



Aerosols perturb mixed-phase microphysics and anvil radiative effects in deep convection

Maor Sela¹, Mathilde Ritman¹, Sadhitro De¹, and Philip Stier¹

¹Department of Physics, University of Oxford, Oxford, UK

Correspondence: Maor Sela (maor.sela@physics.ox.ac.uk)

Abstract. Deep convective clouds (DCCs) transport heat, water, and momentum, affecting Earth’s radiative balance and water cycle. Aerosol particles influence these properties as cloud condensation nuclei, but how anthropogenic aerosols modify mixed-phase cloud microphysics and radiative effects remains unclear. Here, we investigate two sea-breeze-driven convection cases over Houston using high-resolution ICON-HAM-lite simulations, exploring present-day, cleaner, and more polluted conditions by scaling anthropogenic aerosol concentrations and emissions by factors of 3 and 0.3. We ran 10-member ensembles of 24-hour simulations with perturbed initial temperatures. As previous studies have shown, increased aerosols raise cloud droplet numbers and suppress warm rain formation. We further show that aerosols shift microphysics pathways by reducing riming and enhancing evaporation, ice nucleation, and depositional growth. These changes maintain more condensate in cloud ice and snow while reducing transfer to graupel and hail. Consequently, pollution strengthens both longwave warming and shortwave cooling, with shortwave cooling dominating in active convection. The principal responses are consistent across both cases, but their vertical and radiative expression varies with storm organisation: more organised storms produce larger anvils and more sustained cooling, while isolated storms have more heterogeneous cloud structures. In the most aerosol-rich simulations, the net radiative effect is also less robust because of greater ensemble spread during late-stage anvil dissipation. Overall, these results highlight the importance of competing mixed-phase microphysical processes when assessing aerosol effects on deep convective clouds.

1 Introduction

Deep convective clouds (DCCs) are an important component of the atmospheric system as they redistribute heat, moisture and momentum, vertically and horizontally, while also exerting a strong influence on precipitation and the Earth’s energy balance (Emanuel, 1997; Trenberth et al., 2009; Bony et al., 2015; Morrison et al., 2020). Although DCC vertical and horizontal extent can reach tens and hundreds of kilometres (respectively) in some cases, their bulk properties are controlled by processes that take place in the micron to millimetre scale, referred to as cloud microphysics. These processes govern cloud characteris-



tics by controlling hydrometeor formation, growth, sedimentation, and evaporation, making DCCs intrinsically difficult to be represented in models (Khain et al., 2015; Morrison et al., 2020).

25 Atmospheric aerosols affect these microphysical processes by altering cloud droplet number concentration (CDNC) and droplet size distributions. Aerosol particles that act as cloud condensation nuclei (CCN) can increase CDNC, producing smaller and narrower droplet spectra and thereby suppressing collision–coalescence and delaying warm-rain formation (Andreae et al., 2004; Rosenfeld et al., 2008; Tao et al., 2012; Fan et al., 2016). While this CCN–CDNC–droplet-size response is well established for warm clouds, its extension to deep convection remains uncertain because the resulting microphysical and dynamical
 30 responses depend strongly on the thermodynamic and kinematic environment and are sensitive to the representation of cloud microphysics in numerical models (Fan et al., 2016; White et al., 2017; Varble, 2018; Marinescu et al., 2021; Saleeby et al., 2025).

A key uncertainty lies in the mixed-phase and ice-phase processes through which aerosol perturbations may propagate in deep convection. A commonly proposed pathway for the propagation of these perturbations is that the suppression of warm-
 35 rain formation retains more condensate for transport into the mixed-phase region, where the initial perturbation may propagate through coupled microphysical processes, including riming, freezing and ice nucleation, and vapour deposition (Storer and van den Heever, 2013; Khain et al., 2015; Glassmeier and Lohmann, 2016; Heikenfeld et al., 2019). However, the magnitude of the effects propagating through this pathway are not uniquely determined by the initial aerosol perturbation, as changes in condensate type and loading, latent heating, and cloud dynamics can reinforce or counteract one another and are strongly
 40 conditioned by the environmental state (Storer and van den Heever, 2013; Varble, 2018). These interactions can shape the bulk properties of the cloud and its resulting interaction with the environment. However, bulk fields alone are insufficient to distinguish which processes dominate; hence, process-rate diagnostics are required to attribute the large-scale effects to the governing mechanisms (Storer and van den Heever, 2013; Heikenfeld et al., 2019; Herbert et al., 2021).

DCC bulk properties also play an important role in determining their radiative impact. Optically thicker anvils generally
 45 reflect more incoming solar radiation, strengthening shortwave (SW) cooling, whereas the longwave (LW) effect depends strongly on cloud-top temperature and emissivity. Consequently, changes in anvil optical properties, cloud-top height (CTH), extent, and persistence can alter the balance between SW cooling and LW warming. These properties evolve throughout the DCC lifecycle and depend nonlinearly on convective intensity, organisation, and outflow (Jones and Stier, 2025; Ritman et al., 2026), such that changes in anvil morphology can substantially alter the resulting CRE (Koren et al., 2010). Aerosol perturba-
 50 tions further complicate this response by modifying cloud microphysics, including ice-particle size and sedimentation, thereby affecting anvil morphology and persistence (Koren et al., 2010; Fan et al., 2013). The resulting radiative response depends strongly on how high-clouds evolve (Herbert et al., 2021). Because these aerosol-induced changes can act differently across the developing, mature, and dissipating stages of convection, and because radiative effects depend nonlinearly on the resulting cloud properties (Koren et al., 2010), understanding aerosol impacts on DCCs requires considering the evolution of the cloud
 55 over its lifecycle rather than only changes in bulk or time-averaged properties.

Motivated by uncertainty in aerosol–ice-phase microphysics pathway interactions and the complex DCC anvil lifecycle, we investigate the impact of anthropogenic aerosol perturbations on deep convection. To isolate the impact of aerosol–DCC in-



teractions, we examine two sea-breeze cases with different background environments and use a cloud-tracking-based analysis. To resolve the complex interplay between the competing microphysical pathways, we use process-rate diagnostics along the convective lifecycle of DCCs, with a high-resolution cloud-resolving model (CRM) coupled to an interactive aerosol module. We aim to identify the microphysical pathways that control the radiative signature of the simulated DCCs and assess whether the aerosol-driven mechanism is robust across the different backgrounds. The manuscript is structured as follows. Section 2 describes the model setup, experimental design, and analysis technique. In Section 3, we first characterise the storms' bulk properties, then follow the aerosol perturbation through the DCCs using in-cloud microphysics, process rates, and cloud radiative effect, while discussing the generalisability of the results, biases, and limitations. Section 4 summarises the findings and highlights their implications.

2 Methodology

2.1 Model description and setup

We use the ICOSahedral Non-hydrostatic (ICON) modelling framework (Zängl et al., 2015; Hohenegger et al., 2023; Segura et al., 2025) coupled to the reduced-complexity aerosol module HAM-lite (Weiss et al., 2025; De and Stier, 2026) in regional configuration (Heinold et al., 2026), hereafter ICON-HAM-lite. The model is configured with a horizontal grid spacing of ~ 600 m, which permits explicit representation of deep convective updrafts, downdrafts, and the associated anvil evolution. The atmosphere and radiation time steps are set to 1 s and 15 min, respectively, to maintain stable model integration within available computational resources. The atmospheric dynamics are coupled to the Jena Scheme for Biosphere-Atmosphere Coupling in Hamburg (JSBACH; Reick et al. (2021)) for the land surface and use the terrain-following Smooth Level Vertical (SLEVE) coordinate system (Leuenberger et al., 2010) in the vertical, with 95 vertical levels extending to 21.5 km. We employ the Rapid Radiative Transfer Model for General circulation model applications-Parallel (RRTMGP; Pincus et al. (2019)) radiation scheme together with the modified Smagorinsky turbulence closure (Dipankar et al., 2015). Cloud microphysical processes are represented by the two-moment bulk microphysics scheme of Seifert and Beheng (2006), which predicts the mass and number evolution of cloud water, ice, rain, snow, graupel and hail (De and Stier, 2026). ICON-HAM-lite prescribes aerosol size, composition, and mixing state and prognoses aerosol number concentration. It employs four hydrophilic modes (carbonaceous, sulfuric, sea salt, and dust), with the carbonaceous and sulfuric modes representing internally mixed organic and black carbon, and sulfate particles (Weiss et al., 2025). These modes are transported, interact with the model parameterisations, and affect radiation and associated diabatic heating. Anthropogenic and biomass-burning emissions are prescribed, whereas sea-salt and dust emissions are calculated interactively and emitted into their corresponding modes. This reduced-complexity representation neglects the explicit evolution of aerosol size and composition through processes such as nucleation, growth, and ageing (Weiss et al., 2025; Heinold et al., 2026). Aerosol-cloud coupling is represented primarily through aerosol activation and its effect on cloud-droplet number, while mixed-phase processes, including riming and depositional ice growth, are represented by the bulk cloud-microphysics scheme and are not directly parameterised as functions of evolving aerosol properties; aerosols are in fact coupled back to the hydrometeor field through scavenging and wet deposition.



Table 1. Prescribed aerosol composition and properties used in the experiment. Geometric median dry radius $\bar{r}$, standard deviation σ , density ρ , and hygroscopicity κ . Number concentration is the only prognostic tracer per mode. Both $\bar{r}$ and σ were prescribed by the TRACER protocol; ρ and κ were kept unchanged as originally used in Weiss et al. (2025).

	Carbonaceous	Sulfuric	Sea salt	Dust
$\bar{r}$ (μm)	$r_{\text{june}} = 0.068, r_{\text{august}} = 0.0875$	$r_{\text{june}} = 0.015, r_{\text{august}} = 0.0245$	0.72	0.6
σ	$\sigma_{\text{june}} = 1.5, \sigma_{\text{august}} = 1.4$	$\sigma_{\text{june}} = 1.5, \sigma_{\text{august}} = 1.8$	2.0	2.0
ρ (kg m^{-3})	2000	1858	2165	2650
κ	0.166	0.464	1.335	0.140

2.2 Experimental design and case classification

We perform three experiments for each of two sea-breeze case studies, following the Tracking Aerosol Convection Interactions Experiment Model Intercomparison Project (TRACER-MIP) protocol (Saleeby and Fan, 2025). The TRACER campaign (Jensen et al., 2023) took place between 2021 and 2022 in the Houston, Texas, area, which was selected to investigate the influence of anthropogenic aerosols on deep convective clouds (see, e.g., Marinescu et al., 2021; Saleeby and Fan, 2025). Following the TRACER-MIP protocol, we focus on two deep-convective cases that took place on 17 June and 7 August 2022. For each case, we run a 10-member ensemble, generated by adding small Gaussian perturbations to the temperature profile with a standard deviation of $\sigma = 0.01$ K. This ensemble design allows us to quantify the spread associated with small differences in the initial state while retaining the same large-scale forcing and aerosol perturbation. The simulations are centred on Houston, Texas, spanning $96.5^\circ - 93.8^\circ$ W and $28.5^\circ - 30.5^\circ$ N. ERA5 reanalysis data from the Copernicus Climate Change Service (Hersbach et al., 2017) provide the initial and lateral boundary conditions, which are updated hourly and linearly interpolated between updates. Both cases are simulated for 24 h, with an additional spin-up duration of 3 h, so the data are analysed from midnight (local time) on 17 June and 7 August.

The TRACER-MIP protocol prescribes two aerosol modes, ultra-fine and accumulation, with diameters and standard deviations derived from surface-based measurements (Saleeby and Fan, 2025; Chen et al., 2025). To translate the ultra-fine and accumulation modes to ICON-HAM-lite native modes, we use the derived radii and standard deviations to set the sulfuric and carbonaceous modes, respectively. The resulting modal properties are summarised in Table 1. Unlike the TRACER-MIP protocol, which constructs synthetic aerosol profiles from these observations, we initialise the aerosol distributions using Copernicus Atmospheric Monitoring Service (CAMS; Inness et al. (2019)) reanalysis data. Fig. 1 shows the domain-mean aerosol initial condition profiles. The carbonaceous-mode number concentrations are comparable to those of the synthetic TRACER-MIP accumulation mode, whereas the sulfuric mode reaches a peak number concentration approximately six times that of the synthetic ultra-fine mode.

Following the TRACER-MIP protocol, we conduct two aerosol perturbation experiments in addition to the baseline simulation (referred to as ref): a cleaner (x03), and a more polluted experiment (x3) for each of the simulated days. We obtain the cleaner and polluted experiments by scaling the carbonaceous and sulfuric aerosol modes by factors of 0.3 and 3, respectively,

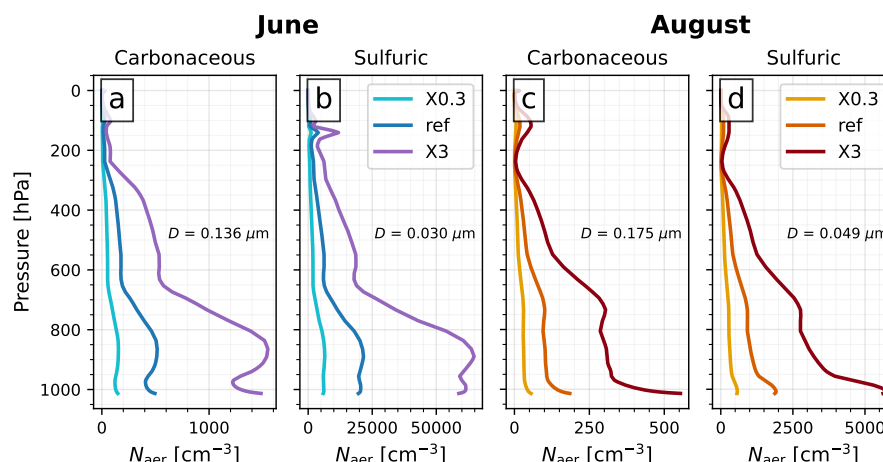


Figure 1. CAMS-derived domain-mean initial state of the aerosol number concentration profiles for the (a and c) carbonaceous and (b and d) sulfuric aerosol modes for the June (blue shades) and August (orange shades) cases, respectively. These profiles are used to define the reference (ref), cleaner (x03), and more polluted (x3) experiments, with the effective modal diameters annotated in each panel. Note differences in number concentration extents.

in both the initial conditions and the lateral boundary conditions. The anthropogenic components of the prescribed emissions, which include sectors such as industry, aviation, energy production, and transport, are scaled by the same factors as the initial and boundary conditions. The emissions data are provided as monthly mean mass fluxes and converted to number fluxes according to the fixed modal properties shown in Table 1. Overall, we run 60 simulations: 10 perturbed initial-condition ensemble members for three aerosol-level experiments in two case studies.

Beyond their different background aerosol conditions, the two cases are characterised by contrasting dynamical and thermodynamic environments. The June case features widespread, more organised convection that forms preferentially near the coastline, with an afternoon sea-breeze-induced thunderstorm. The August case, under relatively cleaner aerosol conditions, features a morning sea-breeze front and an early-afternoon thunderstorm, with more isolated, popcorn-like DCCs (see Fig. 3). This difference is consistent with the pre-convective environments shown in Fig. 2. In June, the highest convective available potential energy (CAPE) values are more spatially localised, and the convective inhibition (CIN) distribution is more confined to specific regions. In August, weaker capping and a broader region of convective instability, combined with stronger but more scattered CAPE, are consistent with more localised and independent convective initiation. These two cases provide contrasting environments in which we evaluate the effects of similar relative aerosol perturbations.

2.3 DCC tracking and lifecycle normalisation

To isolate individual deep convective systems and analyse their lifecycle evolution, we use the *tobac* cloud tracking framework, following the approach of Ritman et al. (2026). Convective updrafts are defined as regions in which vertical velocity exceeds

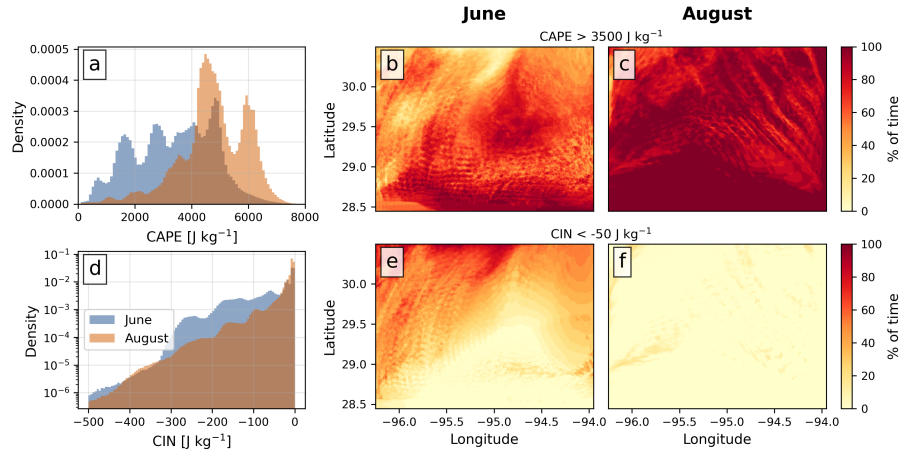


Figure 2. Convective environment of the June and August cases during the 5 h preceding the analysed DCC period. (a) Normalised probability density functions (PDF) of CAPE. (b,c) Fraction of time that CAPE exceeds 3500 J kg^{-1} at each grid point for the June and August cases, respectively. (d) Normalised probability density functions of CIN. (e,f) Fraction of time that CIN is smaller than -50 J kg^{-1} (i.e., stronger capping) at each grid point for the June and August cases, respectively.

1 m s^{-1} . The resulting convective ice towers and their anvil outflows are defined as regions where the total frozen condensate exceeds 0.01 g kg^{-1} . We track the detected updrafts and frozen-condensate clouds separately, using tobac's centroid tracking and a modified cloud overlap test (described in detail in Ritman et al. (2026)). The results of the two tracking operations are then combined such that only ice clouds originating from convective updrafts are retained and termed deep convective clouds.

Fig. 3 illustrates the tracking procedure for the two reference simulations. The overlaid contours show the footprint of the tracked DCCs, which we use to separate the convective systems analysed in the remainder of the paper. On average, we track roughly 30 systems per ensemble member using these tracking criteria, although the exact number varies between members, experiments, and cases.

We define the DCC lifecycle to begin at the onset of glaciation (Heikenfeld et al., 2019), providing a consistent physical reference point for comparing the subsequent growth, mature, and dissipating stages across cases while excluding the preceding initiation phase. To compare systems with different absolute lifetimes, we map each DCC lifecycle onto a common relative time coordinate. For each system, we linearly interpolate the active period onto equally spaced points between 0 and 1. This allows DCCs to be compared at equivalent relative stages of development, rather than mixing systems sampled at different lifecycle phases. However, it removes information about the absolute duration of the systems, which is therefore not interpreted from the normalised composites.

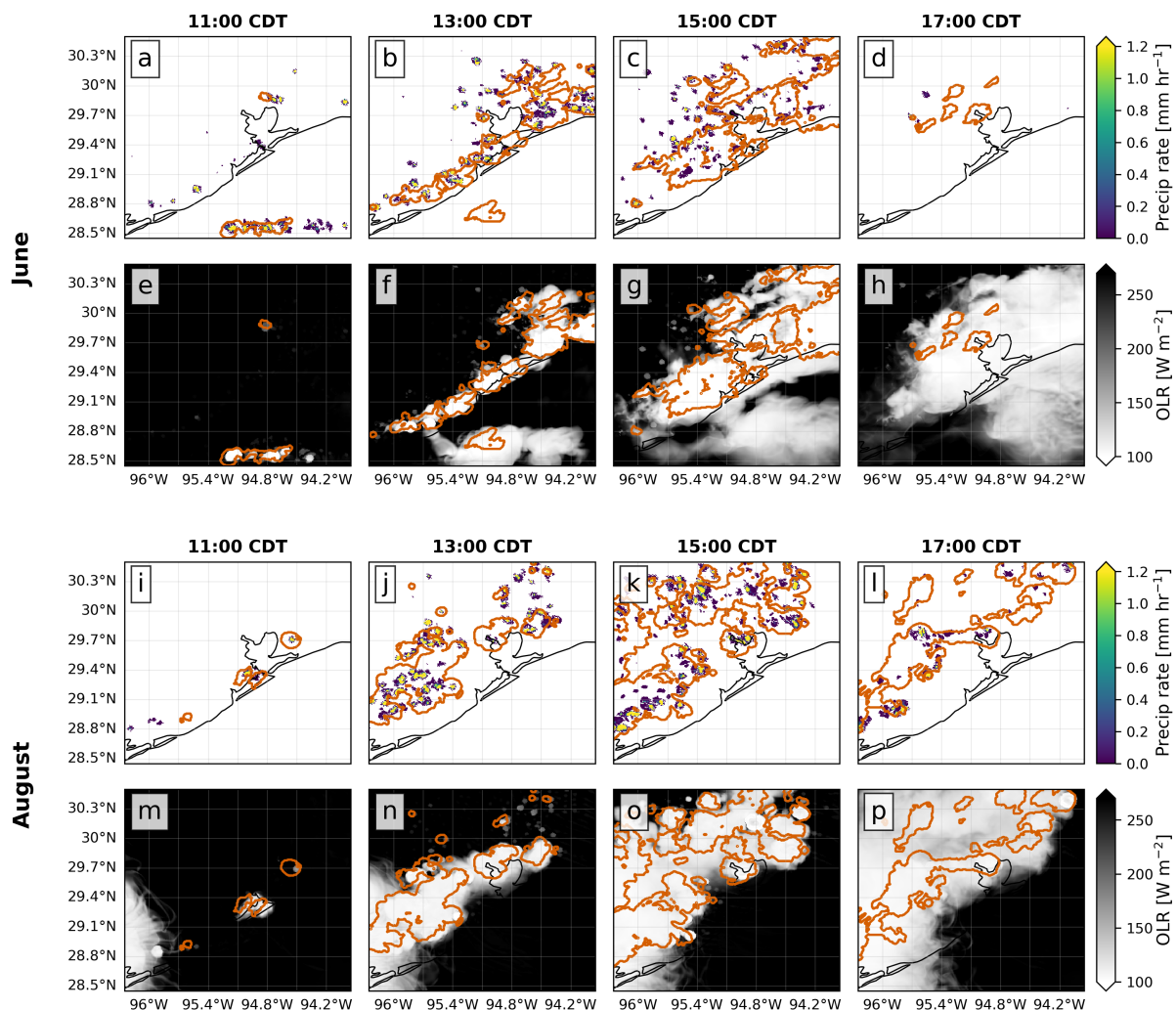


Figure 3. Instantaneous maps of (a-d and i-l) surface precipitation, and (e-h and m-p) OLR during the span of the convective storms on June 17 (a-h) and August 7 (i-p), 2022, as simulated by the reference run. Overlaid contours (orange) indicate tracked DCC cells.

3 Results and discussion

3.1 Domain-mean response to aerosol perturbation

150 We begin by looking at the aerosol effect across the entire domain. Increasing aerosol concentration produces more numerous cloud droplets (Fig. 4c,e). At the same time, the domain-mean low-level cloud-water mass remains broadly similar across the aerosol experiments (Fig. 4d,f). Thus, despite the strong response in droplet number, domain-mean low-level cloud-water mass shows no corresponding signal. This suggests a redistribution of the available liquid water among a larger population of

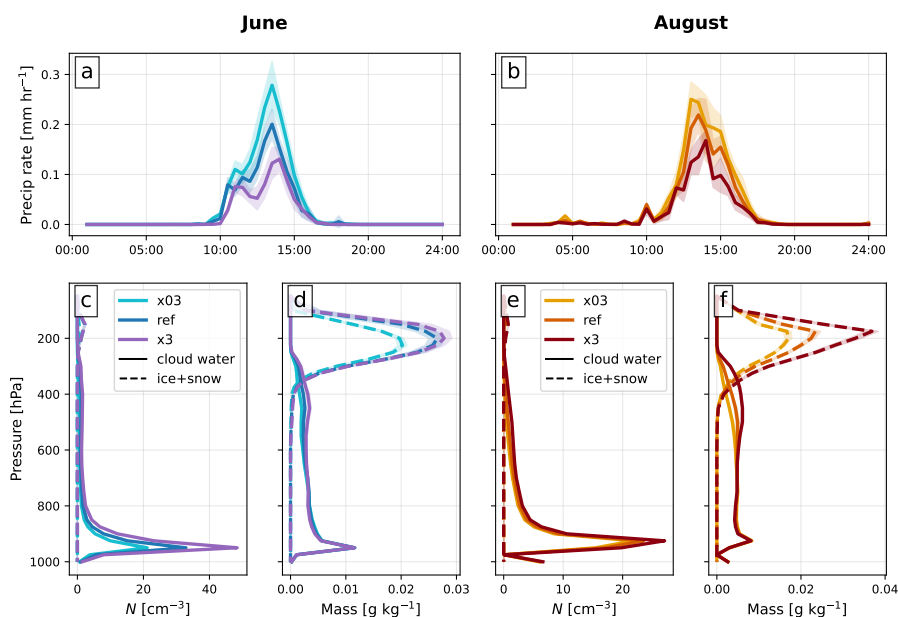


Figure 4. Domain mean of (a and b) surface precipitation, (c and e) cloud droplet (solid) and ice crystal (dashed) number concentration, and (d and f) cloud water (solid) and ice (dashed) mass mixing ratio. Lines show the ensemble mean, and shading indicates the ensemble standard deviation for both the June (blue shades) and August (orange shades) cases.

droplets, although changes in in-cloud liquid water may be obscured by the domain averaging. The increase in number concentration therefore implies a smaller mean droplet size, which is consistent with the activation response to enhanced aerosol concentrations expected based on past work (see, e.g., Storer and van den Heever, 2013; Fan et al., 2013; Simpkins, 2018). Additionally, we find reduced surface precipitation rates (Fig. 4a,b), consistent with the expected suppression of collision-coalescence, as smaller droplets are less efficiently converted to precipitation through autoconversion and accretion (Pinsky et al., 2007; Barthlott et al., 2022). Thus, both cases exhibit warm-rain suppression as expected by the literature.

Since warm-rain formation is suppressed, a larger fraction of the liquid condensate is expected to remain in the cloud and be transported to higher altitudes, which we find to be true. In both cases, the polluted simulations retain more cloud water above the 600 hPa level, meaning more water is transported above the melting layer (Fig. 4d,f), creating an enhanced reservoir of supercooled liquid water in the mixed-phase region. Once lofted into this region, the liquid condensate can follow several pathways, including freezing, collection through riming, evaporation, or indirectly contribute to deposition onto ice and snow. The simultaneous increase in upper-level cloud water and ice mass in Fig. 4d,f therefore suggests that the effect of warm-rain suppression extends into the mixed-phase region, altering the subsequent partitioning of condensate among liquid and frozen hydrometeors, as also suggested by, e.g., Rosenfeld et al. (2014), Heikenfeld et al. (2019), Dagan et al. (2020), Dagan and Stier (2020), Wu and Peng (2022).



These results motivate a process-level investigation. Given the increase in supercooled liquid and ice mass aloft, the question is therefore how this additional condensate is partitioned among competing mixed-phase growth pathways. In clean conditions, larger droplets may favour collection by ice particles and riming, whereas in polluted conditions the smaller droplet population may favour a stronger vapour-depositional growth pathway, consistent with a Wegener-Bergeron-Findeisen (WBF; Storelvmo and Tan (2015)) type mechanism. However, the domain mean cannot distinguish these pathways unambiguously, because it combines shallow convection, deep convective cores, stratiform cloud, and different stages of storm evolution. We therefore focus the next part of the analysis on Lagrangian-tracked convective cells (see Section 2.3), starting with their bulk hydrometeor structure and morphology.

3.2 Bulk DCC properties and morphology response

In both cases, the storm's broad evolution is largely fixed across aerosol perturbations. Fig. 5 shows the time evolution of DCC vertical frozen-hydrometeor occurrence, defined where the mass mixing ratio exceeds 0.1 mg kg^{-1} . Across all the experiments, the systems develop an initial deep convective phase, with graupel and hail extending through the mixed-phase region, followed by the persistence of upper-level cloud ice and snow, which characterise the anvil cloud. For each case, the timing and vertical extent of the frozen hydrometeor regions remain broadly similar across aerosol experiments. This indicates that the aerosol perturbation does not fundamentally alter bulk storm evolution, such as storm size and lifetime, but instead primarily affects hydrometeor partitioning. The absence of a clear aerosol-driven change in bulk storm evolution is consistent with recent observations over Houston, where relationships between aerosol loading and convective cell area weakened after accounting for the synoptic environment (Al-Jabri et al., 2025).

Despite similar bulk evolution, the DCC morphology is sensitive to aerosol perturbations. The regions occupied by graupel and hail become less extensive in the polluted simulations, while cloud ice and snow occupy a larger fraction of the upper cloud and anvil region. The DCC mean vertical profiles show this more directly, where increasing aerosol loading increases droplet number concentration (Fig. 6c,i), enhances cloud water in the mixed-phase region (Fig. 6a,g), increases ice number aloft (Fig. 6d,j), and reduces rain and hail mass through much of the column (Fig. 6e,f,k,l). Similar aerosol-induced shifts in hydrometeor partitioning and mixed-phase processes have been reported in previous process-resolving studies (see, e.g., Heikenfeld et al., 2019; Wu and Peng, 2022). The response is consistent with the domain-mean profiles (Fig. 4), but is more pronounced. This stronger aerosol signal within DCCs relative to the domain mean is also evident in the CDNC response. The maximum polluted-to-clean ratio reaches ~ 4 -fold within DCCs, compared to ~ 2.5 -fold in the domain mean (Fig. 6c,i vs. Fig. 4c,e). Because warm clouds are abundant across the domain, one can expect a discernible aerosol effect in the domain mean. Given the aerosol-rich Houston environment (Fig. 1), the weaker domain mean response is consistent with an updraft-limited activation regime (Reutter et al., 2009), in which CDNC becomes more sensitive to updraft velocity than to CCN concentration when CCN are abundant. Weaker vertical motions outside the tracked DCCs would limit supersaturation and the activation of additional CCN, compressing the domain-mean CDNC response relative to that within DCCs. Differences in DCC occurrence among the aerosol experiments may also contribute to this contrast.

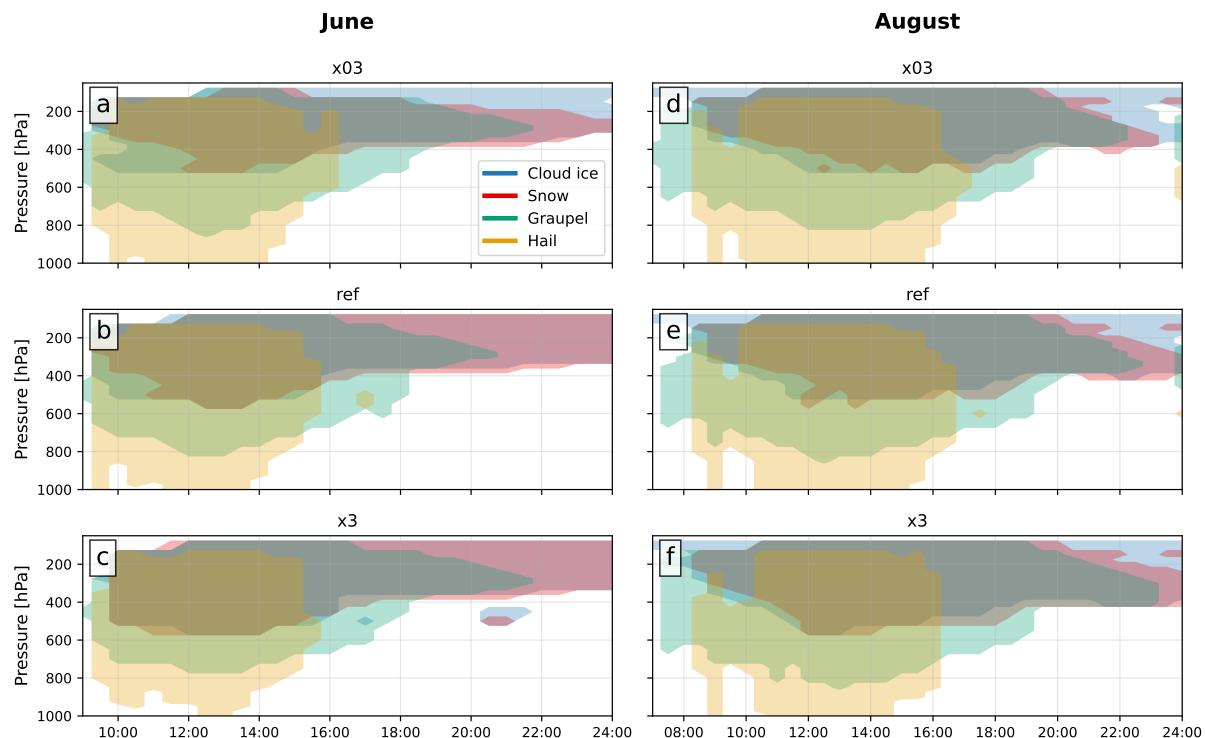


Figure 5. Pressure-time evolution of the ice habits distribution for the (a and d) reference, (b and e) clean, and (c and f) polluted cases for the June and August cases (left- and right-hand side panels, respectively). Filled contours indicate regions where the mixing ratio of cloud ice (blue), snow (red), graupel (green), and hail (yellow) exceeds 0.1 mg kg^{-1} , averaged across members.

Note that the two cases diverge in the low-level cloud-water response. In the June case, cloud water increases with aerosols (Fig. 6a), whereas in the August case the mean decreases (Fig. 6g), although this response exhibits greater variability. This uncertain behaviour is likely a manifestation of the underlying difference in the convective case, consistent with previous studies showing that aerosol responses depend strongly on the prevailing cloud regime and environmental conditions (e.g., Khain et al., 2005; Dagan and Stier, 2020). The more isolated, popcorn-like convection in August produces a cloud population more likely to span a wider range of developmental stages and cloud-top heights, so the mean profile integrates contributions from systems that are less vertically aligned than in the organised June case. Smaller, more isolated convective systems may also be more susceptible to environmental mixing because of their greater cloud-edge fraction (e.g., Tsai and Wu, 2017; Abramian et al., 2025). Although the near-cloud-base cloud-water response is thus more uncertain in August and can differ in sign from the June case, the CDNC response remains clear: polluted conditions still produce much higher CDNC, implying smaller mean droplets, while rain mass remains reduced relative to the clean case. Aloft, the August case shows a less distinct anvil-shaped ice profile, with ice extending upward from the melting layer and no clear separation between the aerosol cases in the mixed-phase region (Fig. 6h); these features are consistent with a less organised, congestus-like DCC population spanning

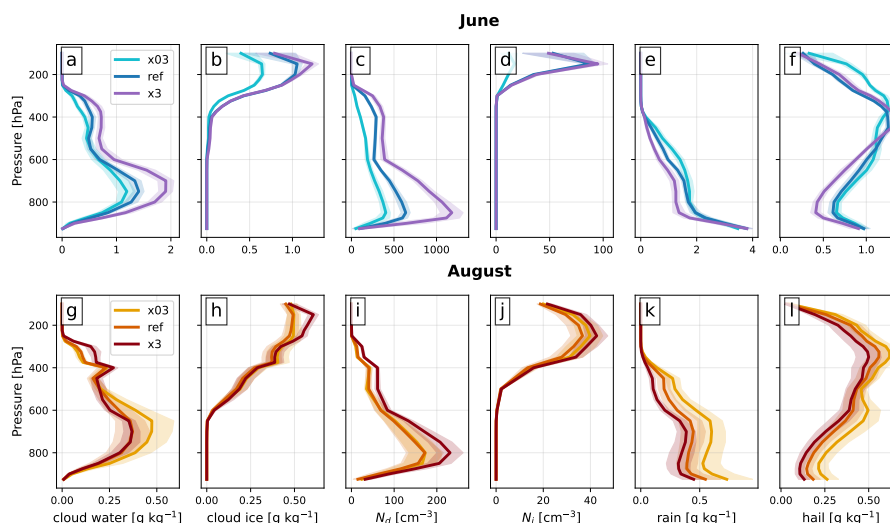


Figure 6. In-DCC mean vertical profiles of cloud hydrometeor properties of (a and g) cloud water mass mixing ratio, (b and h) cloud ice mass mixing ratio, (c and i) cloud droplet number concentration, (d and j) ice crystal number concentration, (e and k) rain mass mixing ratio, and (f and l) hail mass mixing ratio for the June (blue shades) and August (orange shades) cases. Lines show the ensemble mean, and shading indicates the ensemble standard deviation.

215 a broader range of developmental stages and vertical structures. Such case-dependent aerosol responses are consistent with TRACER observations of isolated deep convection, for which aerosol effects on cloud development and precipitation varied with convective and environmental conditions (Wang et al., 2025). Observational analyses over Houston further indicate that apparent relationships between aerosol loading and convective morphology can weaken after accounting for the synoptic environment (Al-Jabri et al., 2025), highlighting the difficulty of separating aerosol effects from meteorological controls.

220 These differences in hydrometeor partitioning suggest altered competition among mixed-phase growth processes. In cleaner conditions, larger droplets are more readily collected by ice particles, favouring riming growth into denser, faster-sedimenting graupel and hail. In polluted conditions, the smaller droplet population is less efficiently collected because small droplets have lower inertial collision efficiencies and are more likely to follow the airflow around falling ice particles (Pruppacher, 2010; Lamb and Verlinde, 2011). This reduced collection efficiency is consistent with a larger reservoir of suspended cloud water,

225 cloud ice, and snow, and with a reduced transfer of condensate into precipitating hydrometeor classes.

These hydrometeor profiles therefore support the interpretation that aerosol loading shifts the storm microphysics away from efficient riming and sedimentation and toward pathways that maintain more condensate in smaller liquid and ice categories. However, morphology and hydrometeor profiles alone cannot determine by which microphysical pathways this shift is caused. Therefore, we next analyse the microphysical process-rate diagnostics over the normalised lifecycle of in-cloud DCC systems.



230 3.3 Microphysical competition

To identify the processes responsible for the aerosol-driven hydrometeor differences shown above, we analyse the DCC process rates over the normalised cloud lifecycle, so that differences between aerosol cases reflect lifecycle-relative process evolution rather than differences in the timing of individual systems. Fig. 7 shows the polluted-minus-clean (i.e., x_3 minus x_{03}) process-rate differences for the June and August cases. Positive values (red) indicate processes that are enhanced in the polluted simulations, while negative values (blue) indicate processes that are enhanced in the clean simulations. The process-rate diagnostics reveal a changing balance between two competing mixed-phase pathways as aerosol loading increases. The first pathway involves riming, whereby supercooled cloud droplets and rain are collected by falling ice particles and transferred into denser hydrometeor classes, favouring graupel and hail growth and, ultimately, sedimentation. With increasing aerosol loading, riming of cloud droplets and rain becomes less pronounced throughout much of the mixed-phase region (Fig. 7a,d,g,j). This reduction is consistent with the smaller graupel and hail regions in the polluted experiments shown in Fig. 5 and the reduced hail mass in the DCC profiles (Fig. 6f,l). The response is consistent with the size dependence of collection and riming efficiencies, whereby larger droplets have greater inertia and are more efficiently collected by falling ice particles (e.g., Pruppacher, 2010; Lamb and Verlinde, 2011; Wang and Ji, 2000; Hashino et al., 2014; Seifert and Beheng, 2006). Aerosol-induced changes in riming and the vertical redistribution of mixed-phase processes have also been reported in deep-convection simulations (Fan et al., 2016; Heikenfeld et al., 2019).

The second pathway involves enhanced ice nucleation and depositional growth, both of which increase in the polluted experiments throughout the cloud lifecycle (Fig. 7b,e,h,k), although the aerosol perturbations are not directly coupled to the ice nucleation particle (INP) treatment. At around 300 hPa, both cases also show elevated cloud droplet evaporation (Fig. 7f,l), co-located with enhanced deposition (Fig. 7e,k). Enhanced nucleation increases the number of ice crystals and potentially the total ice surface area available for vapour deposition (Khain et al., 2015; Fan et al., 2016), while increased evaporation provides an additional source of water vapour. Since ice nucleation in the microphysics scheme responds primarily to temperature and supersaturation with respect to ice, changes in the vapour field provide an indirect pathway through which the aerosol perturbation may influence both nucleation and depositional growth. Smaller droplets are less efficiently collected by falling ice particles, reducing riming (e.g., Wang and Ji, 2000; Pruppacher, 2010; Lamb and Verlinde, 2011), while this might be true, the model does not preferentially sort droplets by their sizes during cloud evaporation (e.g., Small et al., 2009; Karset et al., 2020). The co-location of enhanced droplet evaporation and ice deposition is consistent with a WBF-type pathway, whereby conditions subsaturated with respect to liquid water but supersaturated with respect to ice permit this phase transition (Tao et al., 2012; Storelvmo and Tan, 2015; Omanovic et al., 2024).

However, evaporation need not be the sole contributor to enhanced deposition. Changes in convective ascent and associated cooling can also modify ice supersaturation and depositional growth, while aerosol-induced changes in latent heating can alter updraft strength in deep convection (e.g., Tao et al., 2012; Storer and van den Heever, 2013). Such dynamical feedbacks are not necessarily monotonic because enhanced condensate loading can partly offset aerosol-induced latent-heating effects on buoyancy (Igel and van den Heever, 2021; Varble, 2018). Likewise, enhanced deposition does not intrinsically imply the

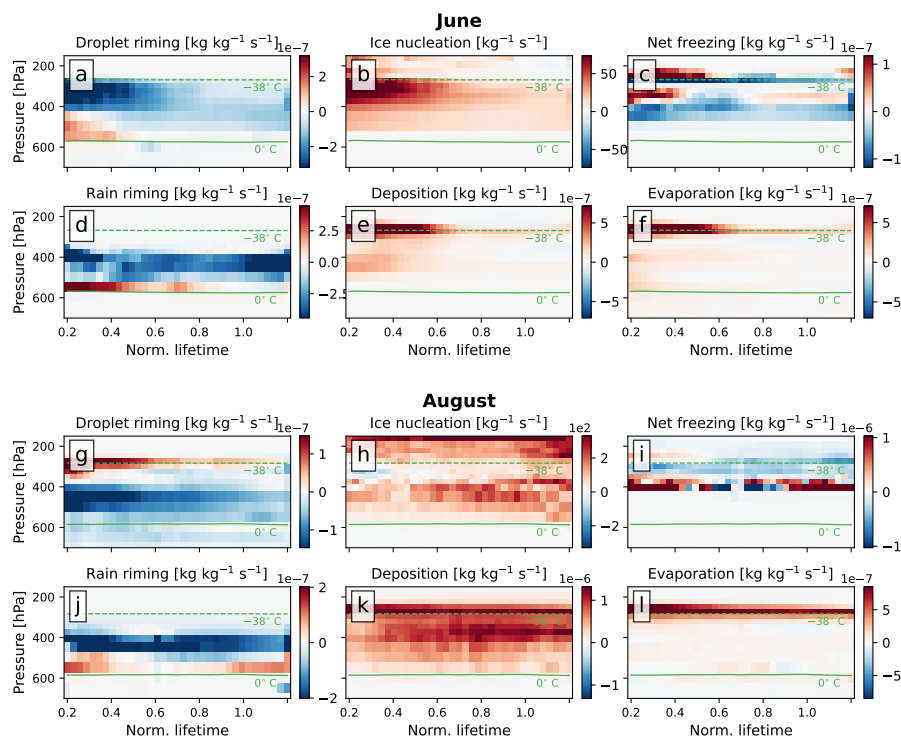


Figure 7. Polluted-minus-clean differences in selected microphysical process rates for the (a–f) June and (g–l) August cases. The x-axis represents the normalised DCC lifecycle from the onset of glaciation, with each DCC lifetime rescaled from 0 to 1. Positive values indicate larger process rates in the polluted simulation, and negative values indicate larger process rates in the clean simulation. The solid and dashed green curves are the 0°C and -38°C isotherms, respectively. Process rates are averaged over the isolated DCCs across members.

formation of smaller, slowly sedimenting ice; its effect depends on which frozen hydrometeor population receives the deposited mass and on their particle-size distribution (Heikenfeld et al., 2019). In our simulations, however, enhanced deposition occurs alongside increased ice number and mass, reduced riming, and reduced sedimentation, consistent with a redistribution toward the smaller frozen categories rather than precipitating-ice growth.

Both cases show reduced riming under polluted conditions across a substantial part of the mixed-phase region, although the altitude and timing of the response differ. The net-freezing response is less systematic. Rather than showing a uniform enhancement under polluted conditions, its sign varies with height and lifecycle stage (Fig. 7c,i). This suggests that the aerosol effect on net freezing depends on altitude and lifecycle stage rather than representing a uniform enhancement throughout the DCC. The case dependence is particularly evident in August, where the riming and freezing responses are less vertically coherent than in June, consistent with the more isolated and heterogeneous August DCC population. Similar sensitivity of aerosol responses to the convective and environmental context has been reported for isolated DCCs (Wang et al., 2025). The common reduction in riming therefore suggests a robust aerosol sensitivity within our simulations, whereas the broader mixed-



phase response remains dependent on lifecycle stage and convective environment. Such variability is consistent with previous studies showing that aerosol effects can emerge from competing and compensating microphysical pathways (Tao et al., 2012; Fan et al., 2016; Heikenfeld et al., 2019; Dagan et al., 2020).

Overall, the process-rate analysis aligns with the results suggested by the bulk hydrometeor fields. Aerosol loading suppresses warm-rain and riming pathways, maintains more condensate in smaller liquid and ice categories, and promotes vapour-mediated ice growth. This shift provides a process-level explanation for why polluted conditions in these cases produce less rain and hail but more upper-level ice and snow. Because the same pathway modifies the amount and distribution of anvil condensate, it should also affect both LW trapping and SW reflection, motivating the cloud-radiative-effect analysis in the next section.

3.4 Radiative effects

Deep-convective anvil CRE depends not only on condensate amount but also on its radiative properties, cloud-top temperature, horizontal extent, and evolution relative to the solar diurnal cycle (Koren et al., 2010; Lutsko et al., 2026). Aerosols can modify these anvil characteristics and thereby alter the resulting CRE (Fan et al., 2013; Herbert et al., 2021). The shift from riming- to deposition-dominated ice production identified in Section 3.3 has direct consequences for the anvil radiative balance. By retaining more condensate in cloud ice rather than transferring it into fast-sedimenting graupel and hail, polluted conditions modify the amount and distribution of anvil condensate and, hence, the competition between LW warming and SW cooling.

This competition is evident in the normalised lifecycle (Fig. 8) and domain-mean diurnal-cycle CRE (Fig. 10). CRE is defined here as the difference between the clear-sky and all-sky outgoing radiation at the top of the atmosphere (TOA), such that positive values indicate a warming effect and vice versa, where net CRE denotes the sum of the SW and LW CRE. In both cases, polluted simulations show enhanced LW warming and SW cooling throughout the DCC lifecycle. SW cooling dominates during the active convective period, resulting in a net cooling response, which relaxes toward warming as the storms dissipate. This behaviour is also evident in the domain-mean diurnal cycle, where the magnitude of both LW warming and SW cooling increases following convective development. The similarity in the timing of the response across aerosol experiments is consistent with the broadly similar storm evolution and lifecycle timing discussed above. However, the amplitude of the response increases with aerosol loading, indicating that the microphysical changes identified in Sections 3.1–3.3 influence the radiative effect without fundamentally changing timing and bulk storm evolution.

The enhanced LW warming is consistent with the larger amount of upper-level ice in the polluted simulations. By suppressing riming and sedimentation, the polluted cases maintain more condensate in slowly sedimenting ice categories, allowing anvil condensate to persist and spread horizontally. More slowly sedimenting, persistent small ice particles have been shown to increase cloud lifetime and coverage and associated LW warming (Fan et al., 2013). This interpretation is also consistent with previous studies linking aerosol perturbations, enhanced high-cloud cover, and positive LW cloud radiative effects in deep convective regimes (e.g., Herbert et al., 2021; Dagan and Stier, 2020). At the same time, the increased ice mass and number enhance the anvil's ability to reflect incoming solar radiation, strengthening the SW cooling. The polluted cases therefore amplify both the SW and LW anvil radiative response. Importantly, the resulting net cooling is not a universal consequence

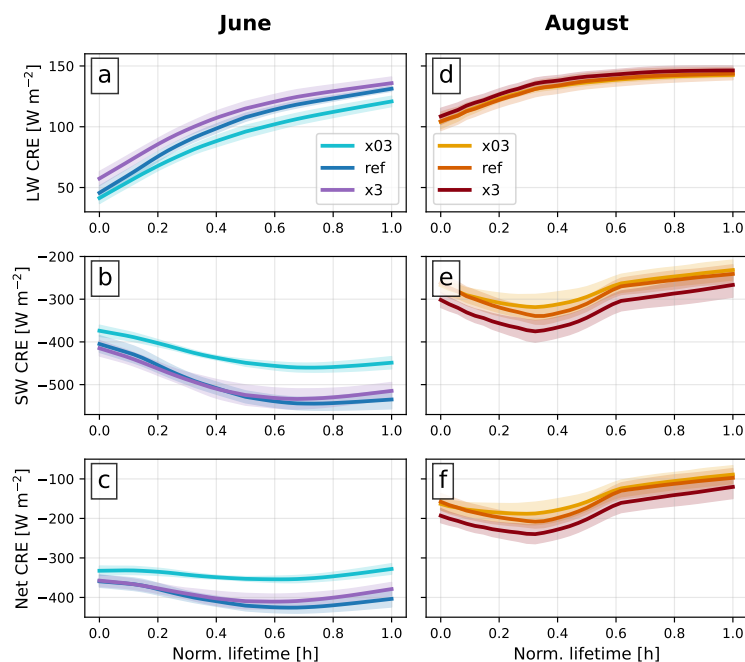


Figure 8. Normalised lifecycle of the (a and d) LW, (b and e) SW, and (c and f) net CRE for both the June (blue shades) and August (orange shades) cases. The x-axis represents the normalised DCC lifecycle from the onset of glaciation, with each DCC lifetime rescaled from 0 to 1. Lines show the ensemble mean, and shading indicates the ensemble standard deviation. In this convention, positive and negative values indicate warming and cooling effects, respectively.

of aerosol perturbations to DCCs. Koren et al. (2010), for example, found aerosol-associated anvil expansion accompanied by optical thinning, for which the reduced SW reflection contributed to a net warming response. This contrast highlights that the sign of the aerosol-induced DCC CRE depends on how microphysical changes project onto anvil optical properties, vertical structure, and lifecycle.

The magnitude and lifecycle expression of this radiative response, however, also depend on the underlying convective regime. Fig. 9 provides further context for this case dependence by showing the normalised anvil area distribution and the fraction of DCCs reaching different CTH pressure thresholds. The August case contains a greater proportion of smaller anvils, with the area distribution peaking at smaller sizes than in June, whereas the June case is characterised by larger systems. Such differences are relevant to anvil evolution because convective intensity, core size, and organisation have been shown to influence anvil extent, structure, and lifecycle (e.g., Jones and Stier, 2025; Ritman et al., 2026).

The CTH statistics (Fig. 9c,d) further show that the two cases differ in vertical structure. Differences in CTH are radiatively important because cloud-top temperature strongly influences the LW component of anvil CRE (Koren et al., 2010). In August, a smaller fraction of DCCs reach the 250 and 200 hPa thresholds, consistent with a larger contribution from congestus-like or less vertically organised systems. At the same time, a larger fraction reaches the highest cloud-top threshold (150hPa). Hence,

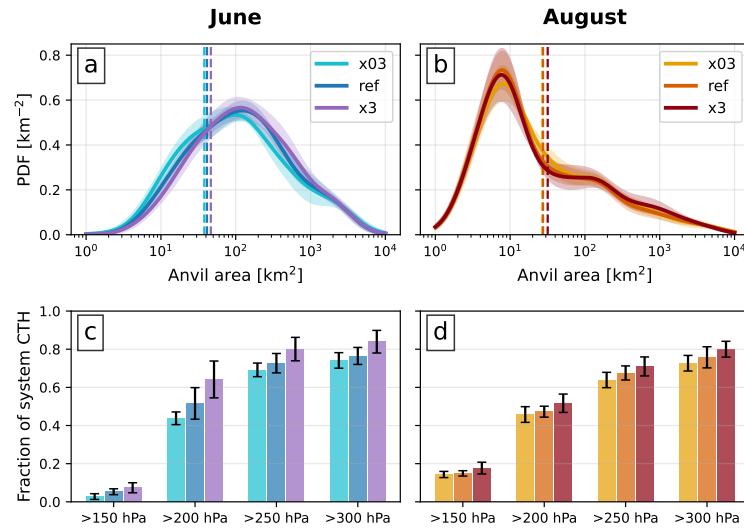


Figure 9. Statistics of (a and b) anvil size Probability density functions (PDF) and (c and d) system cloud-top-height fraction for the June (blue shades) and August (orange shades) cases. The PDF is derived from the two-dimensional projection of the isolated DCC mask after the onset of freezing, with lines and shading indicating the ensemble mean and standard deviation, respectively. Dashed vertical lines indicate the mean anvil area. The fraction of systems whose cloud-top height reaches pressures below 150, 200, 250, and 300 hPa are presented, with bars and error bars indicating ensemble mean and standard deviation, respectively.

August contains a more heterogeneous DCC population, with greater representation of relatively shallow systems while still supporting occasional very deep convection. This broader cloud-depth distribution is consistent with the contrasting convective environments described in Section 2.2. Importantly, the aerosol response in CTH is monotonic in both cases, suggesting that although the overall cloud-depth distribution is dynamically controlled, the aerosol perturbation produces a systematic response within each case.

These differences are also reflected in the DCC CRE lifecycle (Fig. 8). Across all aerosol experiments, the June DCCs maintain stronger SW cooling more consistently through the lifecycle, whereas the SW cooling weakens during the later stages of the August DCCs. This difference is consistent with the contrasting DCC populations shown in Fig. 9, with August characterised by a broader distribution of cloud depths and a greater proportion of smaller anvils. In both cases, stronger SW cooling outweighs the enhanced LW warming, producing a more negative net CRE in the polluted simulations during the storm duration. To quantify the impact of aerosol perturbations on the entire domain, we also calculate the averaged CRE over the 24 h analysis period and the entire domain (see Fig. 10). Quantitatively, the difference in net CRE between the polluted and clean experiments for the entire span of the simulation is $\Delta CRE_{3-0.3}(June) = -0.22 \pm 2.08 \text{ W m}^{-2}$ and $\Delta CRE_{3-0.3}(August) = -3.59 \pm 0.33 \text{ W m}^{-2}$, where the uncertainty denotes the standard error of the mean. The August case shows a more robust cooling response, whereas the June case is less certain. The high uncertainty in June is associated with the late-stage dissipation period, as suggested by Fig. 10a,g. This case dependence reflects the sensitivity of net anvil

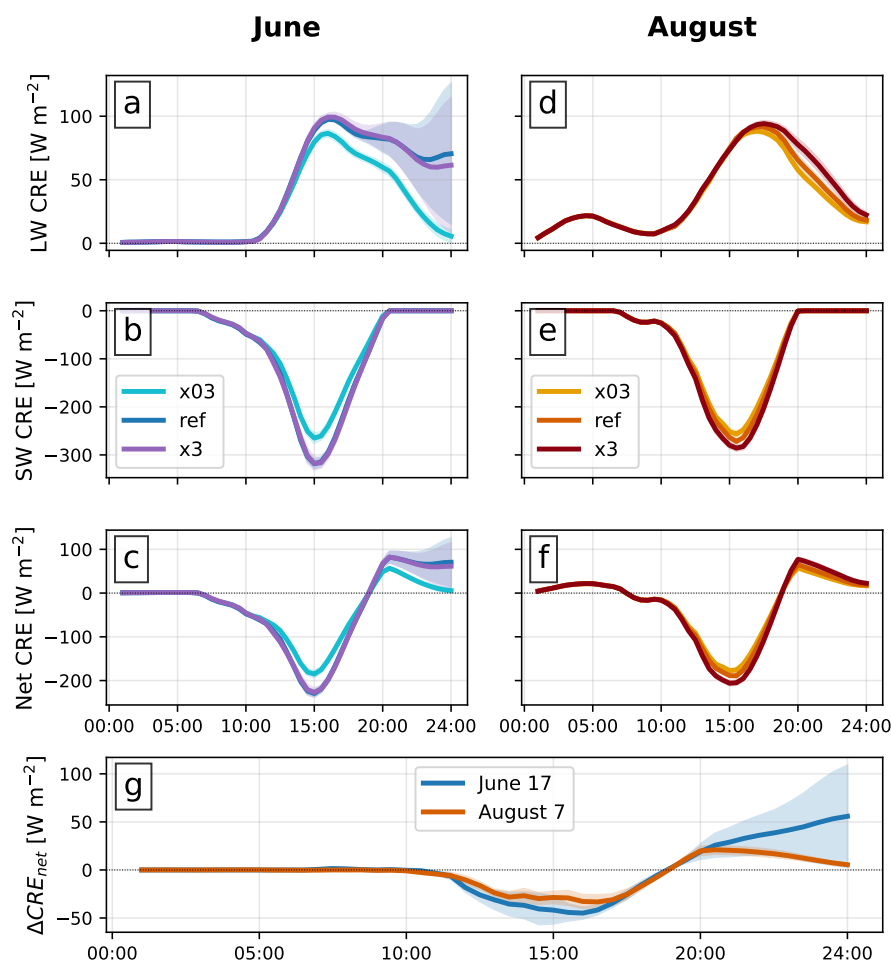


Figure 10. Domain mean diurnal cycle of the (a and d) LW, (b and e) SW, and (c and f) total CRE for both the June (blue shades) and August (orange shades) cases, along (g) the polluted-minus-clean LW CRE for the June (blue) and August (orange) cases. Lines show the ensemble mean, and shading indicates the ensemble standard deviation. In this convention, positive and negative values indicate warming and cooling effects, respectively.

340 CRE to several factors, including optical depth and cloud-top height (Koren et al., 2010), lifecycle and timing relative to the solar cycle (Lutsko et al., 2026), and differences in anvil structure associated with the convective organisation (Jones and Stier, 2025). For example, Koren et al. (2010) reported a net warming effect from aerosol–DCC interactions in the tropics associated with changes in the wind profile, whereas our simulations produce a more negative net CRE without a significant change in the vertical distribution of the wind fields.

345 The difference in net CRE between the polluted and clean experiments shown in Fig. 10g diverges strongly in the final hours of the June case, after the main convective systems have dissipated. Both the mean and error of the June net



CRE are highly sensitive to the post-storm averaging window. For example, a 24 h window starting at 04:00 instead of 06:00 reduces the contribution of this late-stage divergence and yields $\Delta CRE_{3-0.3}(\text{June}) = -4.37 \pm 0.99 \text{ W m}^{-2}$ and $\Delta CRE_{3-0.3}(\text{August}) = -4.30 \pm 0.31 \text{ W m}^{-2}$. The substantially larger change in June than in August indicates that the uncertainty in the June full-period estimate is primarily associated with the evolution of residual upper-level condensate after the main convective systems decay. During this dissipation stage, anvil persistence depends on the competing effects of ice growth, sedimentation, sublimation, and environmental conditions (e.g., Jones and Stier, 2025; Gasparini et al., 2022), while the representation of these processes remains sensitive to assumptions within the cloud-microphysics parameterisation (e.g., Khain et al., 2015; Sela et al., 2026). Importantly, Fig. 6 show that the two most aerosol-rich June experiments (i.e., ref and x3) also produce the highest ice mass and number concentrations. Thus, once active convection ceases, these experiments leave a more numerous population of relatively low-mass ice particles whose persistence becomes increasingly controlled by sedimentation, sublimation and deposition, and the surrounding environment. Because the two-moment scheme derives sedimentation from the predicted mass and number moments through an assumed particle-size distribution, changes in q_i and N_i propagate into the parameterised sedimentation response (Bolt and Omanovic, 2026). Residual anvil dissipation is also influenced by environmental humidity, mixing, and radiatively driven circulations (Gasparini et al., 2022; Jones and Stier, 2025), which may contribute to the member-to-member spread during this late stage. Despite this late-stage sensitivity, both cases show that aerosol-induced changes in mixed-phase condensate partitioning propagate into stronger, competing LW and SW anvil radiative effects.

4 Summary and conclusions

In this study, we investigate how anthropogenic aerosol perturbations modify deep convection in two distinct sea-breeze cases over Houston, and how those changes propagate into the cloud radiative effect, using a high-resolution CRM coupled to an interactive aerosol module. Across both cases, increasing aerosol loading consistently increases CDNC and suppresses warm-rain formation, consistent with established aerosol–cloud responses (Fan et al., 2016; Heikenfeld et al., 2019). The process-rate analysis further reveals a shift in the balance between competing mixed-phase pathways, with reduced riming and enhanced droplet evaporation, ice nucleation, and vapour deposition. This reorganisation of the hydrometeor budget maintains more condensate in cloud water, ice, and snow, while reducing the transfer into fast-sedimenting graupel and hail. As a result, polluted conditions enhance both LW warming and, predominantly, SW cooling during the active convective period, producing stronger net cooling in both cases. However, the full-period, domain-mean net CRE differences between the polluted and clean experiments are $-0.22 \pm 2.08 \text{ W m}^{-2}$ and $-3.59 \pm 0.33 \text{ W m}^{-2}$ for June and August, respectively. The June estimate is particularly sensitive to the evolution of residual upper-level cloud during the late-stage dissipation period.

At the same time, the two case studies show that the aerosol response is modulated by the background convective environment (Khain et al., 2005). The more organised June systems exhibit a clearer aerosol response in the anvil and sustained SW cooling, whereas the smaller and more isolated August systems show greater vertical variability in hydrometeor structure, a riming response that differs in altitude and lifecycle timing, and weaker late-stage SW cooling. The June case addition-



ally develops a pronounced late-stage LW spread in the domain-mean CRE during storm dissipation, a behaviour absent in August. Because the DCC CRE spread in Fig. 8 remains modest, this variability may reflect differences in the evolution of residual upper-level cloud after the main convective systems decay, highlighting the sensitivity of the late-stage radiative response to the persistence of ice-phase condensate. This dissipation tail introduces substantial uncertainty into June's full-period CRE estimate, particularly as the most aerosol-rich simulations retain greater upper-level ice mass and number populations. Thus, although the principal aerosol-induced microphysical pathways and the enhanced LW warming and SW cooling during active convection are consistent across the two cases, the full-period net radiative response is additionally influenced by case-dependent cloud evolution, particularly during dissipation.

Several limitations remain. The two case studies sample only a narrow part of the full range of possible DCC environments, so additional simulations would be needed to test whether the same response holds under a broader range of thermodynamic conditions, convective organisation, and diurnal timing. The environment around Houston is also already aerosol-rich, so simulations spanning a broader range of aerosol conditions, including more pristine environments, would be needed to assess the relevance of these mechanisms to pre-industrial aerosol conditions and their potential contribution to aerosol effective radiative forcing. Additionally, the sensitivity of the late-stage LW response to residual upper-level cloud evolution motivates cloud-microphysical sensitivity studies, e.g. in the form of a perturbed parameter ensemble (PPE), to quantify parametric sensitivity within the microphysics scheme. Structural uncertainty, potentially associated with modelling assumptions such as saturation adjustment, requires complementary evaluation across different model formulations, for example through MIPs such as TRACER-MIP (Saleeby and Fan, 2025). Observational constraints remain essential to evaluate simulated hydrometeor partitioning and anvil evolution.

These findings have implications for understanding the influence of anthropogenic aerosols on atmospheric hydrological and radiative budgets. In particular, they demonstrate that aerosol-induced changes in mixed-phase hydrometeor partitioning can affect both precipitation formation and, through changes in anvil condensate, the radiative effects of deep convection, even when the overall storm evolution remains broadly similar. Importantly, the radiative consequences of aerosol perturbations are not limited to active convection, as the evolution and dissipation of residual upper-level cloud can substantially influence the time-integrated radiative response. This highlights the importance of representing both the competing mixed-phase microphysical processes and cloud evolution throughout the convective lifecycle when assessing aerosol–deep-convection interactions. Improving these representations and constraining them with observations is therefore important for reducing uncertainty in the radiative and hydrological responses to changing aerosol conditions.



Author contributions. MS performed the simulations and analyses, and prepared the manuscript, with contributions and comments from all
410 co-authors. MS, SD, and PS carried out the study design and interpretation. MR developed and ran the *tobac* tracking algorithm. SD assisted
with the model setup, branch maintenance, and testing.

Competing interests. The authors declare that they have no competing interests.

Acknowledgements. The simulations were performed and analysed on the Levante cluster of the DKRZ with resources granted under project
1368 (<https://www.dkrz.de/en/systems/hpc/hhre-4-levante>, last access: 20 August 2026). MS acknowledges funding from the NERC Doctoral
415 Training Partnership in Environmental Research Grant NE/S007474/1. SD, MR and PS acknowledge funding from the European Union's
Horizon Europe project CleanCloud with grant agreement 101137639 and its UKRI underwrite. MR also acknowledges the General Sir John
Monash Foundation.



References

- Abramian, S., Muller, C., Risi, C., Fiolleau, T., and Roca, R.: How key features of early development shape deep convective systems, *npj Climate and Atmospheric Science*, 8, 258, <https://doi.org/10.1038/s41612-025-01154-1>, 2025.
- Al-Jabri, A., Lebo, Z. J., and Wang, D.: Investigating Aerosol and Meteorological Influences on Convective Clouds in Houston, Texas, during the TRACER/ESCAPE Field Campaigns, *Journal of the Atmospheric Sciences*, 82, 2469 – 2489, <https://doi.org/10.1175/JAS-D-25-0004.1>, 2025.
- Andreae, M. O., Rosenfeld, D., Artaxo, P., Costa, A. A., Frank, G. P., Longo, K. M., and Silva-Dias, M. A. F.: Smoking rain clouds over the Amazon, *Science*, 303, 1337–1342, <https://doi.org/10.1126/science.1092779>, 2004.
- Barthlott, C., Zarbo, A., Matsunobu, T., and Keil, C.: Importance of aerosols and shape of the cloud droplet size distribution for convective clouds and precipitation, *Atmospheric Chemistry and Physics*, 22, 2153–2172, <https://doi.org/10.5194/acp-22-2153-2022>, 2022.
- Bolt, S. and Omanovic, N.: Evaluation of semi-implicit and explicit sedimentation approaches in the two-moment cloud microphysics scheme of ICON, *Geoscientific Model Development*, 19, 595–619, <https://doi.org/10.5194/gmd-19-595-2026>, 2026.
- Bony, S., Stevens, B., Frierson, D. M. W., Bentsen, M., Boucher, O., Dufresne, J.-L., McInerney, D., Webb, M. J., Sherwood, S. C., and Webb, M. J.: Clouds, circulation and climate sensitivity, *Nature Geoscience*, 8, 261–268, <https://doi.org/10.1038/ngeo2398>, 2015.
- Chen, B., Thompson, S. A., Matthews, B. H., Sharma, M., Li, R., Nowotarski, C. J., Rapp, A. D., and Brooks, S. D.: A new technique to retrieve aerosol vertical profiles using micropulse lidar and ground-based aerosol measurements, *Atmospheric Measurement Techniques*, 18, 5841–5859, <https://doi.org/10.5194/amt-18-5841-2025>, 2025.
- Dagan, G. and Stier, P.: Ensemble daily simulations for elucidating cloud–aerosol interactions under a large spread of realistic environmental conditions, *Atmospheric Chemistry and Physics*, 20, 6291–6303, <https://doi.org/10.5194/acp-20-6291-2020>, 2020.
- Dagan, G., Stier, P., Christensen, M., Cioni, G., Klocke, D., and Seifert, A.: Atmospheric energy budget response to idealized aerosol perturbation in tropical cloud systems, *Atmospheric Chemistry and Physics*, 20, 4523–4544, <https://doi.org/10.5194/acp-20-4523-2020>, 2020.
- De, S. and Stier, P.: ..., manuscript in preparation, 2026.
- Dipankar, A., Stevens, B., Heinze, R., Moseley, C., Zängl, G., Giorgetta, M., and Brdar, S.: Large eddy simulation using the general circulation model ICON, *Journal of Advances in Modeling Earth Systems*, 7, 963–986, <https://doi.org/10.1002/2015MS000431>, 2015.
- Emanuel, K. A.: Overview of Atmospheric Convection, pp. 1–28, Springer Netherlands, Dordrecht, ISBN 978-94-015-8828-7, https://doi.org/10.1007/978-94-015-8828-7_1, 1997.
- Fan, J., Leung, L., Rosenfeld, D., Chen, Q., Li, Z., Zhang, J., and Yan, H.: Microphysical effects determine macrophysical response for aerosol impacts on deep convective clouds, *Proceedings of the National Academy of Sciences*, <https://doi.org/10.1073/pnas.1316830110>, 2013.
- Fan, J., Wang, Y., Rosenfeld, D., and Liu, X.: Review of Aerosol–Cloud Interactions: Mechanisms, Significance, and Challenges, *Journal of the Atmospheric Sciences*, <https://doi.org/10.1175/jas-d-16-0037.1>, 2016.
- Gasparini, B., Sokol, A. B., Wall, C. J., Hartmann, D. L., and Blossey, P. N.: Diurnal Differences in Tropical Maritime Anvil Cloud Evolution, *Journal of Climate*, 35, 1655 – 1677, <https://doi.org/10.1175/JCLI-D-21-0211.1>, 2022.
- Glassmeier, F. and Lohmann, U.: Constraining Precipitation Susceptibility of Warm-, Ice-, and Mixed-Phase Clouds with Microphysical Equations, *Journal of the Atmospheric Sciences*, 73, 5003 – 5023, <https://doi.org/10.1175/JAS-D-16-0008.1>, 2016.



- Hashino, T., Chiruta, M., Polzin, D., Kubicek, A., and Wang, P. K.: Numerical simulation of the flow fields around falling ice crystals with
455 inclined orientation and the hydrodynamic torque, *Atmospheric Research*, 150, 79–96, <https://doi.org/10.1016/j.atmosres.2014.07.003>, 2014.
- Heikenfeld, M., White, B., Labbouz, L., and Stier, P.: Aerosol effects on deep convection: the propagation of aerosol perturbations through
convective cloud microphysics, *Atmospheric Chemistry and Physics*, <https://doi.org/10.5194/acp-19-2601-2019>, 2019.
- Heinold, B., Weiss, P., De, S., Kubin, A., Müller, J., Senf, F., Stier, P., and Tegen, I.: ICON coupled to HAM-lite 1.0 in limited-area mode: an
460 efficient framework for targeted kilometer-scale simulations with interactive aerosols, *Geoscientific Model Development*, 19, 6167–6187,
<https://doi.org/10.5194/gmd-19-6167-2026>, 2026.
- Herbert, R., Stier, P., and Dagan, G.: Isolating Large-Scale Smoke Impacts on Cloud and Precipitation Processes Over
the Amazon With Convection Permitting Resolution, *Journal of Geophysical Research: Atmospheres*, 126, e2021JD034615,
<https://doi.org/https://doi.org/10.1029/2021JD034615>, e2021JD034615 2021JD034615, 2021.
- 465 Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horányi, A., Muñoz-Sabater, J., Nicolas, J., Peubey, C., Radu, R., Schepers, D., Simmons,
A., Soci, C., Abdalla, S., Abellan, X., Balsamo, G., Bechtold, P., Biavati, G., Bidlot, J., Bonavita, M., De Chiara, G., Dahlgren, P., Dee, D.,
Diamantakis, M., Dragani, R., Flemming, J., Forbes, R., Fuentes, M., Geer, A., Haimberger, L., Healy, S., Hogan, R., Hólm, E., Janisková,
M., Keeley, S., Laloyaux, P., Lopez, P., Lupu, C., Radnoti, G., de Rosnay, P., Rozum, I., Vamborg, F., Villaume, S., and Thépaut, J.-
N.: Complete ERA5 from 1940: Fifth generation of ECMWF atmospheric reanalyses of the global climate, Copernicus Climate Change
470 Service (C3S) Data Store (CDS), <https://doi.org/10.24381/cds.143582cf>, accessed on 01-DEC-2024, 2017.
- Hohenegger, C., Korn, P., Linardakis, L., Redler, R., Schnur, R., Adamidis, P., Bao, J., Bastin, S., Behraves, M., Bergemann, M., Biercamp,
J., Bockelmann, H., Brokopf, R., Brüggemann, N., Casaroli, L., Chegini, F., Datseris, G., Esch, M., George, G., Giorgetta, M., Gutjahr,
O., Haak, H., Hanke, M., Ilyina, T., Jahns, T., Jungclaus, J., Kern, M., Klocke, D., Kluft, L., Kölling, T., Kornblueh, L., Kosukhin, S.,
Kroll, C., Lee, J., Mauritsen, T., Mehlmann, C., Mieslinger, T., Naumann, A. K., Paccini, L., Peinado, A., Praturi, D. S., Putrasahan, D.,
475 Rast, S., Riddick, T., Roeber, N., Schmidt, H., Schulzweida, U., Schütte, F., Segura, H., Shevchenko, R., Singh, V., Specht, M., Stephan,
C. C., von Storch, J.-S., Vogel, R., Wengel, C., Winkler, M., Ziemann, F., Marotzke, J., and Stevens, B.: ICON-Sapphire: simulating
the components of the Earth system and their interactions at kilometer and subkilometer scales, *Geoscientific Model Development*, 16,
779–811, <https://doi.org/10.5194/gmd-16-779-2023>, 2023.
- Igel, A. L. and van den Heever, S. C.: Invigoration or Enervation of Convective Clouds by Aerosols?, *Geophysical Research Letters*, 48,
480 e2021GL093804, <https://doi.org/https://doi.org/10.1029/2021GL093804>, e2021GL093804 2021GL093804, 2021.
- Inness, A., Ades, M., Agustí-Panareda, A., Barré, J., Benedictow, A., Blechschmidt, A.-M., Dominguez, J. J., Engelen, R., Eskes, H., Flem-
ming, J., Huijnen, V., Jones, L., Kipling, Z., Massart, S., Parrington, M., Peuch, V.-H., Razinger, M., Remy, S., Schulz, M., and Suttie, M.:
The CAMS reanalysis of atmospheric composition, *Atmospheric Chemistry and Physics*, 19, 3515–3556, <https://doi.org/10.5194/acp-19-3515-2019>, 2019.
- 485 Jensen, M. P., Flynn, J. H., Kollias, P., Kuang, C., McFarquhar, G., Powers, H., Brooks, S., Bruning, E., Collins, D., Collis, S. M., Fan, J.,
Fridlind, A., Giangrande, S. E., Griffin, R., Hu, J., Jackson, R. C., Kumjian, M., Logan, T., Matsui, T., Nowotarski, C. J., Oue, M., Rapp,
A. D., Rosenfeld, D., Ryzhkov, A., Sheesley, R., Snyder, J., Stier, P., Usenko, S., van den Heever, S. C., van Lier-Walqui, M., Varble, A.,
and Wang, Y.: Tracking Aerosol Convection Interactions Experiment (TRACER) Field Campaign Report, Tech. Rep. DOE/SC-ARM-23-
038, U.S. Department of Energy, Atmospheric Radiation Measurement (ARM) user facility, <https://doi.org/10.2172/2202672>, 2023.
- 490 Jones, W. K. and Stier, P.: Effects of convective intensity and organisation on the structure and lifecycle of deep convective clouds, *EGU-*
sphere, 2025, 1–26, <https://doi.org/10.5194/egusphere-2025-6391>, 2025.



- Karset, I. H. H., Gettelman, A., Storelvmo, T., Alterskjær, K., and Berntsen, T. K.: Exploring Impacts of Size-Dependent Evaporation and Entrainment in a Global Model, *Journal of Geophysical Research: Atmospheres*, 125, e2019JD031817, <https://doi.org/https://doi.org/10.1029/2019JD031817>, e2019JD031817 10.1029/2019JD031817, 2020.
- 495 Khain, A., Rosenfeld, D., and Pokrovsky, A.: Aerosol impact on the dynamics and microphysics of deep convective clouds, *Quarterly Journal of the Royal Meteorological Society*, 131, 2639–2663, <https://doi.org/https://doi.org/10.1256/qj.04.62>, 2005.
- Khain, A. P., Beheng, K. D., Heymsfield, A., Korolev, A., Krichak, S. O., Levin, Z., Pinsky, M., Phillips, V., Prabhakaran, T., Teller, A., van den Heever, S. C., and Yano, J.-I.: Representation of microphysical processes in cloud-resolving models: Spectral (bin) microphysics versus bulk parameterization, *Reviews of Geophysics*, 53, 247–322, <https://doi.org/https://doi.org/10.1002/2014RG000468>, 2015.
- 500 Koren, I., Remer, L. A., Altaratz, O., Martins, J. V., and Davidi, A.: Aerosol-induced changes of convective cloud anvils produce strong climate warming, *Atmospheric Chemistry and Physics*, 10, 5001–5010, <https://doi.org/10.5194/acp-10-5001-2010>, 2010.
- Lamb, D. and Verlinde, J.: *Physics and chemistry of clouds*, Cambridge University Press, United Kingdom, ISBN 9780521899109, <https://doi.org/10.1017/CBO9780511976377>, 2011.
- Leuenberger, D., Koller, M., Fuhrer, O., and Schär, C.: A Generalization of the SLEVE Vertical Coordinate, *Monthly Weather Review*, 138, 3683 – 3689, <https://doi.org/10.1175/2010MWR3307.1>, 2010.
- 505 Lutsko, N. J., Wall, C. J., and Gasparini, B.: An Analytical Model of the Lifecycle of Tropical Anvil Cloud Radiative Effects, *Journal of Geophysical Research: Atmospheres*, 131, e2025JD045070, <https://doi.org/10.1029/2025JD045070>, e2025JD045070 2025JD045070, 2026.
- Marinescu, P. J., van den Heever, S. C., Heikenfeld, M., Barrett, A. I., Barthlott, C., Hoose, C., Fan, J., Fridlind, A. M., Matsui, T., Miltenberger, A. K., Stier, P., Vie, B., White, B. A., and Zhang, Y.: Impacts of Varying Concentrations of Cloud Condensation Nuclei on Deep Convective Cloud Updrafts—A Multimodel Assessment, *Journal of the Atmospheric Sciences*, 78, 1147 – 1172, <https://doi.org/10.1175/JAS-D-20-0200.1>, 2021.
- Morrison, H., van Lier-Walqui, M., Fridlind, A. M., Grabowski, W. W., Harrington, J. Y., Hoose, C., Korolev, A., Kumjian, M. R., Milbrandt, J. A., Pawlowska, H., Posselt, D. J., Prat, O. P., Reimel, K. J., Shima, S.-I., van Diedenhoven, B., and Xue, L.: Confronting the Challenge of Modeling Cloud and Precipitation Microphysics, *Journal of Advances in Modeling Earth Systems*, 12, e2019MS001689, <https://doi.org/10.1029/2019MS001689>, e2019MS001689 2019MS001689, 2020.
- 515 Omanovic, N., Ferrachat, S., Fuchs, C., Henneberger, J., Miller, A. J., Ohneiser, K., Ramelli, F., Seifert, P., Spirig, R., Zhang, H., and Lohmann, U.: Evaluating the Wegener–Bergeron–Findeisen process in ICON in large-eddy mode with in situ observations from the CLOUDLAB project, *Atmospheric Chemistry and Physics*, 24, 6825–6844, <https://doi.org/10.5194/acp-24-6825-2024>, 2024.
- 520 Pincus, R., Mlawer, E. J., and Delamere, J. S.: Balancing Accuracy, Efficiency, and Flexibility in Radiation Calculations for Dynamical Models, *Journal of Advances in Modeling Earth Systems*, 11, 3074–3089, <https://doi.org/10.1029/2019MS001621>, 2019.
- Pinsky, M. B., Khain, A. P., and Shapiro, M.: Collisions of Cloud Droplets in a Turbulent Flow. Part IV: Droplet Hydrodynamic Interaction, *Journal of the Atmospheric Sciences*, 64, 2462 – 2482, <https://doi.org/10.1175/JAS3952.1>, 2007.
- Pruppacher, H.R. and Klett, J.: *Cloud Particle Interactions*, pp. 568–616, Springer Netherlands, Dordrecht, ISBN 978-0-306-48100-0, https://doi.org/10.1007/978-0-306-48100-0_14, 2010.
- 525 Reick, C. H., Gayler, V., Goll, D., Hagemann, S., Heidkamp, M., and Nabel, J. E. M. S. a.: JSBACH 3 - The land component of the MPI Earth System Model: documentation of version 3.2, MPI für Meteorologie, Hamburg, <https://doi.org/10.17617/2.3279802>, 2021.



- Reutter, P., Su, H., Trentmann, J., Simmel, M., Rose, D., Gunthe, S. S., Wernli, H., Andreae, M. O., and Pöschl, U.: Aerosol- and updraft-limited regimes of cloud droplet formation: influence of particle number, size and hygroscopicity on the activation of cloud condensation nuclei (CCN), *Atmospheric Chemistry and Physics*, 9, 7067–7080, <https://doi.org/10.5194/acp-9-7067-2009>, 2009.
- 530 Ritman, M., Jones, W., and Stier, P.: Convective controls on anvil cloud evolution in the ICON km-scale global climate model, *Atmospheric Chemistry and Physics*, 26, 7105–7126, <https://doi.org/10.5194/acp-26-7105-2026>, 2026.
- Rosenfeld, D., Lohmann, U., Raga, G. B., O'Dowd, C. D., Kulmala, M., Fuzzi, S., Reissell, A., and Andreae, M. O.: Flood or Drought: How Do Aerosols Affect Precipitation?, *Science*, 321, 1309–1313, <https://doi.org/10.1126/science.1160606>, 2008.
- 535 Rosenfeld, D., Andreae, M., Asmi, A., Chin, M., Leeuw, G., Donovan, D., Kahn, R., Kinne, S., Kivekäs, N., Kulmala, M., Lau, W., Schmidt, K. S., Suni, T., Wagner, T., Wild, M., and Quaas, J.: Global observations of aerosol-cloud-precipitation-climate interactions, *Reviews of Geophysics*, <https://doi.org/10.1002/2013rg000441>, 2014.
- Saleeby, S. and Fan, J.: Tracking Aerosol Convection Interaction Experiment (TRACER) Model Intercomparison Project (MIP) roadmap, <https://arm-synergy.github.io/tracer-mip/README.html>, tech. Rep., 2025.
- 540 Saleeby, S. M., van den Heever, S. C., Marinescu, P. J., Oue, M., Barrett, A. I., Barthlott, C., Cherian, R., Fan, J., Fridlind, A. M., Heikenfeld, M., Hoose, C., Matsui, T., Miltenberger, A. K., Quaas, J., Shpund, J., Stier, P., Vie, B., White, B. A., and Zhang, Y.: Model Intercomparison of the Impacts of Varying Cloud Droplet–Nucleating Aerosols on the Life Cycle and Microphysics of Isolated Deep Convection, *Journal of the Atmospheric Sciences*, 82, 2197 – 2217, <https://doi.org/10.1175/JAS-D-24-0181.1>, 2025.
- Segura, H., Pedruzo-Bagazgoitia, X., Weiss, P., Müller, S. K., Rackow, T., Lee, J., Dolores-Tesillos, E., Becker, T., Behrens, J., Biercamp, J., de Bodiere, P., de la Ossa, J., Doblas-Reyes, F. J., Esch, M., Ganz, C., George, G., Giorgetta, M., Hanke, M., Ilyina, T., Jahns, T., Korn, P., Kornbluh, L., Lahn, R., Linardakis, L., Löb, J., Löschke, S., Manzini, E., Mehlmann, C., Mieslinger, T., Quaas, J., Redler, R., Schmidt, H., Schnur, R., Schulzweida, U., Seneviratne, S. I., Senf, F., Sidorenko, D., Specht, M., Stier, P., von Storch, J.-S., Voss, C., Wengel, C., Wieners, K.-H., Ziemann, F., Marotzke, J., and Stevens, B.: nextGEMS: entering the era of kilometer-scale Earth system modeling, *Geoscientific Model Development*, 18, 7735–7761, <https://doi.org/10.5194/gmd-18-7735-2025>, 2025.
- 545 Seifert, A. and Beheng, K. D.: A two-moment cloud microphysics parameterization for mixed-phase clouds. Part 1: Model description, *Meteorology and Atmospheric Physics*, 92, 45–66, <https://doi.org/10.1007/s00703-005-0112-4>, 2006.
- Sela, M., Weiss, P., and Stier, P.: Sensitivity of cloud structure and precipitation to cloud microphysics schemes in ICON and implications for global km-scale simulations, *Geoscientific Model Development*, 19, 8191–8212, <https://doi.org/10.5194/gmd-19-8191-2026>, 2026.
- Simpkins, G. R.: Aerosol–cloud interactions, *Nature Climate Change*, <https://doi.org/10.1038/s41558-018-0195-9>, 2018.
- 555 Small, J. D., Chuang, P. Y., Feingold, G., and Jiang, H.: Can aerosol decrease cloud lifetime?, *Geophysical Research Letters*, 36, <https://doi.org/https://doi.org/10.1029/2009GL038888>, 2009.
- Storelvmo, T. and Tan, I.: The Wegener-Bergeron-Findeisen process? Its discovery and vital importance for weather and climate, *Meteorologische Zeitschrift*, 24, 455–461, <https://doi.org/10.1127/metz/2015/0626>, 2015.
- Storer, R. and van den Heever, S. C.: Microphysical Processes Evident in Aerosol Forcing of Tropical Deep Convective Clouds, *Journal of the Atmospheric Sciences*, <https://doi.org/10.1175/jas-d-12-076.1>, 2013.
- 560 Tao, W.-K., Chen, J.-P., Li, Z., Wang, C., and Zhang, C.: Impact of aerosols on convective clouds and precipitation, *Reviews of Geophysics*, 50, <https://doi.org/10.1029/2011RG000369>, 2012.
- Trenberth, K. E., Fasullo, J. T., and Kiehl, J.: Earth's Global Energy Budget, *Bulletin of the American Meteorological Society*, 90, 311 – 324, <https://doi.org/10.1175/2008BAMS2634.1>, 2009.



- 565 Tsai, W.-M. and Wu, C.-M.: The environment of aggregated deep convection, *Journal of Advances in Modeling Earth Systems*, 9, 2061–2078, <https://doi.org/10.1002/2017MS000967>, 2017.
- Varble, A.: Erroneous Attribution of Deep Convective Invigoration to Aerosol Concentration, *Journal of the Atmospheric Sciences*, 75, 1351 – 1368, <https://doi.org/10.1175/JAS-D-17-0217.1>, 2018.
- Wang, D., Kobrosly, R., Zhang, T., Subba, T., van den Heever, S., Gupta, S., and Jensen, M.: Aerosol impacts on isolated deep convection: findings from TRACER, *Atmospheric Chemistry and Physics*, 25, 9295–9314, <https://doi.org/10.5194/acp-25-9295-2025>, 2025.
- 570 Wang, P. K. and Ji, W.: Collision Efficiencies of Ice Crystals at Low–Intermediate Reynolds Numbers Colliding with Supercooled Cloud Droplets: A Numerical Study, *Journal of the Atmospheric Sciences*, 57, 1001 – 1009, [https://doi.org/10.1175/1520-0469\(2000\)057<1001:CEOICA>2.0.CO;2](https://doi.org/10.1175/1520-0469(2000)057<1001:CEOICA>2.0.CO;2), 2000.
- Weiss, P., Herbert, R., and Stier, P.: ICON-HAM-lite 1.0: simulating the Earth system with interactive aerosols at kilometer scales, *Geoscientific Model Development*, 18, 3877–3894, <https://doi.org/10.5194/gmd-18-3877-2025>, 2025.
- 575 White, B., Gryspeerdt, E., Stier, P., Morrison, H., Thompson, G., and Kipling, Z.: Uncertainty from the choice of microphysics scheme in convection-permitting models significantly exceeds aerosol effects, *Atmospheric Chemistry and Physics*, 17, 12 145–12 175, <https://doi.org/10.5194/acp-17-12145-2017>, 2017.
- Wu, Y. and Peng, Y.: Investigation of aerosol effects on a deep convection case in Southeast China with budget analysis of hydrometeor mass and latent heat, *Atmospheric Research*, 277, 106 275, <https://doi.org/https://doi.org/10.1016/j.atmosres.2022.106275>, 2022.
- 580 Zängl, G., Reinert, D., Rípodas, P., and Baldauf, M.: The ICON (ICOsahedral Non-hydrostatic) modelling framework of DWD and MPI-M: Description of the non-hydrostatic dynamical core, *Quarterly journal of the Royal Meteorological Society*, 141, 563–579, <https://doi.org/https://doi.org/10.1002/qj.2378>, 2015.