AGU ADVANCING EARTH AND SPACE SCIENCES

JGR Atmospheres

REVIEW ARTICLE
10.1029/2024JD041931

Unveiling Aerosol Impacts on Deep Convective Clouds: Scientific Concept, Modeling, Observational Analysis, and Future Direction

Jiwen Fan[1], Yuwei Zhang[2], Zhanqing Li[3], Hongru Yan[4], Thara Prabhakaran[5], Daniel Rosenfeld[6], and Alexander Khain[6]

[1]Environnental Science Division, Argonne National Laboratory, Lemont, IL, USA, [2]Atmospheric, Climate, and Earth Sciences Division, Pacific Northwest National Laboratory, Richland, WA, USA, [3]Department of Atmospheric and Oceanic Sciences, University of Maryland, Baltimore, MD, USA, [4]Key Laboratory for Semi-Arid Climate Change of the Ministry of Education, College of Atmospheric Sciences, Lanzhou University, Lanzhou, China, [5]Indian Institute of Tropical Meteorology, Ministry of Earth Sciences, Pune, India, [6]Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel

Key Points:

- We clarify important conditions for condensational and freezing-induced invigoration to be significant
- We provide recommendations for methodologies of analyzing and simulating aerosol impact on convection based on new findings
- Feedback to circulation and meteorology can be an important part of aerosol impacts on deep convective clouds and a direction of future studies

Correspondence to:

J. Fan,
fanj@anl.gov

Citation:

Fan, J., Zhang, Y., Li, Z., Yan, H., Prabhakaran, T., Rosenfeld, D., & Khain, A. (2025). Unveiling aerosol impacts on deep convective clouds: Scientific concept, modeling, observational analysis, and future direction. *Journal of Geophysical Research: Atmospheres*, *130*, e2024JD041931. https://doi.org/10.1029/2024JD041931

Received 8 JUL 2024
Accepted 2 JUL 2025
Corrected 29 AUG 2025

This article was corrected on 29 AUG 2025. See the end of the full text for details.

Author Contributions:

Conceptualization: Jiwen Fan
Data curation: Thara Prabhakaran, Daniel Rosenfeld
Formal analysis: Yuwei Zhang, Hongru Yan
Funding acquisition: Jiwen Fan
Investigation: Jiwen Fan, Yuwei Zhang, Zhanqing Li, Hongru Yan, Alexander Khain
Methodology: Jiwen Fan, Daniel Rosenfeld
Project administration: Jiwen Fan
Resources: Jiwen Fan
Supervision: Jiwen Fan

Abstract

Aerosols are important environmental factors that can influence deep convective clouds (DCCs) by serving as cloud condensation nuclei. Due to complications in DCC dynamics and microphysics, and aerosol size distribution and composition, understanding aerosol-DCC interactions has been a daunting challenge. Recently, the convective invigoration mechanisms through enhancing latent heating in condensation and ice-related processes that have been proposed in literature are debated for their significance qualitatively and quantitatively. A salient issue arising from these debates is the imperative need to clarify essential knowledge and methodologies in investigating aerosol impacts on deep convection. Here we have presented our view of key aspects on investigating and understanding these invigoration mechanisms as well as the aerosol and meteorological conditions under which these mechanisms may be significant based on new findings. For example, the condensational invigoration is most significant under a clean condition with an introduction of a large number of ultrafine particles, and the freezing-induced invigoration can be significant in a clean condition with a large number of relatively large-size particles being added. We have made practical recommendations on approaches for investigating aerosol impacts on convection with both modeling and observations. We note that the feedback induced by the invigoration via the enhanced latent heating to circulation and meteorology can be an important part of aerosol impacts but is very complicated and varies with different convective storm types. This is an important future direction for studying aerosol-DCC interactions.

Plain Language Summary

Deep convective clouds (DCCs) play a crucial role in Earth's energy and water cycles, significantly impacting human lives and natural systems by generating hazardous weather conditions. Aerosols are recognized as important environmental factors that can influence DCCs. However, understanding aerosol-DCC interactions qualitatively and quantitatively remains a significant challenge, leading to substantial uncertainty. Early studies have proposed some mechanisms regarding the impacts of aerosols on convective storms, focused on the invigoration of convection by cloud condensation nuclei. This topic has recently garnered considerable attention, with ongoing scientific debates. A salient issue arising from these debates is the imperative need to clarify essential knowledge in investigating aerosol impacts on deep convection. Thus, we provide clarity on this subject by identifying key aspects for proper investigation with model simulations and observational analysis and reconciling recent seemingly contradictory findings with new analyses, aiming to promote future advancements in the field of aerosol-cloud interactions.

1. Background

Aerosol interactions with deep convective clouds (DCCs) are complicated because of the complex microphysics, dynamics, and meteorological conditions of these clouds. Convective cloud invigoration can be reflected in various metrics such as increased updraft velocity, heavy rain rates, cloud top height, and lightning, and it was loosely defined in literature. To avoid confusion, here we define "convective invigoration" as enhanced updraft velocity or vertical mass flux in convective cores. Several mechanisms for convective invigoration by aerosols via serving as cloud condensation nuclei (CCN) have been proposed (Fan & Li, 2022 and references therein): (a) enhanced condensational heating due to more efficient condensation (so-called "warm-phase invigoration"), (b)

Validation: Jiwen Fan
Visualization: Yuwei Zhang, Hongru Yan, Daniel Rosenfeld
Writing – original draft: Jiwen Fan
Writing – review & editing: Jiwen Fan, Yuwei Zhang, Zhanqing Li, Hongru Yan, Thara Prabhakaran, Daniel Rosenfeld, Alexander Khain

enhanced latent heating from freezing-induced intensification of ice processes including droplet freezing, riming, and deposition (so-called "cold-phase invigoration"), (c) changed cold pools and secondary convection in the outflows, and (d) enhanced convergence/divergence and moistening of the middle and upper troposphere. It has been understood that aerosol effects on convection and precipitation depend on meteorological conditions such as cloud base height, relative humidity (RH), vertical wind shear, and convective available potential energy (CAPE) (e.g., Fan et al., 2009; Fan, Rosenfeld, et al., 2012; Khain, 2009; Khain et al., 2008; Lebo & Morrison, 2014; Storer & van den Heever, 2013), as reviewed in Tao et al. (2012) and Fan et al. (2016). Thus, it is not surprising that various aerosol effects are found in different DCC cases. In modeling studies, the aerosol effect also depends on different approaches emloyed in aerosol and cloud microphysical parameterizations (Fan, Leung, et al., 2012; Khain et al., 2015; Marinescu et al., 2021; Y. Zhang et al., 2021).

Recently, much attention has been paid to aerosol impacts on convection intensity. Debates have been focused on the effect of latent heating increase as described in warm and cold phase invigoration and a detailed review of studies about two mechanisms is provided by Varble et al. (2023). Regarding the warm phase invigoration, the mechanism was founded on a solid theoretical basis (Fan & Khain, 2021; Khain & Pinsky, 2018; Kogan & Martin, 1994). Numerous modeling studies over the past couple of decades showed larger updraft velocities in polluted DCCs compared to clean DCCs due to enhanced condensation by the increased CCN (e.g., Cotton & Walko, 2021; Fan et al., 2007, 2018; Khain et al., 2012; Lebo, 2018; Sheffield et al., 2015; C. Wang, 2005). How high the maximum supersaturation can reach in updrafts under clean conditions to a large extent measures the magnitude of enhancement on the updraft velocity (Fan et al., 2018; Romps et al., 2023). High supersaturations such as 10% or larger are possible based on theoretical calculations for a typical cumulus cloud condition (Shaw, 2000) and for air parcels (Khain & Pinsky, 2018) as well as in modeling studies (Fan et al., 2018; Grabowski & Morrison, 2016, 2020; Phillips et al., 2007). Varble et al. (2023) also emphasized the significance of this mechanism depends on the substantial supersaturation changes in convective clouds and claimed that such large supersaturation changes have not been observed. However, our current ability of observing supersaturation in strong updrafts of DCCs is limited.

Khain et al. (2012) and Fan et al. (2018) have suggested that the warm phase invigoration was manifested by ultrafine aerosol particles (UAPs) in humid and pristine tropical regions. The detailed physical modeling of Fan et al. (2018) presented that rapid rain formation due to very low droplet number concentrations around cloud bases under pristine conditions reduces droplets for condensation, leaving very high supersaturation conditions in the updrafts, which activate a large number of UAPs from urban pollution and enhance condensation substantially. A recent study showed that UAPs from biomass-burning plumes may also invigorate the scattered convective clouds and precipitation in the Amazon dry season (Shrivastava et al., 2024). The large effects of UAPs have been suggested in other regions like Houston and southeast China (Chen et al., 2017, 2020; Fan et al., 2020).

The core of the recent major debate on the warm-phase invigoration mechanism is whether high supersaturation exists in the updrafts of DCCs under clean aerosol conditions in the real atmosphere. Romps et al. (2023) examined the mean values of the observed quasi-steady state supersaturation of updrafts in shallow cumuli with cloud tops of 4 km in the dry season and claimed that the invigoration in DCCs found in the wet season in Fan et al. (2018) does not exist. In fact, the findings of these two studies are not comparable due to different cloud types; low supersaturation in shallow cumuli shown in Romps et al. (2023) is consistent with literature (e.g., Siebert & Shaw, 2017) and cannot disapprove the existence of high supersaturation and invigoration for DCCs in Fan et al. (2018). The same claim was made by Öktem et al. (2023) by reanalyzing the 17 observed cases from the observational part of Fan et al. (2018) and also focusing on the low-levels (warm-phase of DCCs). The issues with the approaches will be discussed in Section 4.

Regarding the cold-phase invigoration, the effect is through the increase in latent heat at high levels from freezing extra cloud liquid transported from low levels under polluted conditions, because of the suppression and delay of warm rain as a result of much higher cloud droplet numbers (Andreae et al., 2004; Khain et al., 2005; Rosenfeld et al., 2008). The debate is mainly about whether the effect of latent heating from the freezing of extra liquid water is completely offset by the extra-loading effect of the liquid water being carried across the freezing level or not. Rosenfeld et al. (2008) argued that the latent heat of freezing balances the condensate load and showed that the net buoyancy occurs when the condensates offload quickly as large ice hydrometeors (graupel/hail) sediment. Many cloud modeling studies showed that the increased condensate loading effect is smaller than the latent heating effect induced by extra droplet freezing (i.e., freezing, deposition, and riming), leading to a net increase in buoyancy (Chen et al., 2020; Fan et al., 2018; Fan, Rosenfeld, et al., 2012; Khain, 2009; Khain et al., 2008; Lebo et al., 2012; Storer & van den Heever, 2013; Tao & Li, 2016). However, other modeling studies have argued that the condensate loading effect is large enough to completely offset the latent heating effect (Grabowski & Morrison, 2016, 2020). It is interesting that those studies showing positive aerosol effects used bin or bin-emulating microphysics schemes and also conducted real-case simulations while those showing small or negative aerosol impacts often used bulk microphysics schemes and conducted idealized model simulations. Recent parcel model studies also showed a strong offset effect from condensate loading or entrainment is possible theoretically (Igel & van den Heever, 2021; Peters et al., 2023), but it does not mean the entrainment and the condensate loading processes will dominate in reality in every case as parcel models do not simulate real 4-D DCCs (3-D space plus time evolution). Varble et al. (2023) acknowledged the complexity involved in this mechanism but did not present the whole picture that we intend to do in this study.

Concerning the ongoing scientific debate, a salient issue arising from these debates is the imperative need to clarify (a) essential knowledge and whole picture about the mechanisms and (b) the uncertainties and appropriate approaches for investigating aerosol impacts on DCCs from both modeling and observational aspects. This work fills in this gap. We offer practical suggestions on how to appropriately simulate aerosol impacts on DCCs and analyze observational data that are lacking in Varble et al. (2023). Each recommendation is grounded in scientific investigations with new results and justifications, as detailed in Sections 2–4. Our work also presents new analysis and discussion on (a) high supersaturation and updraft velocity, (b) the relationship of supersaturation and quasi-steady-state supersaturation at different droplet concentrations, and (c) the problems and uncertainty of the previous studies. In the following sections, we will present key aspects for proper investigations with model simulaitons and observational analysis, identify challenges, and reconcile recent seemingly contradictory findings, aiming to promote future advancements on this topic.

2. Key Aspects of the Invigoration Mechanisms

2.1. Condensational Invigoration

We adopt the terminology of “condensational invigoration” from Cotton and Walko (2021) to replace the terminology of “warm-phase invigoration” to avoid literal confusion. Recent studies (Öktem et al., 2023; Romps et al., 2023) have shown a misinterpretation of this mechanism based on the literal meaning of “warm-phase,” which have led them to either study shallow warm clouds (Romps et al., 2023) or focus on convection at the warm-cloud part of DCCs (Öktem et al., 2023), claiming that the vertical velocity increase at high levels is due to other reasons such as freezing-induced invigoration.

Since condensational invigoration operates through enhanced vapor condensation, high supersaturation within a storm prior to the introduction of high aerosols is a prerequisite for a significant effect. Moreover, aerosol size distribution plays a critical role in determining its impact. In environments with high concentrations of large-size particles (accumulation and coarse modes), strong nucleation near cloud bases generates a large number of small droplets. These droplets compete for water vapor, suppressing the conversion of cloud droplets into rain and limiting the formation of high supersaturation within the clouds. In contrast, in environments where large-particle concentrations are low but ultrafine aerosol particle (UAP) concentrations are high (e.g., as observed in the Amazon in Fan et al. (2018)), droplet concentrations near cloud bases remain low because UAPs are not readily activated under the typically low supersaturation at cloud bases. This facilitates the conversion of droplets into rain, reduces overall condensation due to fewer droplets, and promotes the development of higher supersaturation. The higher supersaturation subsequently activates a large number of UAPs, enhancing condensation efficiency and bringing down supersaturation. Therefore, the significance is essentially concerned with the secondary activation of a large number of small particles that are often observed with anthropogenic emissions in urban environments (Q.-Q. Li et al., 2023). UAPs are low or even mostly lacking in natural environments without much anthropogenic influence (Meinrat, 2007; Pöschl et al., 2010). UAPs require high supersaturations to be activated. For example, the critical supersaturation to activate ammonia sulfate particles of 20 nm (in dry diameter) is ∼2% (Seinfeld & Pandis, 1997). Besides aerosol sizes, the critical supersaturation for aerosol activation also depends on aerosol hygroscopicity (a measure of solubility), for example, particles of 100 nm with a hygroscopicity of 0.01 (common for organic aerosols) need a supersaturation larger than 1% to activate (Petters & Kreidenweis, 2007).

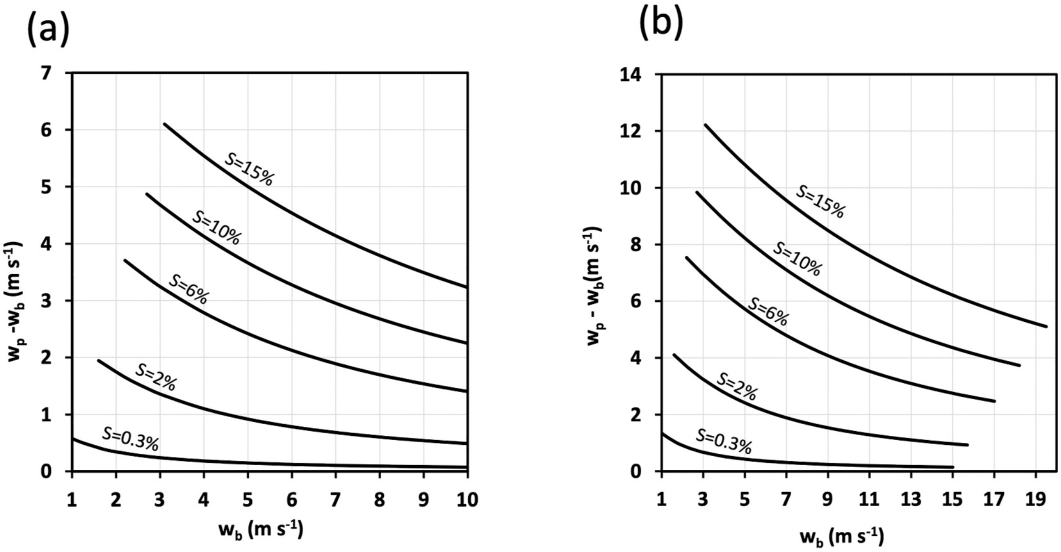

Figure 1. Relationship of updraft velocity enhancement with the baseline vertical velocity (w_b, x-axis) and supersaturation (S, solid lines) based on Equation 1 for (a) the lower atmosphere around 1.5–2 km, using $ar/c_d = 500$ m^2 s^{-2} from Romps et al. (2023) and (b) the middle atmosphere (8 km), using $ar/c_d = 1{,}500$ m^2 s^{-2} which is obtained with $r = 1$ km for deep convective clouds, $c_d = 0.2$ (Peters et al., 2022), and $a = 0.3$ m s^{-2}. Note these values are used just for a rough demonstration of the concept.

When UAPs are ample and able to activate inside convective updrafts, the significance of condensational invigoration should depend on how high the supersaturation (S_b) and updraft velocity (w_b) are in baseline (clean) conditions. Romps et al. (2023) provided a crude estimate of the updraft speed of a polluted updraft (w_p) given a baseline updraft speed (w_b) and supersaturation (S_b, Equation 1), assuming that the supersaturation is removed by enhanced condensation under polluted conditions and the buoyancy is balanced by the drag locally.

$$w_p = \sqrt{w_b^2 + \frac{ar}{c_d} S_b}, \tag{1}$$

where r is the radius of a convective updraft, c_d is the drag coefficient for a cloudy updraft, and a is a velocity acceleration parameter [Equation A8 in Romps et al. (2023)]. Figure 1a shows the updraft velocity enhancement ($w_p - w_b$) in the lower atmosphere, using Equation 1 and the values of r, c_d, and a taken from Romps et al. (2023), with $ar/c_d = 500$ m^2 s^{-2}. The updraft velocity enhancement increases with S_b but decreases with w_b. There is a ~3 m s^{-1} enhancement for w_b of 4 m s^{-1} with 6% supersaturation in the lower atmosphere.

Figure 1b shows the potential updraft velocity enhancement at a high level (~8 km), calculated based on Equation A8 in Romps et al. (2023), with $r = 1$ km, $c_d = 0.2$ (Peters et al., 2022), and $a = 0.3$ m s^{-2} for the 8-km level of a typical tropical atmosphere. For DCCs at 8 km, the radius of a convective updraft can vary a lot, and we select $r = 1$ km based on the simulation with 0.25 km resolution in Fan et al. (2017). The c_d of 0.2 is used based on Peters et al. (2022), and a of 0.3 m s^{-2} is calculated based on Equation A8 in Romps et al. (2023) using the temperature and the saturation vapor mass fraction at ~8 km. Compared to Figure 1a for the lower atmosphere, the updraft velocity enhancement at the higher level is more significant for the same w_b and S, which is mainly due to the large updraft size (r) as a is decreased due to the decrease in temperature and saturation vapor mass fraction, and c_d is assumed not changed. For example, the enhancement is ~6 m s^{-1} for w_b of 4 m s^{-1} with S of 6% at 8 km altitude, in constrast to the ~3 m s^{-1} enhancement at the lower atmsophere for the same w_b and S. The dependence on w_b is smaller compared to that in the lower atmosphere.

Note the buoyancy-drag balance framework in Romps et al. (2023) means that the instantaneous updraft speed is related to the instantaneous buoyancy and does not consider the vertical continuity of convective flow (i.e., treated as a point event). This leads to their argument that the updraft response to the buoyancy is local thus the updraft speed enhancement by condensational heating occurs at the low levels (Öktem et al., 2023). This deviates from

10.1029/2024JD041931

the theory of convective storms, which, fundamentally all arise from buoyancy, that is, their air motions always originate in the form of vertical accelerations (Chapter 7 of Houze (2014)). Because the buoyant accelerations lead to strong updraft velocity, the mass field must adjust, resulting in the buoyancy pressure-perturbation gradient (PPG). The rapidly moving air in convective clouds also leads to entrainment of surrounding environmental air. Finally, the air motions develop rotation, which can enhance entrainment and produce dynamic PPG (Houze, 2014). The above description is basically the continuity equation of convective clouds (Equation 7.2 in Houze (2014)). Therefore, the updraft speed should peak at the height of neutral buoyancy (the basis for some convective cloud parameterizations such as G. J. Zhang and McFarlane (1995)) if other factors (e.g., PPG and entrainment) are not considered. That is why overshoots above the neutral buoyancy level can possibly happen, whereas the buoyancy-drag balance framework cannot explain overshoots.

The buoyant acceleration described above for convective clouds is based on Newton's Second Laws of Motion, that is, with positive buoyancy (the force) exerted on the rising air parcel (the object), the velocity acceleration occurs along the trajectory of the object. This is also the theoretical support of calculating maximum w based on CAPE from a tephigram. Therefore, it may be fine to gain a rough idea of the updraft speed of a polluted updraft based on a baseline velocity and supersaturation using the buoyancy-drag balance framework in Romps et al. (2023). However, it is problematic to apply it to the reality and focus on the responses of w at the low levels only (i.e., a local point) as in Öktem et al. (2023). In Amazon, with the intrusion of middle-level dry air, entrainment affect where w peaks and in the convective cores (reflectivity >35 dBZ), w increases with height and peaks above 7 km (Figures 5d and 5e in Giangrande et al. (2023)). The real-case modeling with detailed spectral-bin microphysics for an Amazon wet season case shows that the largest response of w in the convective cores occurs at the upper levels when only the latent heating at low levels is changed (Fan et al., 2018). Also, condensation can occur significantly in the supercooled and mixed-phase parts of DCCs as shown in Fan et al. (2018, 2020). Fan et al. (2018) showed that the large increase in the condensational heating is at 3–8 km altitudes becasue of activation of ultrafine particles under higher supersaturation (peaking at the height of ∼7 km). Therefore, the condensational invigoration in DCCs (i.e., so-called "warm-phase invigoration") is very different from the warm-cloud invigoration for shallow convection (Koren et al., 2014). In addition, w at low levels of DCCs is usually small and complicated by other significant dynamic processes, such as entrainment, warm rain precipitating, cold pool, etc. Therefore, it can be difficult to see the signal at the low levels from observations. In summary, we argue that aerosol effects on updraft velocity via condensational invigoration should be examined over vertical profiles, and the significant effects likely appear in the middle and high levels.

It is a daunting challenge to study whether high supersaturation exists in DCC updrafts under clean conditions in a real atmosphere because such measurements in convective cores are very difficult to obtain and are thus lacking or very limited. Here, we showed the in situ aircraft observations for deep cumuli during the 2018 Cloud Aerosol Interaction and Precipitation Enhancement Experiment (CAIPEEX IV) carried out during the monsoon season in India (Prabhakaran et al., 2023). Since supersaturation cannot be measured directly, the quasi-steady supersaturation (S_{qss}) is derived based on observations. Figure 2 shows the results from three flights (04, 18, and 20 July 2018) that sampled some convective cores at the developing stage of deep cumuli during the Indian summer monsoon season (19 cloud passes in total) as described in Bera et al. (2021). The three flights sampled clouds developed in the intermediate aerosol conditions (with the maximum droplet numbers of ∼600 cm^{-3} at cloud bases). The data with raindrops were filtered out and there was no data with ice (the freezing level is at ∼5 km altitude).

Typically, the uncertainty for temperature is ±0.25°C from −50°C to +50°C, for dew point temperature is ±0.2°C from −40°C to +60°C, and for RH is ±3% (see Supplemental Material in Prabhakaran et al. (2023)). Near the cloud edges, the measurement uncertainties for the temperature and humidity can be very large (>100%) due to the wetting of temperature probes and evaporation. In this analysis, we tried to exclude the samples at cloud edges by applying the thresholds of $w > 2$ m s^{-1}, droplet number concentrations (N_c) > 1 cm^{-3}, and liquid water content (LWC) of >0.005 g m^{-3}. We further removed a single point at ∼2.1 km altitude having S_{qss} of 11.2%, w of 8.31 m s^{-1}, LWC of 0.0069 g m^{-3}, and N_c of 1 cm^{-3} since such a strong updraft with high supersaturation but low LWC at this low altitude is suspicious.

Among these samples (867 in total), there are about 3% of samples having quasi-steady state supersaturation (S_{qss}) of >5%. These high supersaturations (>5%) are associated with low N_c (Figure 2a) and they can appear at any updraft speed and any height within the selected data sets (Figures 2a and 2b). The high supersaturations at the

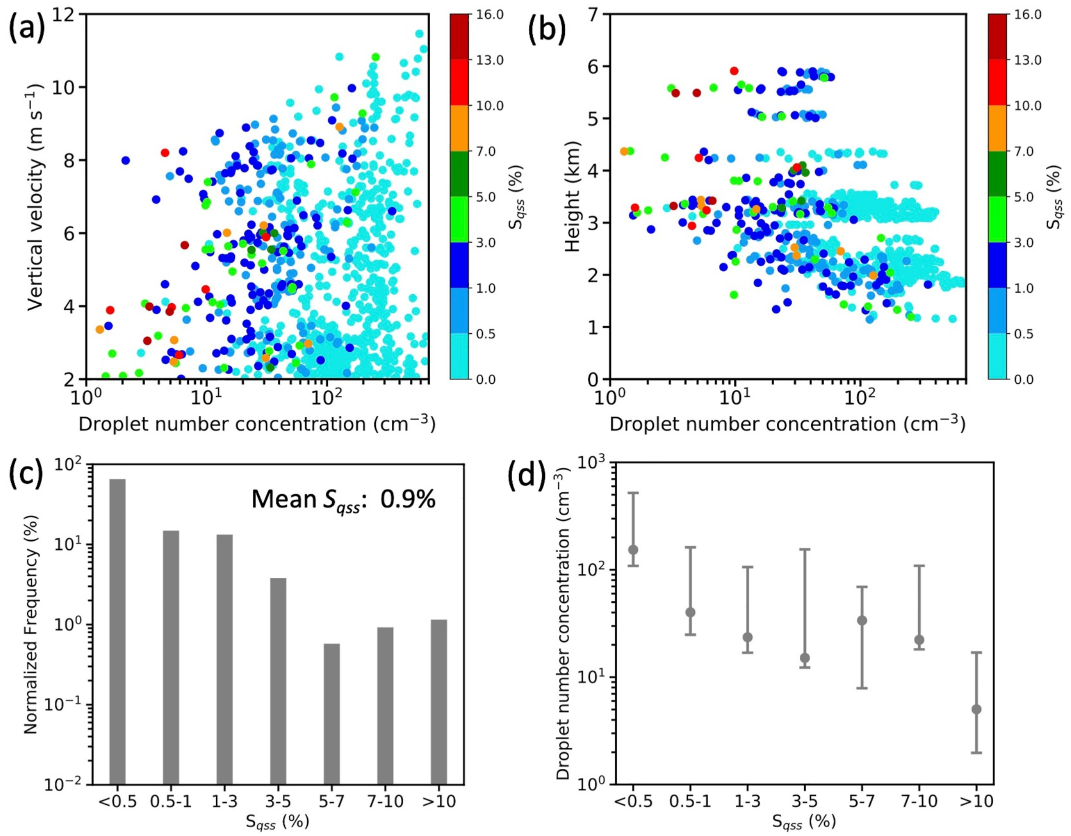

Figure 2. Relationships of (a) S_{qss} with w (y-axis) and N_c (x-axis), (b) the vertical profile of S_{qss} as a function of N_c, (c) normalized frequency of S_{qss}, and (d) median and 10th to 90th percentiles of N_c for each S_{qss} bin for three moderately clean deep cumuli cases (04, 18, and 20 July 2018) from CAIPEEX IV. The data shown here are filtered with $w > 2$ m s^{-1}, $N_c > 1$ cm^{-3}, and liquid water content (LWC) of >0.005 g m^{-3} and removal of a suspicious data point at ~2.1 km altitude that has S_{qss} of 11.2%, w of 8.31 m/s, and LWC of 0.0069 g m^{-3}. The total data points are 867.

low levels (2–3 km altitude; Figure 2b) are associated with large updraft velocity, which can be caused by gust fronts (Houze, 2004 and references therein). There are ~1% of the supersaturations with a value ≥10% (Figure 2c). The mean N_c for $S_{\mathrm{qss}} \geq 10\%$ is only several cm^{-3} (Figure 2d). The maximum S_{qss} is 16.7% for the 04 July case ($w = 4.0$ m s^{-1}, $N_c = 3.3$ cm^{-3}), 14.4% for the 18 July 2018 case ($w = 3.1$ m s^{-1}, $N_c = 3.3$ cm^{-3}), and 12.7% for the 20 July 2018 case ($w = 3.9$ m s^{-1}, $N_c = 1.6$ cm^{-3}). High supersaturations occur when N_c is low, and in general, supersaturation increases with the decrease of N_c (Figure 2d), which is defined by the theoretical derivation (Khain & Pinsky, 2018). The model simulation of a case from the Observations and Modeling of the Green Ocean Amazon (GoAmazon) field campaign - the C_BG simulation as described by Fan et al. (2018) also shows a similar relationship of supersaturation with N_c (Figure 3b). Note that the observed supersaturations vary a lot based on the frequency distribution with a mean value of only 0.9% (Figure 2c). A similar variability and a low mean supersaturation (1.4%) are also seen in the model simulation of the Amazon case (Figure 3a).

2.2. Freezing-Induced Invigoration

Regarding "cold-phase invigoration," we adopt the term "freezing-induced invigoration" to be more straightforward. This mechanism mainly concerns the relatively large-size aerosol particles that are activated at cloud bases at low supersaturation levels because warm rain suppression and delay are most significant when cloud droplet number concentrations are high at cloud bases. This is in contrast to the characteristics of condensational invigoration induced by ultrafine particles that does not have a significant delay in rain because the fast production of warm rain is needed to produce low droplet number thus high supersaturation in updrafts with the supply of moisture from low levels for activation of ultrafine particles.

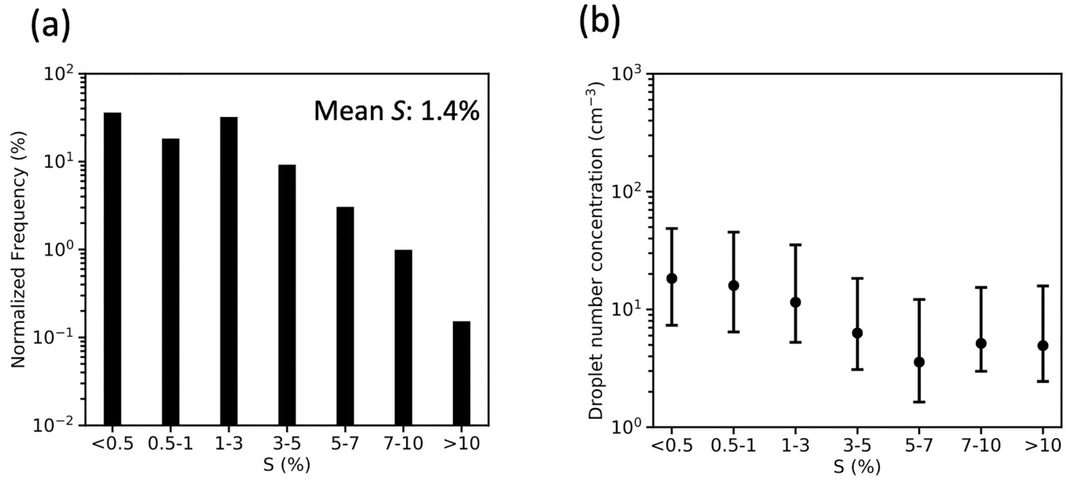

Figure 3. (a) Normalized frequency of supersaturation and (b) median and 10th to 90th percentiles of N_c for each supersaturation bin S from the C_BG simulation (background aerosols with UAPs) described by Fan et al. (2018), sampled with the same criteria as Figure 2, that is, $w > 2$ m s^{-1}, $N_c > 1$ cm^{-3}, and liquid water content >0.005 g m^3 for the developing stage of the deep convection clouds at 16:00–17:10 UTC.

The processes important to this mechanism are complicated and exhibit pronounced temporal and vertical evolution. It starts from high droplet nucleation near cloud bases, resulting in warm-rain suppression. Subsequently, cloud droplets are vertically transported to higher levels, where freezing occurs in the cold phase. Stronger droplet freezing enhances ice processes such as riming and deposition. Throughout this process, both condensate loading and offloading occur, influencing cloud dynamics. Condensate offloading refers to the sedimentation of large hydrometeors, mainly rain, graupel, and hail. It should be noted that the definition of freezing in some studies include riming (e.g., Rosenfeld et al., 2008); here freezing refers to droplet heterogenous and homogenous freezing only.

Freezing does not cause mass change, nor does riming. The increase in condensate loading occurs during the vertical transport of droplets prior to freezing or after freezing where stronger depositional growth occurs. The loading effect can occur at different stages of convective development and at varying vertical levels, as elaborated by Fan and Khain (2021). Therefore, the compensation by condensate loading to the effect of latent heating involves convoluted processes. Furthermore, the effect of condensate off-loading due to sedimentation of large hydrometeors can be significant, which increases buoyancy. During the convective transport, autoconversion and accretion in the updrafts can be strong and form rain then fall out of the clouds. Also, in the cold phase, when riming forms graupel and hail, which can induce positive feedback-collecting more droplets to form larger sizes. The sedimentation of hail and graupel can have a strong off-loading effect and lead to a large invigoration as shown in parcel model tests of Rosenfeld et al. (2008). However, when the off-loading effect is assumed to be smaller compared to the assumptions used in Rosenfeld et al. (2008) in the parcel model with the offloading effect by accretion of cloud droplets at lower levels neglected, no significant invigoration was suggested (Igel and van den Heever, 2021). To recap, enhanced latent heating induced by droplet freezing occurs at higher altitudes, condensate loading takes place during vertical transport, and condensate off-loading happens when rain, graupel, and hail form at low and middle levels. As a result, the overall effects—whether invigoration, suppression, or no effect—depend on the relative magnitudes of these three effects, which can be highly case-dependent.

Here are two additional key points often overlooked when discussing the freezing-induced effects:

1. Stronger Ice Processes: The enhanced ice processes, such as riming and deposition, as a result of increased ice formation due to freezing have been identified as an important factor for convective invigoration (e.g., Fan, Rosenfeld, et al., 2012; Khain, 2009; Khain et al., 2008; Lebo et al., 2012; Storer & van den Heever, 2013). Note that in mixed-phase regimes the enhancement of ice deposition by aerosols cannot be accurately simulated when the Wegener–Bergeron–Findeisen process is parameterized (i.e., not explicitly calculated based on supersaturation) in the bulk schemes that do not predict water supersaturation.

2. Latent Heating Profile Changes: Changes in the vertical profile of latent heating by freezing-induced invigoration, characterized by increased heating at higher levels and reduced heating at lower levels, can induce alterations in circulation and convergence, thereby enhancing secondary convection (Fan, Rosenfeld, et al., 2012). Feng et al. (2018) demonstrated that the microphysical scheme that simulates stronger vertical heating gradients can lead to the generation of stronger potential vorticity, often manifesting as a mesoscale convective vortex anomaly in the mid-troposphere.

In summary, the effectiveness of the freezing-induced invigoration mechanism depends on convoluted microphysical and dynamical processes, which may lead to a wide range of possible outcomes. A general conclusion about the significance of the mechanism requires a large body of studies across global tropical and subtropical regions (convective storms at midlatitudes generally have cold cloud bases, which limits the suppression of wain rain - necessary for effective freezing-induced invigoration). Significant effects can occur when there is a high number of aerosol particles in the environment that can be activated at cloud bases under a weak wind shear condition (Fan, Rosenfeld, et al., 2012). Cloud parcel model tests can be useful to show a variety of possibilities. However, real-case simulations of the temporal and vertical evolution of the convective microphysical processes for DCCs are needed to understand the effectiveness of this mechanism in realistic storms.

3. Key Aerosol and Meteorological Conditions for Significant Aerosol Effects

Indeed, the significance of both mechanisms depends on aerosol and meteorological conditions. For condensational invigoration, the effect is manifested when adding a high concentration of UAPs to a background with few large aerosol particles. This means that at cloud bases, droplet numbers would be fewer, forming warm rain more quickly and falling out, depleting droplets for condensation and resulting in a high supersaturation scenario that can activate a larger number of UAPs. Favorable meteorological conditions would be a warm and humid planetary boundary layer with an abundant moisture supply into the updrafts. In addition, warmer cloud bases (indicating the deeper warm layer depth) can help boost the effects (Gayatri et al., 2022; Z. Li et al., 2011; Prabha et al., 2011). We may thus see the significance of the condensational invigoration mechanism in many tropical and subtropical storms, as shown in several modeling studies (Chen et al., 2017, 2020; Fan et al., 2007, 2018, 2020; Khain et al., 2012; Lebo, 2018; Manoj et al., 2021; Sheffield et al., 2015).

For freezing-induced invigoration, its significance can be much less compared to condensational invigoration when the same percentage of latent heating is increased, as shown in the sensitivity tests conducted by Fan et al. (2018) and Lebo (2018). The latent heating increase in the freezing-induced invigoration mechanism is less effective in enhancing storm intensity since it occurs at the higher part of the storm. However, when the latent heat increase at high levels is much larger than the increase at low levels, this mechanism can be significant because stronger upper-level heating can enhance air convergence at low and middle levels by increasing the atmospheric instability. Such a situation can occur when large-size (>100 nm) aerosols are abundant, leading to a strong suppression of warm rain and enhanced ice processes in the mixed-phase regime. Fan, Rosenfeld, et al. (2012) showed such a case—scattered convective storms formed under warm and humid conditions with weak wind shear. The case was from the Department of Energy's Atmospheric Radiation Measurement in China (ARM-China) field campaign in southeast China. The addition of large concentrations of fine aerosols (accumulation mode) suppressed the warm rain and caused an up to 3 K d^{-1} latent heat increase around 9–10 km by aerosols, 2–3 times higher compared to the increase at low levels [Figure 1e of Fan, Rosenfeld, et al. (2012)]. The change in the latent heating profile by the freezing-induced invigoration is in strong contrast to that by the condensational invigoration as shown in Fan et al. (2018; Figure 4 in the paper) from the GoAmazon field campaign, where adding UAPs caused an increase in latent heat that was much larger at low and middle levels compared to the high levels. Thus, the dominant mechanism for convective invigoration is different between the two cases (added large-size aerosols vs. UAPs). The sensitivity tests carried out by Fan, Rosenfeld, et al. (2012) showed that weak wind shear was critical for the freezing-induced invigoration to be significant (at least for that case study). Many studies have suggested that in cases where strong wind shear exists and/or cloud bases are cold, aerosols could suppress convection and precipitation due to strong evaporative cooling of the small cloud droplets and/or less efficient ice-growing processes (Fan et al., 2009; Fan, Rosenfeld, et al., 2012; Iguchi et al., 2008; Khain et al., 2008; Lebo & Seinfeld, 2011; Morrison, 2012).

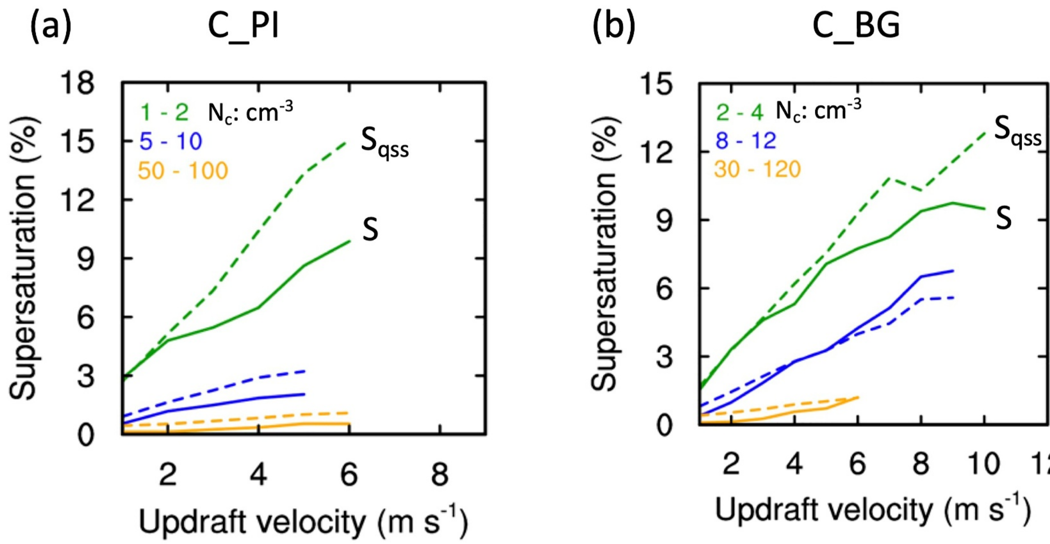

Figure 4. Relationships of supersaturation (S, solid) and quasi-steady-state supersaturation (S_{qss}; dashed) with the updraft velocity in convective cores for three droplet number (N_c) groups in the WRF-SBM simulations of (a) C_PI (the pristine condition) and (b) C_BG (background aerosols with UAPs) from Fan et al. (2018). For each N_c group and each updraft bin, values are averaged over the data with cloud droplet mass $>10^{-4}$ kg kg^{-1} and $S > 0$ or $S_{qss} > 0$ during the deep convective period, as defined by Fan et al. (2018).

4. Critical Aspects of Modeling Aerosol Impacts on DCCs

Modeled aerosol impacts on DCCs highly depend on cloud microphysics parameterizations and the aerosol setup (Fan, Leung, et al., 2012; Khain et al., 2015; Marinescu et al., 2021). Here we want to highlight the basic parameterizations and configurations needed for modeling aerosol impacts on DCCs.

4.1. Predicted Supersaturation and Explicit Calculations of Activation and Condensation

As depicted in Section 2, the magnitude of supersaturation determines how significant condensational invigoration can be. Therefore, resolving convection and predicting supersaturation is the premise for simulating this effect. The higher the model resolution, the better (large-eddy simulations are preferred). However, we are often limited by computing resources. We often choose a grid spacing of 0.5–1 km to balance the situation. Since activation of relatively large-size particles and secondary activation of smaller particles in-cloud are important to condensational invigoration, aerosol activation should be calculated explicitly based on aerosol size distribution (SD) and its hygroscopicity with predicted supersaturation.

Another requirement for simulating condensational invigoration is to explicitly calculate droplet diffusional growth (condensation/evaporation) based on prognostic supersaturation and droplet size distribution. The assumptions of saturation adjustment and quasi-equilibrium supersaturation distort or diminish aerosol effects, as detailed by Fan et al. (2020) and Khain and Pinsky (2018) with theoretical derivations. Figures 4a and 4b shows the differences between true supersaturation (S) and S_{qss} in the clean cases in the central Amazon with UAPs (right, C_BG) and without UAPs (left, C_PI) for different droplet conditions and updraft speeds during the deep convective period, based on simulations by Fan et al. (2018). S is, in general, very different from S_{qss} in C_PI, which is an ultra-clean case (Figure 4a). The very low N_c (1–2 cm^{-3}) occurs in convective cores , because of low droplet nucleation and very efficient rain formation. S_{qss} is generally much larger than S, meaning that using S_{qss} would activate more UAPs, so larger condensation heating would be achieved. However, in the case of UAPs (Figure 4b), the differences between S and S_{qss} are much smaller, and S can be larger than S_{qss}. In the case of $S > S_{qss}$, using S_{qss} for S may underestimate condensation. Based on Figure 4, high S_{qss} (>10%) from CAIPEEX 2018 could be an overestimation of the true S.

Appropriately predicting S requires the explicit calculation of droplet condensation and evaporation, which is a function of droplet number and size besides supersaturation (i.e., the integrated droplet surface area; Khain & Pinsky, 2018). The common approach for modeling clouds is to calculate condensation based on saturation adjustment or other bulk approaches without considering aerosol effect, then calculate N_c through activation. Such an approach diminishes condensational invigoration by aerosols. Changing to the predicted supersaturation helps not only better agree with bin model results in aerosol effects but also simulates a thunderstorm agreeing better with observations (Y. Zhang et al., 2021).

4.2. Prognostic Aerosol Size Distribution

Since the significance of the condensational and freezing-induced invigoration effects strongly depends on aerosol sizes, predicting the aerosol SD is essential. Prescribed droplet number, aerosol number, or aerosol SD at each time step are often used in many past modeling studies (e.g., Grabowski & Morrison, 2016, 2020; Heikenfeld et al., 2019), should be avoided. Fan, Leung, et al. (2012) and Y. Wang et al. (2013) elaborated on how the fixed aerosol SD caused excessive droplet nucleation, thus extremely high but small cloud droplets, which led to an opposite aerosol impact on convection compared to predicted aerosol SD. Fine and coarse soluble aerosols have opposite effects on DCCs, as found by a recent observational study by Liu et al. (2022). Predicting aerosol SD during the model integration allows the relatively large-size particles that are activated first at cloud bases to be removed so that smaller particles have a chance to be activated next. This is particularly important for the effect of UAPs because UAPs would not be activated easily when relatively large-size particles are always present in the situation of a fixed SD.

A common approach to prognose aerosols in cloud-resolving models (CRM) without chemistry included is the prediction of aerosol number concentrations and mass with a fixed SD shape and prescribed initial SD, as in Y. Wang et al. (2013). Such an approach is certainly better than the fixed aerosol SD at each timestep, but it does not fully account for the effects of SD change. To fully represent aerosol SD evolution, the shape needs to be predicted. Bin representations to predict aerosol SD change is considered to be a better approach for such a consideration, as used in many studies that have simulated significant aerosol effects (e.g., Fan et al., 2007, 2018; Khain et al., 2005, 2012).

In summary, when carrying out modeling studies of aerosol impacts on DCCs, the first thing to check is whether the model has prognostic aerosol SD for aerosol activation and physics-based condensation and evaporation. For UAP effects, it is important to consider whether the concentration of large-size particles (>100 nm) is low and the ratio of UAP to large-size particles is high.

4.3. Open Versus Closed Lateral Boundary Conditions

There are two common approaches in simulating clouds at a cloud-resolving scale: (a) idealized CRM or large-eddy simulations (LES) initialized with a sounding-like profile, forced by horizontally homogeneous large-scale and/or surface flux forcing, and often supplied with periodic lateral boundary conditions; and (b) real-case simulations in which the initial and lateral boundary conditions are provided by global/regional analysis/reanalysis or forecasts with open-boundary conditions and realistic land-surface conditions. It is often called a "limited area model" simulation (Zhu et al., 2012). Dagan et al. (2022) demonstrated that idealized simulations with periodic boundary conditions are problematic for simulations of aerosol impacts on DCCs. It is responsible for the substantial increase in humidity in the middle and upper troposphere by aerosols, leading to inflated convective invigoration over a 60-day model integration in Abbott and Cronin (2021). The fixed droplet number setup can be another potential source of the large increase in moisture by aerosols due to unrealistically large evaporation in the polluted condition. For investigating aerosol impacts on DCCs, we recommend the real-case model setup with open boundary conditions at CRM or LES scales, as done in many of our simulations which were also evaluated using observations (e.g., Fan, Leung, et al., 2012, 2013, 2018, 2020).

4.4. Limitations of Parcel Models in Assessing Aerosol Impacts on DCCs

Parcel models are commonly used in studying cloud microphysical processes and aerosol impacts on DCCs (Khain et al., 2000; Pinsky & Khain, 2002; Rosenfeld et al., 2008). Recent parcel model simulations have questioned the significance of freezing-induced invigoration (Igel & van den Heever, 2021; Peters et al., 2023). Indeed, cloud parcel models are a useful tool for navigating a spectrum of possible results. However, they are not realistic simulations of the convoluted cloud microphysical and dynamic processes in the course of convective development. Convective temporal evolution is important to both invigoration mechanisms as discussed in Section 2, particularly freezing-induced invigoration. Latent heating, condensate loading, and entrainment may peak at different times and vertical levels of convective development, mattering to the significance of aerosol effects (Fan & Khain, 2021; Pinsky et al., 2013, 2014), but the spatiotemporal evolution of these processes is difficult to represent in cloud parcel models. For example, the vertical profile change in latent heating may induce circulation changes spatially and intensify storms. However, this cannot be considered in parcel model studies. Another example is that entrainment may exert effects on the periphery, but not on the strong convective cores of DCCs, which cannot be considered as well. Therefore, we emphasize that cloud parcel model results should be treated as the test of the possibilities and 3-D real-case simulations of DCCs are needed for gaining the assessment of aerosol impacts on DCCs in realistic conditions.

Since DCCs can be sensitive to any small perturbation in environments (Morrison, 2012), to achieve more robust results, perturbing both aerosol and meteorological conditions by carrying out a large number of ensemble simulations would be preferable. Recent modeling studies using this approach have suggested that aerosol-induced changes in precipitation are statistically significant in their studied DCC cases (X. Li et al., 2021; Miltenberger et al., 2018).

5. Critical Aspects of the Observational Analysis of Aerosol Impacts on DCCs

Studying aerosol impacts on convective invigoration (in particular, concerning updraft intensity) based on observations is challenging because of (a) the lack of measurements of key cloud microphysical and thermodynamical properties in convective cores with strong updrafts, and (b) measurements are often incomplete, or not reliable/long enough to address the aerosol and meteorology co-variability issue. Also, surface aerosol measurements used in past analyses may not represent aerosols supplied into the updrafts to affect clouds. Furthermore, such analysis often neglects the role of aerosol size distribution. Therefore, observational data alone may not be sufficient or robust enough to either prove or disprove aerosol invigoration effects. However, observational analyses can be useful in validating model simulations, proposing hypotheses or pointing directions for further studies. For example, given the observational sample sizes are too small to be conclusive (only 17 cases), Fan et al. (2018) employed the observational analysis to motivate a detailed, real-case physical modeling work to understand why and under what conditions the impacts of ultrafine aerosol particles can be significant. Observational results could highly depend on data analysis approaches (e.g., Fan et al., 2018; Öktem et al., 2023; Oza, 2023), which can also be inconclusive by itself. Here we discuss the appropriate methodologies to analyze data for examining aerosol impacts on convection.

5.1. Mean Versus Maximum Supersaturation

Supersaturation cannot be directly measured. The quasi-steady state assumption is applied to derive observed S_{qss}. This assumption may work in most parts of clouds with weak updrafts and a relatively high number of cloud droplets (e.g., 100 cm^{-3} or larger). However, the assumption breaks down when N_c is low, particularly in strong updrafts as discussed in Section 3. Since S_{qss} represents the best possible observations that we can currently achieve, we have to accept this uncertainty in the observational analysis.

Since strong updrafts determine the storm intensity, the analysis needs to focus on those strong updrafts. Strong updraft grid points are only a very small fraction of cloudy points (e.g., 0.64% for strong updrafts defined as the top 10th percentile of cloudy points with $w > 2$ m s^{-1}, as shown in Figure 5a at an altitude of 5 km). At the very least, convective cores should be examined. Note that the fraction of convective cores, defined as having $w > 2$ m s^{-1} and total condensate >1e^{-5} kg kg^{-1} here, covers only 6.3% of the cloudy area (Figure 5a) and that high supersaturation occurs in these convective cores in model simulations (Figure 5a, colored contours).

High supersaturation is very unstable and may quickly decrease by condensation, particularly with new droplet formation. It is thus deemed to occur in a low frequency but can play an important role. There is a very large variability of S in convective clouds and averaging supersaturation over cloudy points or convective cores leads to a small mean supersaturation as shown in Figure 2c (observation) and Figure 3a (C_BG). Even in the pristine condition of Amazon (C_PI), the mean supersaturation in cloud updrafts ($w > 1$ m s^{-1}) is only 1.6% (Figure 5c). The corresponding value for C_BG (background aerosol condition of the Manaus region) is only 0.7% (Figure 5d). However, significant condensation invigoration occurs when UAPS are added to both cases as shown in Fan et al. (2018), because it is driven by the small fraction of high supersaturations. Therefore, it is problematic to claim no invigoration based on the changes of mean S (e.g., Romps et al., 2023). In the 2009 CAIPEEX field campaign data set that Romps et al. (2023) used, although the mean S is only 0.8%, there are 0.6% of the samples

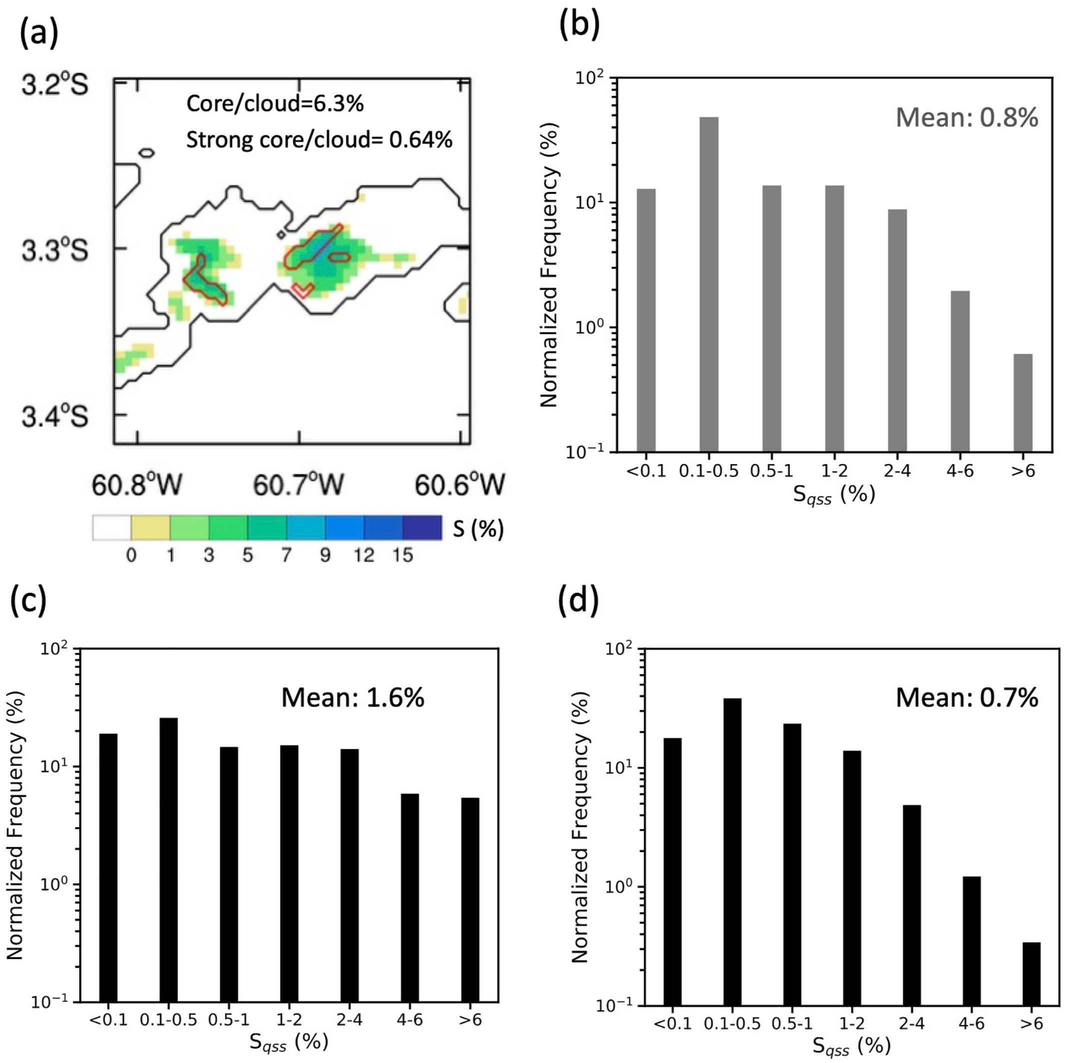

Figure 5. (a) Supersaturation S (colored contours) at the 5-km altitude at 17:25 UTC from the C_BG simulation. The black line denotes the cloudy area defined as the total hydrometeor mass larger than 10^{-5} kg kg^{-1}, and the red line displays the convective core (cloudy points with $w > 2$ m s^{-1}). (b) Normalized frequencies of S_{qss} from the 2009 CAIPEEX campaign (Prabha et al., 2011) for data with $w > 1$ m s^{-1}, following the criteria described by Romps et al. (2023). CAIPEEX 2009 data are the same as those from Romps et al. (2023). Panels (c) and (d) are the normalized frequencies of supersaturations in C_PI and C_BG, respectively, for $w > 1$ m s^{-1} and cloud droplet mass $>10^{-4}$ kg kg^{-1} during 14:00–17:10 UTC, consistent with the way of sampling in panel (b).

in updrafts ($w > 1$ m s^{-1}) having S larger than 6% (Figure 5b), with a maximum S of 9.9%. S of 6% can lead to a 3.5 m s^{-1} potential increase in w by CCN when the baseline w is ∼3 m s^{-1} in the lower atmosphere and can be a 12 m s^{-1} increase at an altitude of 8 km based on Figures 1a and 1b. In the 2018 CAIPEEX field campaign, which sampled deeper cumuli compared to the 2009 CAIPEEX field campaign, the mean supersaturation is also only 0.9%. Again, there are 1% values larger than 10% (Figure 2c) and the maximum supersaturation is 16.7%. In summary, high or maximum supersaturation should be examined for aerosol impacts on convection, not the mean supersaturation.

5.2. Aerosol-Meteorology Co-Variability

In observational analysis, the most challenging part is to isolate aerosol effects. Examining whether aerosols and meteorology co-vary or not and how the aerosol and meteorology co-variability contributes to observed aerosol-cloud correlations when co-varying does exist is important. Long-term measurements are needed to single out

10.1029/2024JD041931

aerosol effects in regions where the meteorology varies significantly, such as the U.S. Southern Great Plains (Z. Li et al., 2011; Varble, 2018). We note (a) to examine the covariability, aerosol and meteorological conditions associated with the studied DCCs should be sampled before cloud initiation, ensuring minimal influence from clouds and precipitation (cumulus and stratocumulus clouds may occasionally precede DCCs). Ideally, such sampling should be under clear-sky conditions at the middle and lower levels, prior to but close to cloud initiation. After DCC starts, the aerosol and meteorological conditions are already modified by aerosol effects and they should not be used to examine the aerosol and meteorology co-variability for the targeted DCC; (b) the presence of aerosol-meteorology co-variability does not rule out aerosol effects, and more in-depth analysis and modeling are needed to understand how aerosols and meteorology respectively affect the DCCs when co-varying does exist.

Here we show an example of how the relationship between aerosols and CAPE is changed by comparing the sampling with aerosol and meteorology influenced by clouds to that in the clear sky conditions before the cloud. Varble (2018) reexamined the study of Z. Li et al. (2011) and found there is a positive correlation between aerosols and meteorology (e.g., CAPE). In Varble (2018), CAPE was sampled at the start time of cloud objects (denoted by the dark pule arrow in Figure 6a). Aerosols were taken from the median values over the period between the start of cloud objects and the start of warm-based deep clouds (WDCs), which is the range denoted by the red arrow in Figure 6a. The definition of WDCs is exactly the same as Varble (2018), that is, cloud objects with maximum cloud base temperature (CBT) $\geq$ 15°C, minimum cloud top temperature (CTT) ≤ -4.0°C, CAPE > 0, lifted condensation level (LCL) > 15°C, and level of neutral buoyancy (LNB) < LCL. We reproduced the results of Varble (2018) based on the data and data processing methods from the study. That is, Varble18_Prod is very similar to Varble18_copy in Figure 6b. The only difference between Varble18_Prod and Varble18_copy is that the CAPE in Varble18_Prod is calculated by us as a sanity check. Since Varble18's data lack the CAPE values needed for our analysis (clear-sky times within 30 min before the cloud object), we calculate our CAPE values. However, the relationship between aerosols and CAPE is qualitatively changed and there is no such a strong positive correlation between them in Varble18_NewAerMet (blue curve in Figure 6b), when we sampled the aerosols and CAPE consistently from the clear-sky conditions within the 30-min period before the cloud object (denoted by two vertical light purple lines in Figure 6a). The reason for sampling before the cloud object is because there are about 9.4% of selected cases with surface precipitation occurring between the start of cloud objects and the start of WDCs (e.g., 19 June 1998, 12 August 1998, and 10 September 1999). The aerosol concentrations near the surface between the time period would be affected by the precipitation in these cases. Furthermore, we also found that 51% WDCs identified in Varble (2018) overlap in time duration, we excluded these WDCs with start time located at the duration of another WDC to remove the overlapped WDCs, which referred to as Varble18_ NewAerMet_NoOverlap. The result does not show any correlation at all (Figure 6b; red). This means that the positive correlation between aerosols and CAPE is not validated when we have resampled the WDC cases in a more physical and consistent way. Anyhow, the standard errors are large for some aerosol bins so increasing the sample sizes is needed to give a robust estimation of co-varying or not. Other uncertainties in this type of observational analysis include not accounting for aerosol flux into clouds, aerosol size distribution, and hygroscopicity. Therefore, we stress that statistical results of observations are subject to large uncertainties due to limited data samples and unconstrained conditions. A combination of observational analysis and modeling is thus important to providing insightful studies.

5.3. Identification of Convective Initiation and the Convective Period

Identification of convective initiation and the convective period is important for observational analyses of aerosol impacts on convection, particularly for the analysis of radar data. This is because (a) aerosols should be sampled before clouds, as discussed above, and (b) DCCs can have a shallow-cloud stage and then a stratiform/anvil cloud stage, which should be excluded in the analysis. There are different ways of identifying convective initiation, depending on what observations are available. However, using the surface rain initiation time as the convective initiation time such as in Öktem et al. (2023) may be questionable for the convective storms evolving from a period of shallow cumuli. Convective initiation, in this case, can be much earlier than surface rain initiation.

For identifying the convective period, the basic principle is that there should be no significant shallow or stratiform clouds included in the analysis. The recent study by Öktem et al. (2023) did not constrain their analysis to convective periods. Most of the local convective storms had a precipitation period of 1–2 hr, and they used the data sampled from 1-, 2-, 3-, 4-, or 6-hr time intervals centered around the peak of convection. The 3–6 hr time interval would include stratiform clouds and/or shallow clouds. Although the 90th percentile of updraft velocity was selected to be examined, including shallow and stratiform clouds may populate the data set with a high

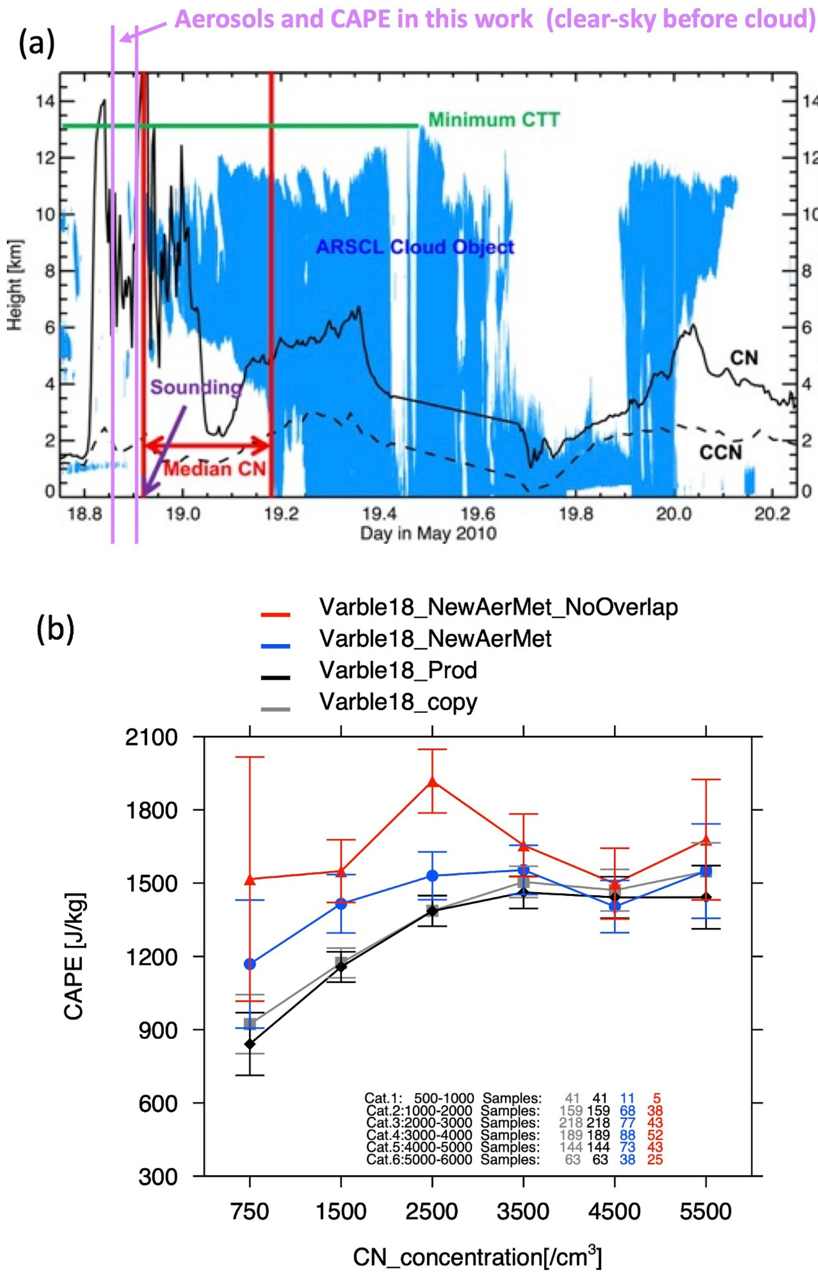

Figure 6. (a) An example time–height cloud object (blue) showing the time for sampling aerosols (CN, red lines) and meteorology (Sounding in dark purple), which is taken from Figure 1 of Varble (2018). © American Meteorological Society. Used with permission. We added the two light purple lines to denote our sampling period of aerosols and meteorology, which is within 30 min before the identified cloud object. (b) Relationship of CN with convective available potential energy (CAPE) from Varble18_copy (gray), Varble18_Prod (black), Varble18_NewAerMet (blue), and Varble18_NewAerMet_NoOverlap (red). Varble18_copy is the result of Varble (2018). Varble18_Prod is the result we have reproduced with the data and data processing from Varble (2018) except that the CAPE is calculated by us, which produces similar values as Varble (2018). In Varble18_NewAerMet, the aerosols and CAPE for each WDC are sampled from the clear-sky conditions within a 30-min period before the cloud object, and other things are the same as Varble18_Prod. The result of Varble18_NewAerMet_NoOverlap is produced by further excluding the overlapping WDCs in Varble (2018). The standard errors of the mean are shown with vertical lines. The numbers in (b) show the sampel sizes for each method.

proportion of weak updrafts. As a result, the 90th percentile value does not necessarily indicate a DCC updraft, particularly the strong updrafts that are most relevant to convective invigoration.

Another note is that N_c varies significantly among the updrafts of a single storm. In other words, even in the storms develop in a polluted environment, there exist clean updrafts with low N_c. Aerosol effects should be analyzed by identifying clean and polluted storms based on the aerosol conditions prior to the onset of a cloud examined, not by analyzing the changes from low to high N_c within a storm. The strong updrafts of each identified storm (such as the top 10th or 25th percentiles of convective cores) may be selected for analysis, and the aerosol effects should be examined based on the changes from clean to polluted storms.

5.4. Convective Invigoration Versus Cloud Invigoration

As defined in Introduction, "convective invigoration" refers to enhanced updraft velocity or vertical mass flux in convective cores. However, the term "cloud invigoration" can refer to taller clouds and larger clouds reflected by cloud top height (CTH), cloud thickness, and cloud cover, which are dominated by stratiform and anvil of DCCs since convective cores have a very small cloud fraction. Observational studies often analyze the relationships of CTT, CTH and/or cloud thickness with aerosols (e.g., Koren et al., 2005; Z. Li et al., 2011). Their increases indicate "cloud invigoration" but do not necessarily mean "convective invigoration," as elaborated by Fan et al. (2013) where they found only up to ¼ of cloud invigoration has convective invigoration occurring. This is because the increases in CTT and cloud thickness in the stratiform and anvil of DCCs can be caused by aerosol microphysical effects, that is, (a) more cloud mass detrainment from convective cores to stratiform as a result of suppression of rain and a larger upward transport of condensate mass, and (b) the reduced ice particle size due to freezing of a large number of smaller droplets in the polluted clouds slows dissipation of stratiform and anvil clouds. Therefore, a better approach for such observational studies is to examine aerosol impacts on CTT/CTH and cloud thickness in convective cores (the tower) and the stratiform and anvil parts of DCCs separately (Yan et al., 2014). If aerosols increase CTT/CTH and cloud thickness in convective cores, convective invigoration likely occurs, which also contributes to increased CTT and cloud thickness in the stratiform and anvil parts. However, if aerosols increase CTT and cloud thickness in the stratiform and anvil but not in the convective cores, convective invigoration is likely lacking, and cloud invigoration in the stratiform and anvil parts would be mainly caused by the aerosol microphysical effect as detailed in Fan et al. (2013). Identification of convective cores in observations is extremely challenging, particularly before the Geostationary Operational Environmental Satellites (GOES-R series). The GOES-R series and surface-based radar data can make the identification relatively more reliable (Cooney et al., 2021; Y. Lee et al., 2021).

The invigoration in convective cores affects weather hazards such as extreme precipitation, hail, and lightning, while cloud invigoration in stratiform and anvil mainly affects aerosol-induced radiative forcing and climate (Nishant et al., 2019; Yan et al., 2014). The effects of aerosols on anvil radiative properties are not yet well understood and their global radiative effect is not quantified. A recent study estimated a global effect of order 0.2–0.5 W m^{-2}, per aerosol doubling, for aerosol effects on tropical convection (Nishant et al., 2019). Observational studies based on satellite and ground data have suggested a net daily mean positive radiative effect of 29.3 W m^{-2} at the TOA and 22.2 W m^{-2} at the surface at the Southern Great Plains (SGP) from the clean to polluted conditions for days with DCCs (Yan et al., 2014). Accounting for the frequency of occurrence of DCCs and aerosol loading, the long-term (10 years) and diurnal mean forcing of DCCs induced by aerosols amounts to 0.5 W m^{-2} warming at the TOA. However, over the tropics, a negative radiative effect at the TOA up to −76 W m^{-2} was seen because shortwave cooling overwhelmed the longwave warming (Peng et al., 2016; Sarangi et al., 2018). Although there is no qualitative agreement in the net forcing, the increased cloud fraction by aerosols consistently enhances the surface cooling in the daytime and nighttime warming, leading to a reduced diurnal temperature variation (~0.5–0.6 K), as observed (Sarangi et al., 2017) and simulated (Fan et al., 2013). A solid understanding of aerosol radiative forcing through DCCs requires quantification of the changes of optically thick and thin anvil clouds by aerosols.

6. Aerosol Impacts Through Feedback to Circulation and Meteorology

Aerosol impacts on convective systems involve the feedback of primary aerosol effects to the circulation and meteorology at various scales, which further affect clouds. The feedback of the primary invigoration (any invigoration mechanism including both condensational and freezing induced) can enhance air convergence and

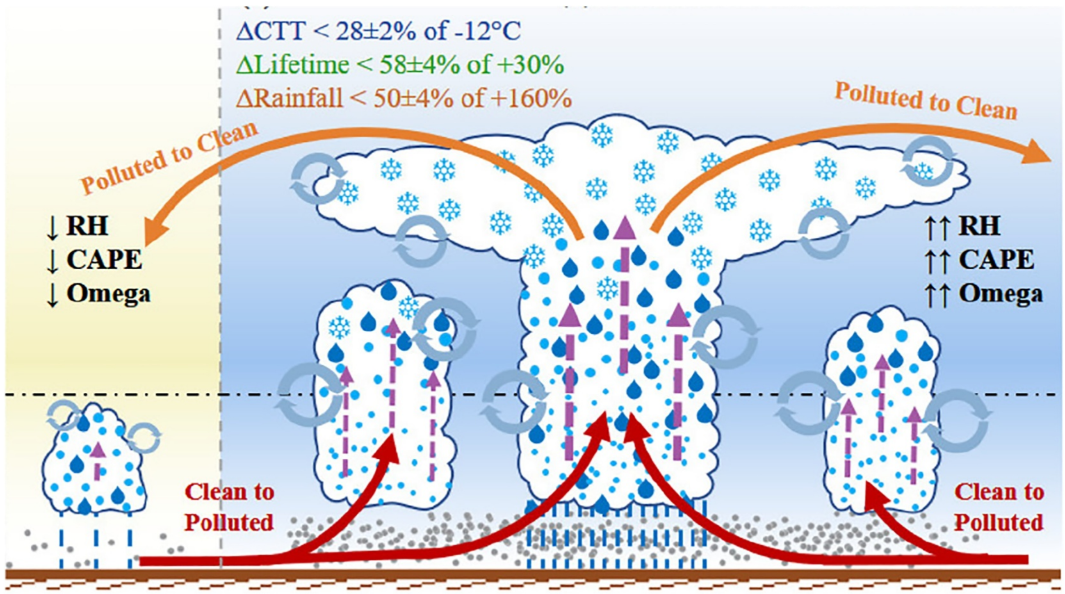

Figure 7. A schematic diagram of aerosol - convection - circulation invigoration for deep tropical convective clouds. The primary invigoration (purple arrows) induces low-level convergence (magenta arrows) that moistens the troposphere (blue) and further invigorates the convection in a positive feedback loop (orange arrows). The descending branch of the circulation suppresses clouds and precipitation over the leaner area [from Zang et al. (2023)].

divergence at various scales (Bell et al., 2008; Blossey et al., 2018; Dagan et al., 2023; Fan, Rosenfeld, et al., 2012; S. S. Lee, 2012; Morrison & Grabowski, 2013; Zang et al., 2023), which can amplify aerosol impacts.

This induced circulation enhances the moisture convergence and convective clouds in a positive feedback loop. Simulations in a domain size of 1,100 km showed that the descending branch of the circulation suppressed the convection away from the aerosol perturbation (S. S. Lee, 2012). A feedback mechanism occurs if aerosols are added to a large-scale system where part of it contains synoptically preferable deep convection, such as the intertropical convergence zone (ITCZ) and the adjacent subtropical trade wind cumulus. The aerosol-induced suppressed rain in the trade wind cumulus enriches the vapor convergence to the deep clouds in the ITCZ, invigorates them and, in turn, enhances the subsidence over the trade wind cumulus, which further suppresses the clouds and precipitation and enhances the vapor convergence into the ITCZ in a positive feedback loop, as simulated by Dagan et al. (2022, 2023).

Such feedback can explain at least part of the observed co-variability of meteorology with aerosols. In other words, this co-variability can be very likely caused by the meteorological feedback induced by the primary invigoration. The feedback is illustrated in Figure 7, which shows the air convergence at low levels induced by the primary invigoration leads to increases in both aerosols and moisture/CAPE. A recent observational study of tropical deep convective clusters showed that this co-variability as a result of the feedback explains approximately half of the indicated enhancement in precipitation (Zang et al., 2023). The feedback can be modulated by environmental conditions and a different feedback mechanism might occur in a dry environment due to strong evaporation. The feedback mechanism requires simulations with a large domain (>500 km or larger) and high model resolutions (a couple of km or finer) and is difficult to capture by small domain simulations or coarse resolutions.

The feedback of the primary aerosol effects on circulation is complicated, depending on the types of mesoscale convective systems (MCSs). Each type of MCSs, such as self-organized MCSs, squall lines, and supercells, has its unique dynamics and microphysics-dynamics feedback. Studies should be conducted respectively for each type over their whole lifecycles.

For tropical self-organized MCSs, as described above, the meteorological feedback due to the enhanced latent heating by aerosols can be large, contributing to further invigoration. For squall lines, the balance between the cold pool and low-level wind shear plays a key role in their convective intensity (Rotunno et al., 1988). Therefore, examining how the balance is changed by aerosols is key in these systems. Both Chen et al. (2020) and Lebo and

Morrison (2014) found that increasing CCN from a clean condition (a couple of hundred of particles per cm^{-3}) lead to a more optimal balance between the cold pool and low-level wind shear due to the reduced cold pool intensity under polluted conditions, thus an increase in low-level convergence and larger updraft speeds were seen. For supercells that are strongly dynamically forced, limited aerosol effects on convective intensity are not a surprise (Heikenfeld et al., 2019; Lin et al., 2021). However, the precipitation and hail number and size in supercells can be notably enhanced (Kalina et al., 2014; Khain et al., 2008; X. Li et al., 2021; Lin et al., 2021; Loftus & Cotton, 2014). Aerosols may also modify large-scale storm systems, such as tropical cyclones, hurricanes, and mid-latitude storm tracks (e.g., Carrio & Cotton, 2011; Y. Wang, Lee, et al., 2014; Y. Wang, Wang, et al., 2014; R. Zhang et al., 2007; Zhao et al., 2018).

Since tropical convection drives much of the global circulation (Riehl & Malkus, 1958), the aerosol effects on tropical convection can modulate the global circulation. The tropics are also the source of many of the world's major aerosol emissions with heavy aerosol loadings such as dust from Africa and the Mideast; biomass burning in the Amazon, Central Africa, and Southeast Asia; and anthropogenic aerosols from India and South China, where extensive studies have been conducted, revealing very strong effects of aerosol-cloud interactions (Z. Li et al., 2016, 2019; Sarangi et al., 2018; Q. Wang et al., 2018). As such, the total effects in extensive polluted regions may amount to such a large magnitude that it can influence the large-scale circulation, including the meridional one transporting excessive energy surplus from the tropics to the energy-deficit regions at high latitudes (Dagan et al., 2022). The feedback of aerosol effects to circulation discussed in this section can be significant but is difficult to be singled out in observational studies. Much more work is needed to prove those hypothesized effects.

7. Summary

Aerosol-cloud-interaction (ACI) has been a major source of uncertainties in estimating climate forcing. Aerosol interactions with DCCs can aslo affect hazardous weather. The aerosol-invigoration effects driven by latent heat release are an important ACI mechanism that has been fervently debated in recent years. To help clarify the effects, we have addressed key aspects of condensational and freezing-induced invigoration mechanisms, as well as the aerosol and meteorological conditions under which these mechanisms become significant. Learning from the merits and limitations in previous studies and based on new findings from this study, we offer practical recommendations on approaches for investigating aerosol impacts on convection using both modeling and observational methods.

Condensational invigoration is caused by an increase in latent heating due to the enhancement of condensation. The resulting increase in buoyancy affects the vertical velocities over the parcel trajectory, and the largest effects should be examined from vertical profiles, not the low levels only. This mechanism can be significant when supersaturation in clean conditions is high, convective intensity is moderate, and there are low number concentrations of large-size aerosol particles (>100 nm) and a high ratio of UAPs to large-size particles. The intrusion of anthropogenic aerosols in the warm and humid tropics and subtropics as shown by Fan et al. (2018, 2020) can demonstrate large effects.

Freezing -induced invigoration is driven by multiple microphysical and dynamical processes occurring at different times and locations during convective development. The processes supporting invigoration of convection include enhanced latent heating, condensate offloading, stronger ice processes (riming and deposition), and even circulation change and convergence induced by larger latent heating at the high levels. Whereas the processes suppressing convection are enhanced condensate loading and entrainment. Freezing -induced invigoration can be significant in warm, humid, and weak wind shear environments with a high number of relatively large-size particles that can be activated around cloud bases.

Our recommendation for appropriately simulating aerosol impacts on convection includes:

- Predicted supersaturation and explicit calculations of activation and condensation, not using saturation adjustment or quasi-steady state assumptions.
- Prognostic aerosol size distribution, not fixed aerosol or droplet numbers, or fixed aerosol SD particularly for the condensational invigoration.
- Real-case forecast-type simulations at CRM or LES scales with open lateral boundary conditions, not idealized periodic boundary conditions.

It should be noted that extra model developments are needed for studying ACI with bulk microphysics schemes in the standard Weather Research and Forecasting (WRF) model, in order to consider explicit calculations of activation and condensation. The examples of such developments for the Morrison scheme are shown in Y. Wang et al. (2013) and Y. Zhang et al. (2021). In addition, a stronger microphysics-radiation coupling is needed to study ACI (i.e., the radiation schemes need to take the droplet and ice effective radii calculated from the cloud microphysics scheme).

For analyzing aerosol impacts on convection with observational data, we recommend the following:

- Identify convective initiation and convective periods of DCCs, excluding contamination of other clouds, such as shallow clouds or stratiform/anvil parts of DCCs. Then sample strong convective cores of each storm (e.g., the top 10th or 25th percentiles of updrafts with $w > 2$ m s^{-1}). The changes of the high and maximum supersaturation and updraft velocity should be examined, not the mean values.
- Examine the aerosol-meteorology covariability using aerosol and meteorology samples with least influenced by clouds (ideally, take the data from clear-sky conditions at the low and middle levels before the studied cloud forms).
- For the studies using CTT (CTH) or CT, examine aerosol impacts on convective cores and the stratiform/anvil parts of DCCs, separately.

Aerosol impacts on convective systems involve complicated feedback processes by the primary aerosol effects (i.e., the invigoration via enhanced latent heating), which might amplify or reduce the aerosol effects, depending on the types of convective systems. The modeling studies with a domain over a few hundred kilometers such as Fan, Leung, et al. (2012) and Fan et al. (2013, 2018) should have captured some of the feedback. The feedback of the primary aerosol effects on circulation at various scales is complicated, depending on the types of convective systems. Therefore, aerosol effects would not be comparable when the studied storm types differ. Systematically studying aerosol impacts on DCCs and their dependence on storm types is crucial for understanding the significance of ACI via DCCs, which calls for tremendous future observational and modeling efforts. Unraveling the feedback induced by the primary invigoration on circulation and meteorology is key to fully understanding aerosol impacts on convection and precipitation qualitatively and quantitatively. Thus, this is an important future focus for aerosol-DCC interactions.

We strongly recommend a combination of observational analysis and modeling for more insightful studies. For DCCs, we also advocate simulating realistic cases and evaluating them with comprehensive observations to show the fidelity of the model simulation. Although there are large uncertainties in cloud physical parameterizations, our recommended model configurations for studying ACI might help qualitatively align model results. Currently robust measurements/calculations of key variables, such as updraft velocity and supersaturation in deep convective cores, are not available to prove or disprove the proposed mechanisms. Future field campaigns focusing on updraft velocity, supersaturation, and cloud and aerosol properties in convective cores should be carefully designed based on the above recommendations. There is currently a lack of understanding regarding the significance of aerosol-deep convective cloud (DCC) interactions on the global scale, necessitating substantial future studies.

Additionally, this work presents new findings including the existence of high supersaturation based on the observational analysis from CAIPEEX IV, the relationship of supersaturation and quasi-steady-state supersaturation at different droplet concentrations, and the analysis of aerosol-meteorology coverability found in Varble (2018) concerning the finding of Z. Li et al. (2011) using a more proper sampling method. We also offer more specific recommendations on the modeling and observational techniques compared to Varble et al. (2023), as summarized above. The general recommendations for future work agree with Fan and Li (2022) and Varble et al. (2023), including (a) integrated observational analysis and detailed modeling work, (b) improving the observations (such as measuring properties in connective cores including supersaturation) and model simulations to reduce uncertainties and increase robustness, and (c) understanding the significance of aerosol impacts over a larger scale and longer time period.

Overall, it is worth noting that the aerosol invigoration effect is conditional upon various aerosol and meteorological conditions. The existence of the aerosol invigoration effect has been based on the conditions in which the invigoration has a physical basis to work. The aerosol invigoration effect on DCCs is most significant in very clean environments. As such, large effects are expected during the transition from preindustrial to industrial periods. For instance, in southeast China, notable increase in heavy precipitation and lightning activity was observed between 1990 and 2006, corresponding to a shift from a clean to a highly polluted environment due to rapid economic development in that region (Yang & Li, 2014). Further support for aerosol invigoration effects can be seen in studies of lightning enhancements over tropical shipping lanes in the remote ocean environment (Thornton et al., 2017) and the lightning declines following the regulation of fuel sulfur emissions with seven-fold reduction by the International Maritime Organization (Wright et al., 2024).

In the present-day environment, the background is not considered clean in most of the continents, unfavorable to aerosol invigoration effects. Therefore, observational evidence should be sought in regions with low aerosol backgrounds, where high concentrations of hydrophilic aerosols—whether natural or anthropogenic—can be introduced. For example, the CAIPEEX cloud seeding experiments provided statistically significant observational evidence in the Indian monsoon seasons, which generally had a relatively clean background because of marine air and plenty of rain (with N_c of ∼200–600 cm^{-3}). These experiments, involving 278 hygroscopic seeding samples, showed that seeded storms exhibited delayed drizzle, enhanced precipitation, and longer lifetimes (Prabhakaran et al., 2023). Given that the seeding material (NaCl) consists of relatively large aerosol particles, the observed effects are more aligned with freezing-induced invigoration. Ideal locations for observational experiments would include the Amazon forest region or tropical ocean areas, which are minimally influenced by pollution but host abundant warm cloud-based DCCs.

Data Availability Statement

The code and data for plotting the figures are available at Fan et al. (2025). The modeling data are available from Fan et al. (2018). The 2018 CAIPEEX data can be available from Bera et al. (2021). The 2009 CAIPEEX data can be available from Romps et al. (2023).

Acknowledgments
Argonne National Laboratory's work was supported by the U.S. Department of Energy (DOE), Office of Science's Biological and Environmental Research (BER) Atmospheric System Research, under Contract DE-AC02-06CH11357. ZL was supported by the U.S. DOE BER Atmospheric System Research under Award Number DE-SC0022919 and by the National Science Foundation under Award Number AGS2126098. Data were obtained from the Atmospheric Radiation Measurement (ARM) User Facility, a U.S. DOE Office of Science user facility managed by the Office of BER.

References

Abbott, T. H., & Cronin, T. W. (2021). Aerosol invigoration of atmospheric convection through increases in humidity. *Science*, *371*(6524), 83–85. https://doi.org/10.1126/science.abc5181

Andreae, M. O., Rosenfeld, D., Artaxo, P., Costa, A. A., Frank, G. P., Longo, K. M., & Silva-Dias, M. A. (2004). Smoking rain clouds over the Amazon. *Science*, *303*(5662), 1337–1342. https://doi.org/10.1126/science.1092779

Bell, T. L., Rosenfeld, D., Kim, K.-M., Yoo, J.-M., Lee, M.-I., & Hahnenberger, M. (2008). Midweek increase in U.S. summer rain and storm heights suggests air pollution invigorates rainstorms. *Journal of Geophysical Research*, *113*(D2), D02209. https://doi.org/10.1029/2007JD008623

Bera, S., Chowdhuri, S., & Prabha, T. V. (2021). A new methodology for the statistical descriptions of Particle-by-Particle measurements of liquid droplets in cumulus clouds. *Quarterly Journal of the Royal Meteorological Society*, *148*, 743–859. https://doi.org/10.1002/qj.4234

Blossey, P. N., Bretherton, C. S., Thornton, J. A., & Virts, K. S. (2018). Locally enhanced aerosols over a shipping lane produce convective invigoration but weak overall indirect effects in cloud-resolving simulations. *Geophysical Research Letters*, *45*(17), 9305–9313. https://doi.org/10.1029/2018GL078682

Carrio, G. G., & Cotton, W. R. (2011). Investigations of aerosol impacts on hurricanes: Virtual seeding flights. *Atmospheric Chemistry and Physics*, *11*(6), 2557–2567. https://doi.org/10.5194/acp-11-2557-2011

Chen, Q., Fan, J., Yin, Y., & Han, B. (2020). Aerosol impacts on mesoscale convective systems forming under different vertical wind shear conditions. *Journal of Geophysical Research: Atmospheres*, *125*(3), e2018JD030027. https://doi.org/10.1029/2018JD030027

Chen, Q., Koren, I., Altaratz, O., Heiblum, R. H., Dagan, G., & Pinto, L. (2017). How do changes in warm-phase microphysics affect deep convective clouds? *Atmospheric Chemistry and Physics*, *17*(15), 9585–9598. https://doi.org/10.5194/acp-17-9585-2017

Cooney, J. W., Bedka, K. M., Bowman, K. P., Khlopenkov, K. V., & Itterly, K. (2021). Comparing tropopause-penetrating convection identifications derived from NEXRAD and GOES over the contiguous United States. *Journal of Geophysical Research: Atmospheres*, *126*(14), e2020JD034319. https://doi.org/10.1029/2020JD034319

Cotton, W. R., & Walko, R. (2021). Examination of aerosol-induced convective invigoration using idealized simulations. *Journal of the Atmospheric Sciences*, *78*(1), 287–298. https://doi.org/10.1175/JAS-D-20-0023.1

Dagan, G., Stier, P., Spill, G., Herbert, R., Heikenfeld, M., van den Heever, S. C., & Marinescu, P. J. (2022). Boundary conditions representation can determine simulated aerosol effects on convective cloud fields. *Communications Earth & Environment*, *3*(1), 71. https://doi.org/10.1038/s43247-022-00399-5

Dagan, G., Yeheskel, N., & Williams, A. I. L. (2023). Radiative forcing from aerosol–cloud interactions enhanced by large-scale circulation adjustments. *Nature Geoscience*, *16*(12), 1092–1098. https://doi.org/10.1038/s41561-023-01319-8

Fan, J., Han, B., Varble, A., Morrison, H., North, K., Kollias, P., et al. (2017). Cloud-resolving model intercomparison of an MC3E squall line case: Part I—Convective updrafts. *Journal of Geophysical Research: Atmospheres*, *122*(17), 9351–9378. https://doi.org/10.1002/2017JD026622

Fan, J., & Khain, A. P. (2021). Comments on "Do ultrafine cloud condensation nuclei invigorate deep convection?". *Journal of the Atmospheric Sciences*, *77*(1), 329–339. https://doi.org/10.1175/JAS-D-20-0218.1

Fan, J., Leung, L. R., Li, Z., Morrison, H., Chen, H., Zhou, Y., et al. (2012). Aerosol impacts on clouds and precipitation in eastern China: Results from bin and bulk microphysics. *Journal of Geophysical Research*, *117*(D16), D00K36. https://doi.org/10.1029/2011JD016537

Fan, J., Leung, L. R., Rosendeld, D., Chen, Q., Li, Z., Yu, H., & Zhang, J. (2013). Microphysical effects determine macrophysical response for aerosol impacts on deep convective clouds. *Proceedings of the National Academy of Sciences*, *110*, 48. https://doi.org/10.1073/pnas.1316830110

Fan, J., & Li, Z. (2022). Aerosol interactions with deep convection. In K. Carslaw (Ed.), *Aerosols and climate*. Elsevier. Retrieved from https://www.sciencedirect.com/book/9780128197660/aerosols-and-climate

Fan, J., Rosenfeld, D., Ding, Y., Leung, L. R., & Li, Z. (2012). Potential aerosol indirect effects on atmospheric circulation and radiative forcing through deep convection. *Geophysical Research Letters*, *39*(9), L09806. https://doi.org/10.1029/2012GL051851

Fan, J., Rosenfeld, D., Zhang, Y., Giangrande, S. E., Li, Z., Machado, L. A. T., et al. (2018). Substantial convection and precipitation enhancements by ultrafine aerosol particles. *Science*, *359*(6374), 411–418. https://doi.org/10.1126/science.aan8461

Fan, J., Wang, Y., Rosenfeld, D., & Liu, X. (2016). Review of aerosol-cloud interactions: Mechanisms, significance, and challenges. *Journal of the Atmospheric Sciences*, *73*(11), 4221–4252. https://doi.org/10.1175/JAS-D-16-0037.1

Fan, J., Yuan, T., Comstock, J. M., Ghan, S., Khain, A., Leung, L. R., et al. (2009). Dominant role by vertical wind shear in regulating aerosol effects on deep convective clouds. *Journal of Geophysical Research*, *114*(D22), D22206. https://doi.org/10.1029/2009JD012352

Fan, J., Zhang, R., Li, G., & Tao, W.-K. (2007). Effects of aerosols and relative humidity on cumulus clouds. *Journal of Geophysical Research*, *112*(D14), D14204. https://doi.org/10.1029/2006JD008136

Fan, J., Zhang, Y., Li, Z., Hu, J., & Rosenfeld, D. (2020). Urbanization-induced land and aerosol impacts on sea breeze circulation and convective precipitation. *Atmospheric Chemistry and Physics*, *20*(22), 14163–14182. https://doi.org/10.5194/acp-20-14163-2020

Fan, J., Zhang, Y., Li, Z., Yan, H., Prabhakaran, T., Rosenfeld, D., & Khain, A. (2025). Unveiling aerosol impacts on deep convective clouds: Scientific concept, modeling, observational analysis, and future direction [Data and code]. Retrieved from https://portal.nersc.gov/project/m4486/Fan_etal_Perspective/

Feng, Z., Leung, L. R., Houze, R. A., Jr., Hagos, S., Hardin, J., Yang, Q., et al. (2018). Structure and evolution of mesoscale convective systems: Sensitivity to cloud microphysics in convection-permitting simulations over the U.S. *Journal of Advances in Modeling Earth Systems*, *10*(7), 1470–1494. https://doi.org/10.1029/2018MS001305

Gayatri, K., Patadeb, S., Fan, J., & Prabhakaran, T. (2022). Pathways of precipitation formation in different thermodynamic and aerosol environments over the Indian Peninsula. *Atmospheric Research*, *266*, 105934. https://doi.org/10.1016/j.atmosres.2021.105934

Giangrande, S. E., Biscaro, T. S., & Peters, J. M. (2023). Seasonal controls on isolated convective storm drafts, precipitation intensity, and life cycle as observed during GoAmazon 2014/5. *Atmospheric Chemistry and Physics*, *23*(9), 5297–5316. https://doi.org/10.5194/acp-23-5297-2023

Grabowski, W. W., & Morrison, H. (2016). Untangling microphysical impacts on deep convection applying a novel modeling methodology. Part II: Double-moment microphysics. *Journal of the Atmospheric Sciences*, *73*(9), 3749–3770. https://doi.org/10.1175/jas-d-15-0367.1

Grabowski, W. W., & Morrison, H. (2020). Do ultrafine cloud condensation nuclei invigorate deep convection? *Journal of the Atmospheric Sciences*, *77*(7), 2567–2583. https://doi.org/10.1175/JAS-D-20-0012.1

Heikenfeld, M., White, B., Labbouz, L., & Stier, P. (2019). Aerosol effects on deep convection: The propagation of aerosol perturbations through convective cloud microphysics. *Atmospheric Chemistry and Physics*, *19*(4), 2601–2627. https://doi.org/10.5194/acp-19-2601-2019

Houze, R. A., Jr. (2004). Mesoscale convective systems. *Reviews of Geophysics*, *42*(4), RG4003. https://doi.org/10.1029/2004RG000150

Houze, R. A., Jr. (2014). *Cloud dynamics* (p. 432). Academic Press/Elsevier.

Igel, A. L., & van den Heever, S. C. (2021). Invigoration or enervation of convective clouds by aerosols? *Geophysical Research Letters*, *48*(16), e2021GL093804. https://doi.org/10.1029/2021GL093804

Iguchi, T., Nakajima, T., Khain, A. P., Saito, K., Takemura, T., & Suzuki, K. (2008). Modeling the influence of aerosols on cloud microphysical properties in the East Asia region using a mesoscale model coupled with a bin-based cloud microphysics scheme. *Journal of Geophysical Research*, *113*(D14), D14215. https://doi.org/10.1029/2007JD009774

Kalina, E. A., Friedrich, K., Morrison, H., & Bryan, G. H. (2014). Aerosol effects on idealized supercell thunderstorms in different environments. *Journal of the Atmospheric Sciences*, *71*(12), 4558–4580. https://doi.org/10.1175/jas-d-14-0037.1

Khain, A., Rosenfeld, D., & Pokrovsky, A. (2005). Aerosol impact on the dynamics and microphysics of deep convective clouds. *The Quarterly Journal of the Royal Meteorological Society*, *131*(611), 2639–2663. https://doi.org/10.1256/qj.04.62

Khain, A. P. (2009). Notes on state-of-the-art investigations of aerosol effects on precipitation: A critical review. *Environmental Research Letters*, *4*(1), 015004. https://doi.org/10.1088/1748-9326/4/1/015004

Khain, A. P., Beheng, K. D., Heymsfield, A., Korolev, A., Krichak, S. O., Levin, Z., et al. (2015). Representation of microphysical processes in cloud-resolving models: Spectral (bin) microphysics versus bulk parameterization. *Reviews of Geophysics*, *53*(2), 247–322. https://doi.org/10.1002/2014RG000468

Khain, A. P., BenMoshe, N., & Pokrovsky, A. (2008). Factors determining the impact of aerosols on surface precipitation from clouds: An attempt at classification. *Journal of the Atmospheric Sciences*, *65*(6), 1721–1748. https://doi.org/10.1175/2007JAS2515.1

Khain, A. P., Ovtchinnikov, M., Pinsky, M., Pokrovsky, A., & Krugliak, H. (2000). Notes on the state-of-the-art numerical modeling of cloud microphysics. *Atmospheric Research*, *55*(3–4), 159–224. https://doi.org/10.1016/s0169-8095(00)00064-8

Khain, A. P., Phillips, V., BenMoshe, N., & Pokrovsky, A. (2012). The role of small soluble aerosols in the microphysics of deep maritime clouds. *Journal of the Atmospheric Sciences*, *69*(9), 2787–2807. https://doi.org/10.1175/2011jas3649.1

Khain, A. P., & Pinsky, M. (2018). *Physical processes in clouds and cloud modeling* (p. 642). Cambridge University Press.

Kogan, Y. L., & Martin, W. J. (1994). Parameterization of bulk condensation in numerical cloud models. *Journal of the Atmospheric Sciences*, *51*, 1728–1739. https://doi.org/10.1175/15200469(1994)051<1728:POBCIN>2.0.CO;2

Koren, I., Dagan, G., & Altaratz, O. (2014). From aerosol-limited to invigoration of warm convective clouds. *Science*, *2014*(6188), 344–1146. https://doi.org/10.1126/science.1252595

Koren, I., Kaufman, Y. J., Rosenfeld, D., Remer, L. A., & Rudich, Y. (2005). Aerosol invigoration and restructuring of Atlantic convective clouds. *Geophysical Research Letters*, *32*(14), L14828. https://doi.org/10.1029/2005GL023187

Lebo, Z. (2018). A numerical investigation of the potential effects of aerosol-induced warming and updraft width and slope on updraft intensity in deep convective clouds. *Journal of the Atmospheric Sciences*, *75*(2), 535–554. https://doi.org/10.1175/JAS-D-16-0368.1

Lebo, Z. J., & Morrison, H. (2014). Dynamical effects of aerosol perturbations on simulated idealized squall lines. *Monthly Weather Review*, *142*(3), 991–1009. https://doi.org/10.1175/MWR-D-13-00156.1

Lebo, Z. J., Morrison, H., & Seinfeld, J. H. (2012). Are simulated aerosol-induced effects on deep convective clouds strongly dependent on saturation adjustment? *Atmospheric Chemistry and Physics*, *12*(20), 9941–9964. https://doi.org/10.5194/acp-12-9941-2012

Lebo, Z. J., & Seinfeld, J. H. (2011). Theoretical basis for convective invigoration due to increased aerosol concentration. *Atmospheric Chemistry and Physics*, *11*, 5407–5429. https://doi.org/10.5194/acp-11-5407-2011

Lee, S. S. (2012). Effect of aerosol on circulations and precipitation in deep convective clouds. *Journal of the Atmospheric Sciences*, *69*(6), 1957–1974. https://doi.org/10.1175/JAS-D-11-0111.1

Lee, Y., Kummerow, C. D., & Ebert-Uphoff, I. (2021). Applying machine learning methods to detect convection using Geostationary Operational Environmental Satellite-16 (GOES-16) advanced baseline imager (ABI) data. *Atmospheric Measurement Techniques*, *14*(4), 2699–2716. https://doi.org/10.5194/amt-14-2699-2021

Li, Q.-Q., Guo, Y.-T., Yang, J.-Y., & Liang, C.-S. (2023). Review on main sources and impacts of urban ultrafine particles: Traffic emissions, nucleation, and climate modulation. *Atmospheric Environment*, *19*, 100221. https://doi.org/10.1016/j.aeaoa.2023.100221

Li, X., Zhang, Q., Fan, J., & Zhang, F. (2021). Notable contributions of aerosols to the predictability of hail precipitation. *Geophysical Research Letters*, *48*(11), e2020GL091712. https://doi.org/10.1029/2020gl091712

Li, Z., Lau, W. K., Ramanathan, V., Wu, G., Ding, Y., Manoj, M. G., et al. (2016). Aerosol and monsoon interactions in Asia. *Reviews of Geophysics*, *54*(4), 866–929. https://doi.org/10.1002/2015RG000500

Li, Z., Niu, F., Fan, J., Liu, Y., Rosenfeld, D., & Ding, Y. (2011). Long-term impacts of aerosols on the vertical development of clouds and precipitation. *Nature Geoscience*, *4*(12), 888–894. https://doi.org/10.1038/ngeo1313

Li, Z., Wang, Y., Guo, J., Zhao, C., Cribb, M. C., Dong, X., et al. (2019). East Asian Study of Tropospheric Aerosols and their Impact on Regional Clouds, Precipitation, and Climate (EAST-AIRCPC). *Journal of Geophysical Research: Atmospheres*, *124*(23), 13026–13054. https://doi.org/10.1029/2019JD030758

Lin, Y., Fan, J., Jeong, J., Zhang, Y., Homeyer, C. R., & Wang, J. (2021). Urbanization-induced land and aerosol impacts on storm propagation and hail characteristics. *Journal of the Atmospheric Sciences*, *78*(3), 925–947. https://doi.org/10.1175/JAS-D-20-0106.1

Liu, F., Mao, F., Rosenfeld, D., Pan, Z., Zang, L., Zhu, Y., et al. (2022). Opposing comparable large effects of fine aerosols and coarse sea spray on marine warm clouds. *Communications Earth & Environment*, *3*(1), 232. https://doi.org/10.1038/s43247-022-00562-y

Loftus, A. M., & Cotton, W. R. (2014). Examination of CCN impacts on hail in a simulated supercell storm with triple-moment hail bulk microphysics. *Atmospheric Research*, *147–148*, 183–204. https://doi.org/10.1016/j.atmosres.2014.04.017

Manoj, M. G., Lee, S.-S., & Li, Z. (2021). Competing aerosol effects in triggering deep convection over the Indian Region. *Climate Dynamics*, *56*(5–6), 1815–1835. https://doi.org/10.1007/s00382-020-05561-3

Marinescu, P. J., van den Heever, S. C., Heikenfeld, M., Barrett, A. I., Barthlott, C., Hoose, C., et al. (2021). Impacts of varying concentrations of cloud condensation nuclei on deep convective cloud updrafts – A multimodel assessment. *Journal of the Atmospheric Sciences*, *78*(4), 1147–1172. https://doi.org/10.1175/JAS-D-20-0200.1

Meinrat, O. A. (2007). Aerosols before pollution. *Science*, *315*(5808), 50–51. https://doi.org/10.1126/science.1136529

Miltenberger, A. K., Field, P. R., Hill, A. A., Shipway, B. J., & Wilkinson, J. M. (2018). Aerosol–cloud interactions in mixed-phase convective clouds – Part 2: Meteorological ensemble. *Atmospheric Chemistry and Physics*, *18*(14), 10593–10613. https://doi.org/10.5194/acp-18-10593-2018

Morrison, H. (2012). On the robustness of aerosol effects on an idealized supercell storm simulated with a cloud system resolving model. *Atmospheric Chemistry and Physics*, *12*(16), 7689–7705. https://doi.org/10.5194/acp-12-7689-2012

Morrison, H., & Grabowski, W. W. (2013). Response of tropical deep convection to localized heating perturbations: Implications for aerosol-induced convective invigoration. *Journal of the Atmospheric Sciences*, *70*(11), 3533–3555. https://doi.org/10.1175/jas-d-13-027.1

Nishant, N., Sherwood, S. C., & Geoffroy, O. (2019). Aerosol-induced modification of organised convection and top-of-atmosphere radiation. *npj Climate and Atmospheric Science*, *2*(1), 33. https://doi.org/10.1038/s41612-019-0089-1

Öktem, R., Romps, D. M., & Varble, A. C. (2023). No warm-phase invigoration of convection detected during GoAmazon. *Journal of the Atmospheric Sciences*, *80*(10), 2345–2364. https://doi.org/10.1175/JAS-D-22-0241.1

Oza, A. (2023). Reproducibility trial: 246 biologists get different results from same data sets. *Nature*, *622*(7984), 677–678. https://doi.org/10.1038/d41586-023-03177-1

Peng, J., Li, Z., Zhang, H., Liu, J., & Cribb, M. (2016). Systematic changes in cloud radiative forcing with aerosol loading for deep clouds in the tropics. *Journal of the Atmospheric Sciences*, *73*(1), 231–249. https://doi.org/10.1175/JAS-D-15-0080.1

Peters, J. M. (2022). The influence of shear on deep convection initiation. Part I: Theory. *Journal of the Atmospheric Sciences*, *79*, 1669–1689. https://doi.org/10.1175/JAS-D-21-0145

Peters, J. M., Lebo, Z. J., Chavas, D. R., & Su, C.-Y. (2023). Entrainment makes pollution more likely to weaken deep convective updrafts than invigorate them. *Geophysical Research Letters*, *50*(12), e2023GL103314. https://doi.org/10.1029/2023GL103314

Petters, M. D., & Kreidenweis, S. M. (2007). A single parameter representation of hygroscopic growth and cloud condensation nucleus activity. *Atmospheric Chemistry and Physics*, *7*(8), 1961–1971. https://doi.org/10.5194/acp-7-1961-2007

Phillips, V. T. J., Donner, L. J., & Garner, S. (2007). Nucleation processes in deep convection simulated by a cloud-system-resolving model with double-moment bulk microphysics. *Journal of the Atmospheric Sciences*, *64*(3), 738–761. https://doi.org/10.1175/jas3869.1

Pinsky, M., & Khain, A. P. (2002). Effects of in-cloud nucleation and turbulence on droplet spectrum formation in cumulus clouds. *Quarterly Journal of the Royal Meteorological Society*, *128*(580), 1–33. https://doi.org/10.1256/003590002321042072

Pinsky, M., Mazin, I. P., Korolev, A., & Khain, A. (2013). Supersaturation and diffusional droplet growth in liquid clouds. *Journal of the Atmospheric Sciences*, *70*(9), 2778–2793. https://doi.org/10.1175/jas-d-12-077.1

Pinsky, M., Mazin, I. P., Korolev, A., & Khain, A. (2014). Supersaturation and diffusional droplet growth in liquid clouds: Polydisperse spectra. *Journal of Geophysical Research: Atmospheres*, *119*(22), 12872–12887. https://doi.org/10.1002/2014jd021885

Pöschl, U., Martin, S. T., Sinha, B., Chen, Q., Gunthe, S. S., Huffman, J. A., et al. (2010). Rainforest aerosols as biogenic nuclei of clouds and precipitation in the Amazon. *Science*, *329*(5998), 1513–1516. https://doi.org/10.1126/science.1191056

Prabha, T. V., Khain, A., Maheshkumar, R. S., Pandithurai, G., Kulkarni, J. R., Konwar, M., & Goswami, B. N. (2011). Microphysics of pre-monsoon and monsoon clouds as seen from in situ measurements during the Cloud Aerosol Interaction and Precipitation Enhancement Experiment (CAIPEEX). *Journal of the Atmospheric Sciences*, *68*(9), 1882–1901. https://doi.org/10.1175/2011jas3707.1

Prabhakaran, T. V., Murugavel, P., Konwar, M., Malap, N., Gayatri, K., Dixit, S., et al. (2023). CAIPEEX: Indian Cloud Seeding Scientific Experiment. *Bulletin of the American Meteorological Society*, *104*(11), E2095–E2120. https://doi.org/10.1175/BAMS-D-21-0291.1

Riehl, H., & Malkus, J. S. (1958). On the heat balance in the equatorial trough zone. *Geophysica*, *6*, 503–538.

Romps, D. M., Latimer, K., Zhu, Q., Jurkat-Witschas, T., Mahnke, C., Prabhakaran, T., et al. (2023). Air pollution unable to intensify storms via warm-phase invigoration. *Geophysical Research Letters*, *50*(2), e2022GL100409. https://doi.org/10.1029/2022GL100409

Rosenfeld, D., Lohmann, U., Raga, G. B., O'Dowd, C. D., Kulmala, M., Fuzzi, S., et al. (2008). Flood or drought: How do aerosols affect precipitation? *Science*, *321*(5894), 1309–1313. https://doi.org/10.1126/science.1160606

Rotunno, R., Klemp, J. B., & Weisman, M. L. (1988). A theory for strong, long-lived squall lines. *Journal of the Atmospheric Sciences*, *45*(3), 463–485. https://doi.org/10.1175/1520-0469(1988)045<0463:atfsll>2.0.co;2

Sarangi, C., Kanawade, V. P., Tripathi, S. N., Thomas, A., & Ganguly, D. (2018). Aerosol-induced intensification of cooling effect of clouds during Indian summer monsoon. *Nature Communications*, *9*(1), 3754. https://doi.org/10.1038/s41467-018-06015-5

Sarangi, C., Tripathi, S. N., Kanawade, V. P., Koren, I., & Pai, D. S. (2017). Investigation of the aerosol–cloud–rainfall association over the Indian summer monsoon region. *Atmospheric Chemistry and Physics*, *17*(8), 5185–5204. https://doi.org/10.5194/acp-17-5185-2017

Seinfeld, J. H., & Pandis, S. N. (1997). *Atmospheric chemistry and physics, from air pollution to climate change*. John Wiley & Sons.

Journal of Geophysical Research: Atmospheres 10.1029/2024JD041931

Shaw, R. A. (2000). Supersaturation intermittency in turbulent clouds. *Journal of the Atmospheric Sciences*, *57*(20), 3452–3456. https://doi.org/10.1175/1520-0469(2000)057<3452:SIITC>2.0.CO;2

Sheffield, A. M., Saleeby, S. M., & van den Heever, S. C. (2015). Aerosol-induced mechanisms for cumulus congestus growth. *Journal of Geophysical Research: Atmospheres*, *120*(17), 8941–8952. https://doi.org/10.1002/2015JD023743

Shrivastava, M., Fan, J., Zhang, Y., Rasool, Q., Zhao, B., Shen, J., et al. (2024). Intense formation of secondary ultrafine particles from Amazonian vegetation fires and their invigoration of deep convective clouds and precipitation. *One Earth*, *7*(6), 1029–1043. https://doi.org/10.1016/j.oneear.2024.05.015

Siebert, H., & Shaw, R. A. (2017). Supersaturation fluctuations during the early stage of cumulus formation. *Journal of the Atmospheric Sciences*, *74*(4), 975–988. https://doi.org/10.1175/JAS-D-16-0115.1

Storer, R. L., & van den Heever, S. C. (2013). Microphysical processes evident in aerosol forcing of tropical deep convective clouds. *Journal of the Atmospheric Sciences*, *70*, 430–446. https://doi.org/10.1175/JAS-D-12-076.1

Tao, W., & Li, X. (2016). The relationship between latent heating, vertical velocity, and precipitation processes: The impact of aerosols on precipitation in organized deep convective systems. *Journal of Geophysical Research: Atmospheres*, *121*(11), 6299–6320. https://doi.org/10.1002/2015JD024267

Tao, W.-K., Chen, J.-P., Li, Z., Wang, C., & Zhang, C. (2012). Impact of aerosols on convective clouds and precipitation. *Reviews of Geophysics*, *50*(2), RG2001. https://doi.org/10.1029/2011RG000369

Thornton, J. A., Virts, K. S., Holzworth, R. H., & Mitchell, T. P. (2017). Lightning enhancement over major oceanic shipping lanes. *Geophysical Research Letters*, *44*(17), 9102–9111. https://doi.org/10.1002/2017GL074982

Varble, A. (2018). Erroneous attribution of deep convective invigoration to aerosol concentration. *Journal of the Atmospheric Sciences*, *75*(4), 1351–1368. https://doi.org/10.1175/jas-d-17-0217.1

Varble, A. C., Igel, A. L., Morrison, H., Grabowski, W. W., & Lebo, Z. J. (2023). Opinion: A critical evaluation of the evidence for aerosol invigoration of deep convection [preprint]. *EGUsphere*. https://doi.org/10.5194/egusphere-2023-938

Wang, C. (2005). A modeling study of the response of tropical deep convection to the increase of cloud condensation nuclei concentration. 1. Dynamics and microphysics. *Journal of Geophysical Research*, *110*(D21), D21211. https://doi.org/10.1029/2004JD005720

Wang, Q., Li, Z., Guo, J., Zhao, C., & Cribb, M. C. (2018). The climate impact of aerosols on the lightning flash rate: Is it detectable from long-term measurements? *Atmospheric Chemistry and Physics*, *18*(17), 12797–12816. https://doi.org/10.5194/acp-18-12797-2018

Wang, Y., Fan, J., Zhang, R., Leung, L. R., & Franklin, C. (2013). Improving bulk microphysics parameterizations in simulations of aerosol effects. *Journal of Geophysical Research: Atmospheres*, *118*(11), 5361–5379. https://doi.org/10.1002/jgrd.50432

Wang, Y., Lee, K.-H., Lin, Y., Levy, M., & Zhang, R. (2014). Distinct effects of anthropogenic aerosols on tropical cyclones. *Nature Climate Change*, *4*(5), 368–373. https://doi.org/10.1038/nclimate2144

Wang, Y., Wang, M., Zhang, R., Ghan, S. J., Lin, Y., Hu, J., et al. (2014). Assessing the effects of anthropogenic aerosols on Pacific storm track using a multiscale global climate model. *Proceedings of the National Academy of Sciences*, *111*(19), 6894–6899. https://doi.org/10.1073/pnas.1403364111

Wright, C. J., Thornton, J. A., Jaeglé, L., Cao, Y., Zhu, Y., Liu, J., et al. (2024). Lightning declines over shipping lanes following regulation of fuel sulfur emissions. *arXiv:2408.07207 [physics.ao-ph]*. https://doi.org/10.48550/arXiv.2408.07207

Yan, H., Li, Z., Huang, J., Cribb, M. C., & Liu, J. (2014). Long-term aerosol-mediated changes in cloud radiative forcing of deep clouds at the top and bottom of the atmosphere over the Southern Great Plains. *Atmospheric Chemistry and Physics*, *14*, 7113–7124. https://doi.org/10.5194/acp-14-7113-2014

Yang, X., & Li, Z. (2014). Increases in thunderstorm activity and relationships with air pollution in southeast China. *Journal of Geophysical Research: Atmospheres*, *119*(4), 1835–1844. https://doi.org/10.1002/2013JD021224

Zang, L., Rosenfeld, D., Pan, Z., Mao, F., Zhu, Y., Lu, X., & Gong, W. (2023). Observing aerosol primary convective invigoration and its meteorological feedback. *Geophysical Research Letters*, *50*(12), e2023GL104151. https://doi.org/10.1029/2023GL104151

Zhang, G. J., & McFarlane, N. A. (1995). Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model. *Atmosphere-Ocean*, *33*(3), 407–446. https://doi.org/10.1080/07055900.1995.9649539

Zhang, R., Li, G., Fan, J., Wu, D. L., & Molina, M. J. (2007). Intensification of Pacific storm track linked to Asian pollution. *Proceedings of the National Academy of Sciences*, *104*(13), 5295–5299. https://doi.org/10.1073/pnas.0700618104

Zhang, Y., Fan, J., Li, Z., & Rosenfeld, D. (2021). Impacts of cloud microphysics parameterizations on simulated aerosol-cloud-interactions for deep convective clouds over Houston. *Atmospheric Chemistry and Physics*, *21*(4), 2363–2381. https://doi.org/10.5194/acp-21-2363-2021

Zhao, C., Lin, Y., Wu, F., Wang, Y., Li, Z., Rosenfeld, D., & Wang, Y. (2018). Enlarging rainfall area of tropical cyclones by atmospheric aerosols. *Geophysical Research Letters*, *45*(16), 8604–8611. https://doi.org/10.1029/2018GL079427

Zhu, P., Dudhia, J., Field, P., Wapler, K., Fridlind, A., Varble, A., et al. (2012). A limited area model (LAM) intercomparison study of a TWP-ICE active monsoon mesoscale convective event. *Journal of Geophysical Research*, *117*, D11. https://doi.org/10.1029/2011JD016447

Erratum

The originally published version of this article contained some errors. The eighth and ninth sentences of the second paragraph of Section 5.2 have been corrected as follows: "The only difference between Varble18_Prod and Varble18_copy is that the CAPE in Varble18_Prod is calculated by us as a sanity check. Since Varble18's data lack the CAPE values needed for our analysis (clear-sky times within 30 min before the cloud object), we calculate our CAPE values." The Acknowledgments have been revised as follows: "Argonne National Laboratory's work was supported by the U.S. Department of Energy (DOE), Office of Science's Biological and Environmental Research (BER) Atmospheric System Research, under Contract DE-AC02-06CH11357. ZL was supported by the U.S. DOE BER Atmospheric System Research under Award Number DE-SC0022919 and by the National Science Foundation under Award Number AGS2126098. Data were obtained from the Atmospheric Radiation Measurement (ARM) User Facility, a U.S. DOE Office of Science user facility managed by the Office of BER." This may be considered the authoritative version of record.