, and destructive behaviors. Coarse aerosols enhance fast warm rain and impede condensational growth on liquid droplets and ice-related processes, which initially inhibits the rapid intensification of the vortex. However, the short-lived cold pool, caused mostly by widespread weak rainfall and evaporation in the lower levels, is unable to weaken the vortex in the later stages as much as for fine and ultrafine aerosol experimental configurations.

In contrast, both fine and ultrafine perturbations initially strengthen the vortex but degrade its intensity more effectively in the later stages. This is because of the influence of a strong cold pool that develops as a result of the combined effects of intense rainfall and the total latent cooling from microphysical processes such as evaporation, sublimation, and melting. In *UFine*, the role of intense rainfall is dominant, driven by the most intense rain rates, whereas in *Fine*, latent cooling from these microphysical processes plays a more significant role due to the large quantities of solid hydrometeors—which are greater than in the *Coarse* and *UFine* simulations. Such a scenario, where the generated cold pool propagates outward overwhelming the inflow, can only occur in the early stages of the TC development, when the vortex is spinning up and the radial winds are sufficiently weak.

Fine aerosol perturbation was found to be the most effective in reducing the total energy, total destructive potential and rainfall, by approximately 17%, 25%, and 22%, respectively, when considering the entire integration period (i.e., ca. 36 hr) for this weak, developing vortex. This mitigation in total TC energy, total destructive potential, and rainfall was less effective when using coarse aerosols with reductions of 16%, 21%, and 16% in total energy, total destructive potential and rainfall, respectively. Smaller reductions were found for ultrafine aerosol perturbation, with reductions in total energy of 11%, total destructive potential of 20%, and rainfall of 8%.

For practical solutions, targeted seeding strategies using ultrafine and/or fine aerosols hold the most promise, as they are easier to produce and their application is more feasible from a logistics perspective. Despite fine aerosols being the most effective (viz., they lead to the greatest reductions in total energy, total destructive potential, and rainfall), there is a small tradeoff between effectiveness and feasibility; ultrafine aerosols have less impact than fine aerosols but having the smallest particle size means they require the lowest seeding mass flux. Depending upon the aerosol size used, several to 10 large aircraft might be deployed over a few hours to disperse aerosol along the circular path at a distance of 200 km from the vortex center, at 300 m altitude to fill up the 600 m low-level layer. Validating these proposed strategies through numerical simulations would be an interesting topic for future research.

It is also important to note that although the vortex may be weakened after initial invigoration by fine and ultrafine aerosol injections, there could still be re-intensification later on. This raises questions about how the vortex behaves after 36 hr, which will depend on the vortex intensifying to overwhelm the outward flow component of the cold pool and would be a bifurcation point. However, studying these is beyond the scope of the current study. We also emphasize that these results are based on a single idealized case that neglects the effects of the large-scale environment, including varying meteorological conditions and dust effects. These environmental factors could influence aerosol characteristics, cyclogenesis, and aerosol-cyclogenesis interactions. Therefore, further research that investigates TC-aerosol interactions at different stages of the TC lifecycle incorporating relevant environmental conditions is necessary. Additionally, since the findings are based on experiments with sea salt aerosol particles, it would be valuable to explore how these interactions might differ with aerosols of varying compositions. In future work, aerosol scavenging by precipitation could be considered to enhance the representation of aerosol sinks in the SBM model.

Despite these limitations, the findings in this study establish the quantitative sensitivity of a TC to aerosol size and concentration during its early stages. In this way, this study serves as a catalyst for future investigations aiming to identify effective intervening strategies for TCs at both initial and mature stages. In our ongoing work, we will examine the sensitivity of various seeding strategies under the dynamic conditions of a fully developed TC.

Data Availability Statement

The data and the WRF-FSBM code on which this article is based are available in Tran, Fan, et al. (2024).

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