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What Drives Negative K_{DP} Above the Melting Layer? The Hidden Role of Secondary Ice Production

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Key Points:

- Negative K_{DP} regions above the melting layer in deep convection are frequently detected by both C- and S-band radars
- Negative K_{DP} is linked to lightning and likely reflects electrically aligned ice particles in deep convection
- Simulated negative K_{DP} occurs primarily when secondary ice production, especially fragmentation of freezing drops, is included

Supporting Information:

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

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Abstract

Secondary ice production (SIP) remains a major uncertainty in understanding deep convection because its rates and controlling conditions are poorly constrained. Using the Tracking Aerosol Convection Interactions Experiment (TRACER) C-band and operational S-band radar observations, Houston Lightning Mapping Array (HLMA) lightning data, and large-eddy simulations (LESs), this study investigates negative K_{DP} observed above the melting layer as a potential signature of SIP and storm electrification. Among 413 observed deep convective cells, 238 cases (~57.6%) exhibit negative K_{DP} above the melting layer, and 49 of 123 lightning-active cases (~39.8%) show negative K_{DP} near HLMA sources. These observations suggest that negative K_{DP} is associated with small, vertically oriented, electrically aligned ice particles. LESs with spectral bin microphysics reproduce negative K_{DP} with high likelihood only when SIP is included, especially fragmentation of freezing drops, which increases ice water path, decreases liquid water path, and reduces accumulated precipitation by ~11.2% in the examined case.

Plain Language Summary

Thunderstorms contain many small ice particles that influence lightning, rainfall, and how storms grow, but scientists still do not know exactly how these particles form in natural clouds. This study uses radar and lightning observations from the 2022 TRACER field campaign near Houston, Texas, together with high-resolution storm simulations, to examine a radar signature that may reveal these ice particles. The study found that many isolated deep convective clouds exhibited an unusual radar signature, called negative K_{DP}, above the melting layer, where ice particles are present. These signals were detected by both research and operational weather radars, and some occurred near lightning activity. This suggests that the signals may be caused by small ice particles that become vertically aligned by electric fields inside thunderstorms. Simulations showed that this radar signature is reproduced mainly when ice multiplication processes are included, especially when freezing drops break apart. This process increases ice, reduces liquid water, and decreases accumulated rainfall by about 11.2% in the examined case. These results help clarify how ice production, storm electrification, and precipitation are connected in deep convection.

1. Introduction

Specific differential phase, K_{DP}, is a polarimetric radar variable that describes the range derivative of differential propagation phase, Φ_{DP}, between horizontally and vertically polarized radar waves. In rain and in many populations of horizontally oriented anisotropic ice particles, K_{DP} is typically positive because horizontally polarized waves experience a larger phase shift than vertically polarized waves (Ryzhkov & Zrnić, 2019). Negative K_{DP}, particularly above the melting layer, is therefore an unusual and physically informative signature. It has commonly been interpreted as evidence that anisotropic ice particles have been rotated from their preferred aerodynamic, near-horizontal orientation toward a more vertical orientation by strong storm electric fields (Caylor & Chandrasekar, 1996; Ryzhkov & Zrnić, 2007; Weber et al., 2024; Zrnic & Ryzhkov, 1999). Because such electric fields develop during thunderstorm electrification, negative K_{DP} above the melting layer provides a potentially valuable radar-based indicator of ice particle alignment, charge structure, and lightning initiation processes.

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Early polarimetric radar studies reported signatures consistent with rapid changes in ice-particle orientation associated with thunderstorm electrification and lightning discharges. For example, Metcalf (1995) showed that hydrometeor reorientation associated with thunderstorm electrification could be observed using S-band polarimetric radar. Caylor and Chandrasekar (1996), using Florida Convection and Precipitation Electrification project (CaPE) observations, further linked decreases in K_{DP} prior to lightning and rapid recovery after lightning to the electrical alignment and relaxation of ice crystals. Their observations suggested that negative K_{DP} values, often on the order of -0.3 to $-0.5°$ km^{-1}, could develop in cloud-top ice regions several kilometers above the 0°C level. Subsequent studies expanded this phenomenology across different storm environments, radar wavelengths, and lightning data sets (e.g., Biggerstaff et al., 2017; Figueras i Ventura et al., 2013; Mattos et al., 2016, 2017; Wang et al., 2024a, 2024b), showing that negative K_{DP} is not simply collocated with lightning, but may mark the development of strong electric fields and charge-layer boundaries that control where lightning initiates and propagates. For example, Mattos et al. (2016, 2017) showed that K_{DP} in the glaciated and upper mixed-phase regions became increasingly negative as lightning activity increased, and that negative K_{DP} developed before the first cloud-to-ground lightning in isolated compact thunderstorms. Biggerstaff et al. (2017) provided a more detailed electrical interpretation by showing that intracloud flashes in a Florida mesoscale convective system (MCS) initiated near the zero-K_{DP} boundary surrounding a negative-K_{DP} region, rather than within the negative-K_{DP} core itself. Wang et al. (2024a, 2024b) showed that composite K_{DP} in an isolated thunderstorm changed from positive to negative several minutes before the first intracloud flash, with the most negative cloud-top values occurring shortly before lightning initiation.

However, the microphysical interpretation of negative K_{DP} is not straightforward. A negative propagation phase signature does not require all particles in the radar sampling volume to be vertically oriented. Large dry aggregates and graupel often have small intrinsic K_{DP}, while contributing strongly to reflectivity. As a result, negative K_{DP} may arise from a mixed population in which small, anisotropic, electrically aligned ice crystals dominate the propagation phase, while larger graupel or aggregates dominate Z_H and partially mask the expected Z_{DR} signature (Hubbert et al., 2014). Large rimed particles are also central to noninductive charge separation and storm electrification through collisions with ice crystals in mixed-phase regions, whereas smaller anisotropic ice crystals may be the primary carriers of the negative K_{DP} propagation signal (Berdeklis & List, 2001; Reynolds et al., 1957; Saunders, 2008; Saunders et al., 2006; Takahashi, 1978). Thus, negative K_{DP} above the melting layer should be interpreted as an integrated signature of particle shape, size distribution, orientation, and electric-field strength rather than as a simple hydrometeor category. This raises a central unresolved question: what supplies the small anisotropic ice particles responsible for negative K_{DP} above the melting layer?

Secondary ice production (SIP) can substantially enhance ice particle number concentrations in mixed-phase clouds under appropriate thermodynamic and microphysical conditions, beyond what is predicted by primary ice nucleation alone (Field et al., 2017; Korolev & Leisner, 2020). Observational and modeling studies have shown that SIP processes can occur over a broader temperature range than the classic Hallett–Mossop mechanism (-3 to -8°C; Hallett & Mossop, 1974) and can substantially alter ice number concentrations in convective clouds, thereby influencing cloud evolution, cloud electrification, storm evolution, and surface precipitation (e.g., Grzegorczyk, Wobrock, Dziduch, & Planche, 2025; Grzegorczyk, Wobrock, Canzi, et al., 2025a, 2025b; Huang et al., 2022; Korolev et al., 2024; Lawson et al., 2015; Lawson et al., 2022; Lawson et al., 2023; Phillips et al., 2018, 2020; Qu et al., 2022; Sun et al., 2025; Vongpaseut & Barthe, 2025). Huang et al. (2022) indicated that simulations including multiple SIP processes reproduced high-ice-water-content regions in tropical oceanic convective clouds across different temperature levels, with simulated ice number concentrations and radar reflectivity closely matching the observed magnitudes and vertical distributions. These results suggest that representing multiple SIP processes is important for improving numerical simulations of convective clouds. SIP processes therefore provide plausible pathways for replenishing the small ice particles that, once exposed to strong electric fields, may become vertically aligned and produce negative K_{DP}.

Recently, the second-generation C-band Scanning ARM (Atmospheric Radiation Measurement) Precipitation Radar (CSAPR2) observed more than 400 isolated deep convective cells under a wide range of environmental conditions during the Tracking Aerosol Convection Interactions Experiment (TRACER), conducted in the vicinity of Houston, Texas, in 2022 (Huang, McFarquhar, Patil, et al., 2025; Jensen et al., 2025; Mise et al., 2026); the National Science Foundation (NSF)-sponsored Experiment of Sea Breeze Convection, Aerosols, Precipitation, and Environment (ESCAPE; Kollias et al., 2025; Patil et al., 2026), a shorter-duration synergistic field

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campaign, overlapped with part of TRACER in the same region. An examination of these cells revealed frequent occurrences of negative K_{DP} regions located above the melting layer. We hypothesize that these negative K_{DP} signatures arise from small nonspherical ice particles produced by SIP processes in mixed-phase regions and subsequently aligned toward a vertical orientation by the in-cloud electric field. The hypothesis is tested by combining unique TRACER observations and large-eddy simulations (LESs) with the Weather Research and Forecasting (WRF) model. The scientific questions addressed in this study are as follows:

- To what extent can the WRF model reproduce regions of negative K_{DP} observed above the melting layer during TRACER, with and without SIP processes?
- Which specific SIP mechanism exerts the primary control on the formation of these negative K_{DP} signatures?
- How do SIP processes influence characteristics of convective clouds?

The remainder of the paper is organized as follows. Section 2 describes the TRACER observations and LES numerical experiments. Section 3 presents results from both observations and modeling. Section 4 summarizes the conclusions and discussion.

2. Data and Method

2.1. Observations

This study utilizes unique CSAPR2 radar observations collected during the TRACER Intensive Operational Period (IOP), conducted from June to September 2022, to examine negative K_{DP} signatures observed above the melting layer. CSAPR2 is a C-band, dual-polarization Doppler weather radar with a maximum range of 110 km and a range resolution of 100 m. During TRACER, CSAPR2 employed the Multisensor Agile Adaptive Sampling (MAAS) framework (Kollias et al., 2020) for real-time storm tracking. The CSAPR2 optimized convective cell tracking data set (Oue et al., 2023), which includes Plan Position Indicator (PPI) and Range Height Indicator (RHI) scans, provides a comprehensive collection of high-resolution radar observations of convective cells. In addition to CSAPR2 observations, data from the nearest operational Next-Generation Weather Radar (NEXRAD) S-band dual-polarization radar, KHGX, are used as supplementary observations. Through manual examination of both CSAPR2 and KHGX data, a total of 413 deep convective cells were identified for analysis (Mise et al., 2026). Houston Lightning Mapping Array (HLMA) flash-level data during ESCAPE (Bruning et al., 2024; Logan et al., 2023) are used to examine lightning activity associated with the deep convective cells.

2.2. Model Configuration

In this study, quasi-idealized LES simulations are conducted using the WRF model version 4.7.1 with a fast spectral bin microphysics (FSBM) scheme (Shpund et al., 2019) at a horizontal grid spacing of 250 m, which is the inner edge of horizontal grid spacings for convergence of convective vertical velocities (Jeevanjee, 2017). The horizontal domain size is 100 km × 100 km, and the model uses 81 vertical levels at a grid spacing of ∼250 m extending up to 20 km. The initial horizontally homogeneous thermodynamic environment is specified using hourly model-level fields from the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis v5 (ERA5) data set (Hersbach et al., 2020), provided at a horizontal grid spacing of 0.25°. This data set has been evaluated against all available TRACER IOP radiosonde soundings at the TRACER sites, with mean biases in temperature, relative humidity, and horizontal winds falling within the uncertainty ranges of the radiosonde sensors (Huang, McFarquhar, Patil, et al., 2025; Mise et al., 2026). Convection initiation adopts an updraft nudging method, as in our previous study (Huang et al., 2022).

In the FSBM scheme, aerosols are represented by a prognostic size distribution discretized into 43 logarithmically spaced dry-size bins. To reduce computational cost, ice particles are represented by two size distributions: one for low-density ice particles, including ice crystals ($D \leq 263.22$ μm) and snow ($D > 263.22$ μm), and one for high-density ice particles, represented by hail in this study (Shpund et al., 2019). The scheme represents droplet activation, condensation and evaporation, ice nucleation and droplet freezing, deposition and sublimation, collision–coalescence, raindrop breakup, ice aggregation, riming, and melting (Shpund et al., 2019). In addition to the Hallett-Mossop (HM, or rime splintering) mechanism included in the original FSBM scheme, parameterizations of three additional SIP mechanisms, namely fragmentation of freezing drops (FFD, Phillips et al., 2018), fragmentation during sublimation (SUBK, Deshmukh et al., 2022), and fragmentation during ice–ice collisions (FIIC, Phillips et al., 2017), are implemented. The hail option of the FSBM scheme is used, so fragmentation

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during hail–hail and hail–snow collisions is considered in this study. Detailed parameterizations of FFD, SUBK, and FIIC are described in Phillips et al. (2018), Deshmukh et al. (2022), and Phillips et al. (2017), respectively.

To identify the location of dominant processes, especially SIP processes, responsible for producing negative K_{DP} regions, sensitivity experiments on SIP processes, including all four SIP processes off (SIPsOff), only one SIP process on (HMOn: only the Hallett-Mossop process on; FFDOn: only fragmentation of freezing drops on; SUBKOn: only fragmentation during sublimation; FIICOn: only fragmentation during ice–ice collisions), and all four SIP processes on (SIPsOn), are performed. In order to compare directly with the TRACER CSAPR2 observations, dual-polarization radar variables in each simulation are derived using a C-band radar observation operator (Ryzhkov et al., 2011).

3. Results

3.1. Observed Negative K_{DP}

Manual examination of CSAPR2 RHI scans from 413 deep convective cells identifies 238 cases (~57.6%) in which at least one scan contains a spatially coherent region of negative K_{DP} above the 0°C level. Isolated negative values confined to one or a few radar gates are excluded. To assess the reproducibility of this classification, the RHI scans are independently examined by two coauthors. Our classifications differ for less than 1.7% of the cells, indicating high inter-reviewer consistency. Many of these negative K_{DP} signatures are also evident in the collocated observations from the NEXRAD S-band radar KHGX (e.g., Figure S1 in Supporting Information S1), providing additional support for the observed signatures, although agreement between the two radars alone does not rule out retrieval artifacts.

Lightning activity is evaluated using quality-controlled HLMA very-high-frequency (VHF) source locations. HLMA events are considered valid if they have a reduced chi-square value $\chi^2 < 2.0$ and are detected by at least six stations. The reduced χ^2 value measures the normalized timing-residual misfit of the HLMA VHF source-location solution, with values closer to 1 indicating better agreement between observed and fitted station arrival times. This quality-control criterion retains well-constrained lightning source locations and filters out poorly constrained solutions (Thomas et al., 2004). To associate the quality-controlled HLMA sources with individual RHI scans, sources are required to occur within ±30 s of the scan time and within 1° of the RHI azimuth. A cell is classified as lightning active when at least one valid HLMA source satisfies these temporal and azimuthal criteria during CSAPR2 sampling interval. Based on this definition, 123 of the 413 cells (~29.8%) are classified as lightning active. This percentage represents a lower-bound estimate because HLMA observations are unavailable for 47 convective cells sampled during 26–30 June 2022. The larger number of negative-K_{DP} cases (238) than lightning-active cases (123) indicates that negative K_{DP} was not always accompanied by lightning detected by HLMA during the sampled RHI scans. This discrepancy may arise from the portion of the convective-cell life cycle sampled by CSAPR2, the limited two-dimensional spatial sampling of the RHI scans, or variations in HLMA detection efficiency and station geometry. Alternatively, some cells may have contained in-cloud electric fields strong enough to modify the canting-angle distribution of ice particles but below the threshold required to initiate detectable lightning.

Among the 123 lightning-active cases, 49 cases (~39.8%) exhibit negative K_{DP} above the melting layer together with lightning activity near the negative K_{DP} region. These cases include both simultaneous occurrence within the same RHI scan and short lead-lag relationships, in which negative K_{DP} precedes lightning activity or lightning activity precedes enhanced negative K_{DP} over one or more RHI scan intervals. This count should be interpreted as a conservative subset of the lightning-active cases, because spatial proximity and/or temporal consistency between lightning and the negative K_{DP} region are required. If only simple co-occurrence is considered, where negative K_{DP} and lightning activity occur anywhere within the same cell without requiring spatial or temporal correspondence, the number of overlapping cases would be larger. Because the RHI scans provide only intermittent two-dimensional slices through evolving three-dimensional convective cells, the spatial and temporal relationship between negative K_{DP} regions and lightning activity needs to be examined on a case-by-case basis. An isolated convective case from 24 August 2022 was selected for its clear, spatially coherent negative K_{DP} above the 0°C level in multiple CSAPR2 RHI scans, concurrent NEXRAD observations, and active lightning (Figure 1). The K_{DP} fields shown in Figure 1c1–1c4 are estimated using the current U.S. National Weather Service (NWS) operational algorithm (Reimel & Kumjian, 2021; Ryzhkov et al., 2005). These estimates are consistent in

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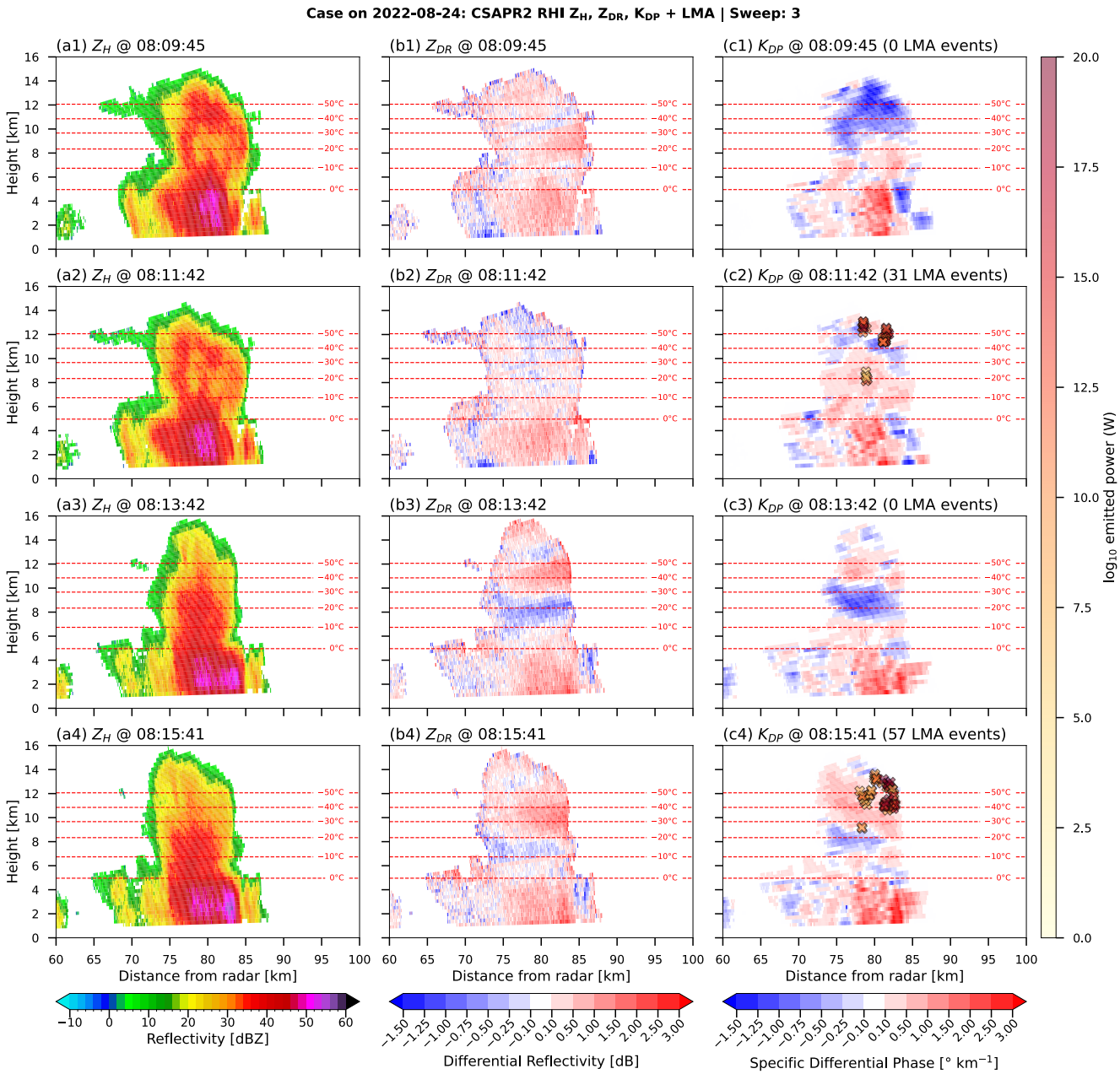

Figure 1. CSAPR2 RHI observations for the 24 August 2022 case, showing the evolution of storm structure and lightning activity from 08:09:45 to 08:15:41 UTC. Columns show (a1–a4) horizontal reflectivity (Z_H, in dBZ), (b1–b4) differential reflectivity (Z_{DR}, in dB), and (c1–c4) specific differential phase (K_{DP}, in ° km^{-1}). HLMA sources are overlaid on the K_{DP} panels where present, with marker color indicating $\log_{10}$ emitted power in W; event counts are given in each panel title. Red dashed lines denote ERA5 environmental isotherms from 0 to −50°C at 10°C intervals.

magnitude and spatial patterns with those in the CSAPR2 optimized convective-cell-tracking data set (Oue et al., 2023), where K_{DP} is estimated using both the technique of Hubbert and Bringi (1995) and its updated version in the LiDAR Radar Open Software Environment (LROSE) (Figure S1 in Supporting Information S1).

In this convective cell observed by CSAPR2 between 08:09:45–08:15:41 UTC, the Z_H fields indicate a persistent deep convective echo extending above the −50°C level, with the strongest reflectivity located in the low-level core region below the 0°C level (Figures 1a1–1a4). The corresponding Z_{DR} fields show positive values mainly below and near the 0°C level, consistent with liquid or melting hydrometeors, while reduced or negative Z_{DR} appears in the upper mixed-phase and glaciated regions above the 0°C level (Figures 1b1–1b4). In the K_{DP} field at 08:09:45 UTC,

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positive values are mainly confined to the lower part (below the −10°C level) of the convective cell, whereas pronounced negative K_{DP} values $< -1°$ km^{-1} appear above the −20°C level (Figure 1c1). About 2 min later (08:11:42 UTC), HLMA sources are observed and concentrated near the same upper-level negative K_{DP} region while the magnitude of negative K_{DP} decreased (Figure 1c2). This sequence then repeated during the cell evolution, with newly formed negative K_{DP} followed by subsequent lightning enhancement (Figures 1c3 and 1c4).

The coherent near-zero or weakly negative Z_{DR} signatures, particularly those in Figures 1b3 and 1b4, broadly coincide with the negative K_{DP} regions in Figures 1c3 and 1c4. Near-isometric graupel or aggregates can mask the intrinsic Z_{DR} of anisotropic ice particles by contributing approximately equal horizontal and vertical backscatter, driving Z_{DR} toward 0 dB. This reflects a fundamental difference between amplitude and phase polarimetric measurements: near-isometric particles can strongly influence Z_{DR} through their backscatter but contribute negligibly to differential propagation phase and therefore do not similarly dilute K_{DP}. Consequently, negative K_{DP} produced by electrically aligned ice crystals can coexist with near-zero Z_{DR} when near-isometric particles dominate backscatter (Hubbert et al., 2014). The negative K_{DP} is consistent with a population of anisotropic ice particles having a net vertical or canted orientation. Because CSAPR2 employs simultaneous H/V transmission, however, cross-coupling associated with canted, aligned ice crystals may also influence the magnitude and sign of the measured Z_{DR}. Accordingly, negative Z_{DR} is treated here as supporting evidence for electrical alignment rather than as a unique indicator of particle habit.

Manual examination indicates that 28 of the 49 cases (∼57.1%) exhibit a similar sequence, in which negative K_{DP} signatures above the 0°C level are followed by enhanced lightning activity (not shown). This temporal evolution is consistent with electrical alignment of small nonspherical ice particles before lightning and their relaxation toward a near-horizontal aerodynamic orientation after a discharge weakens the local electric field. Previous polarimetric studies reported rapid changes in radar-inferred particle orientation associated with lightning discharges (e.g., Hendry & McCormick, 1976; Krehbiel et al., 1996; Metcalf, 1995; Metcalf, 1997). The physical basis for electrical alignment is independently supported by laboratory experiments (Foster & Hallett, 2002; Saunders & Rimmer, 1999) and calculations of electrical and aerodynamic torques (Weinheimer & Few, 1987). However, this interpretation remains an inference for the TRACER cases because particle habit and ambient orientation were not directly measured in situ. In the next section, high-resolution LES simulations and sensitivity experiments are used to examine how SIP processes contribute to the development of these negative K_{DP} signatures above the melting layer.

3.2. Simulated Negative K_{DP}

To examine whether numerical models can reproduce the observed negative-K_{DP} region above the melting layer, and to evaluate the role of SIP processes in producing this polarimetric signature, LES simulations and sensitivity experiments described above are conducted for the representative isolated deep convection case on 24 August 2022 (Figure 1). Thermodynamic and kinematic profiles for this case (Figure S2 in Supporting Information S1) are extracted from the nearest ERA5 grid point at 0700 UTC on 24 August 2022, about one hour before convection initiation, and are used as the model input sounding. The sounding features a near-surface inversion and a surface-based convective available potential energy of 1793 J kg^{-1} (Figure S2 in Supporting Information S1). The prescribed aerosol size distribution is represented by three lognormal modes, with total number concentrations of 818.5, 660.0, and 380.0 cm^{-3}; median diameters of 23.7, 57.0, and 170.0 nm; and geometric standard deviations of 1.578, 1.420, and 1.477, respectively (Figure S3 in Supporting Information S1). These parameters are derived from mean scanning mobility particle sizer (SMPS) surface observations (Kuang et al., 2024) collected between 0600 and 0700 UTC on 24 August 2022 during the TRACER field campaign. The vertical aerosol profile is constructed by applying the vertical scaling factor from the TRACER Model Intercomparison Project (MIP) (Fan et al., 2024) to the surface aerosol concentration.

The simulations with (SIPsOn) and without (SIPsOff) SIP reproduce the observed deep convective structure, with similar echo-top heights of ∼16 km compared with the CSAPR2 RHI observations (Figure S4 in Supporting Information S1). However, the simulated Z_H is generally stronger than observed, especially in the lower to middle troposphere (below the −20°C level). In particular, the simulated 50–60 dBZ core extends to ∼7–8 km, whereas observed Z_H exceeding 50 dBZ is mainly confined below ∼5 km. Similarly, simulated reflectivity exceeding 40 dBZ extends to ∼14 km, compared with ∼12 km in the observations. These differences indicate a vertical over-

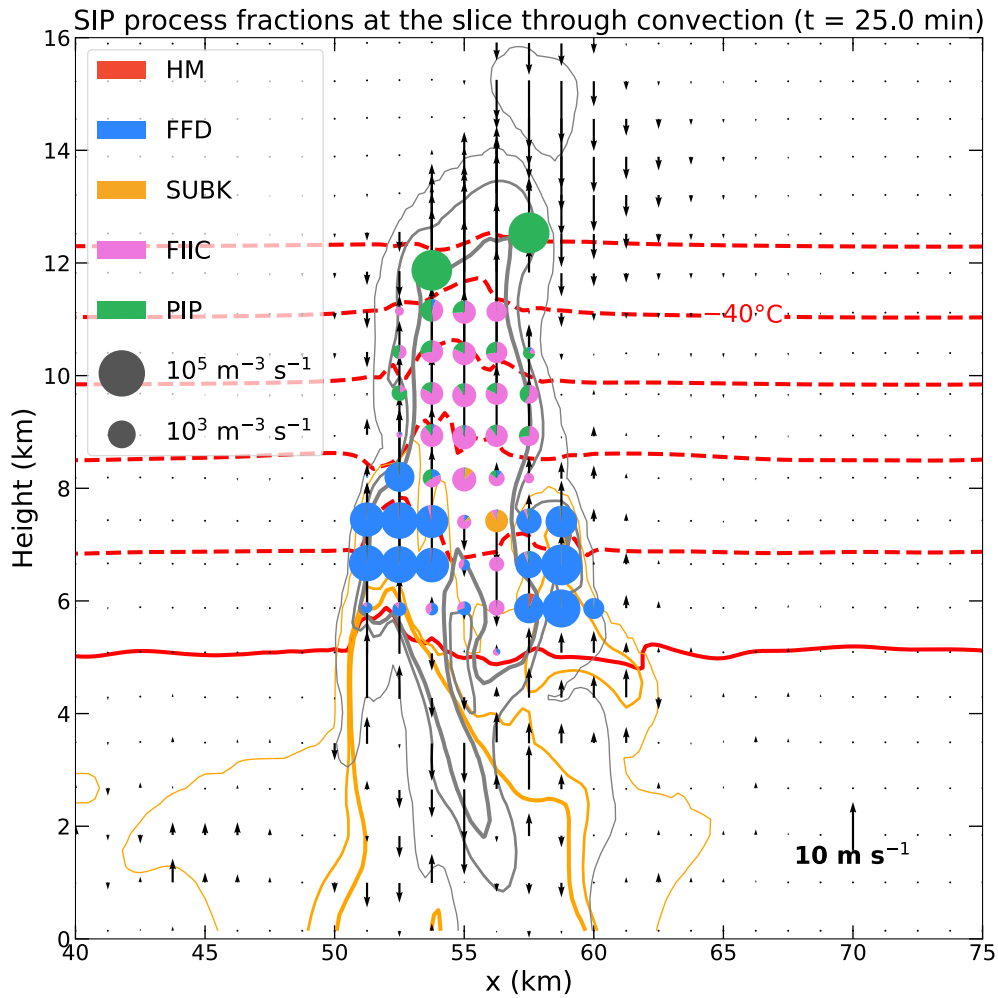

Figure 2. Vertical cross section through the simulated convective cell in the SIPsOn experiment at $t = 25$ min, near the time of peak total water path (Figure 4c), showing the fractional contributions of ice-particle production processes to the ice particle number production rate. Processes include primary ice production (PIP; green), including both homogeneous and heterogeneous ice nucleation, and secondary ice production by HM (red), FFD (blue), SUBK (orange), and FIIC (magenta). Marker size indicates the total ice particle number production rate. Gray and orange contours show ice water content (IWC) and liquid water content (LWC), respectively, at 0.01, 1.0, and 2.0 g m^{-3}. Red contours denote temperature isotherms from 0 to −50°C at 10°C intervals. Vectors show vertical velocity, with the reference vector corresponding to 10 m s^{-1}.

extension of high Z_H by ∼2–3 km, primarily associated with excessive simulated large hail above the freezing level and large raindrops below it (not shown). A future implementation of the hail-melting scheme proposed by Ryzhkov et al. (2013) may help reduce this bias. Compared with SIPsOff, SIPsOn slightly lowers the upper extent of the $Z_H \geq 50$ dBZ core and reduces its horizontal coverage (Figure S4 in Supporting Information S1). This reduction suggests that SIP redistributes condensate from large precipitating particles toward smaller ice particles, thereby reducing the positive Z_H bias. Nevertheless, a substantial bias remains, with the simulated high-Z_H core extending ∼2 km higher than observed. To further examine the relative contributions of ice-particle production processes in convection, we calculate the fractional contribution of each ice-production process to the total ice particle number production rate along a cross section through the convective cell at $t = 25$ min (Figure 2). Within the mixed-phase updraft core above the freezing level ($x = \sim 50$–55 and ∼57.5–60 km, at heights of ∼6–8.5 km), FFD dominates ice particle number production, with the largest production rates concentrated near −10°C (∼7 km height), outside the typical temperature range for HM (−3 to −8°C). Outside the liquid-rich updraft region, primary ice production (PIP) dominates ice particle number production mainly at temperatures <− 40°C. FIIC dominates in the ice-rich region between about −20 and −40°C (∼8–11 km height) and is also locally dominant near −15°C within the dendritic growth layer (about −10 to −20°C), but its ice particle number

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production rate is ∼1–2 orders of magnitude lower than that of FFD. At a later stage (t = 35 min), FIIC dominates ice particle number production above the 0°C level as supercooled liquid water becomes depleted (not shown). This regional and temporal dependence is consistent with Huang et al. (2022), who reported FFD dominance in warmer mixed-phase regions during the early developing stage of convection and FIIC dominance in colder regions and during later stages. These results suggest that ice particle number production in the simulated convective cell is mainly controlled by FFD in the active mixed-phase updraft, while PIP and FIIC dominate in regions outside the liquid-rich updrafts where supercooled liquid water is limited (less than 0.01 g m^{-3}). Advection further redistributes small ice particles, so their concentration maxima need not coincide with regions of maximum local production.

To examine K_{DP} across all sensitivity experiments, K_{DP} is calculated using the C-band dual-polarization radar simulator of Ryzhkov et al. (2011). Weinheimer and Few (1987) calculated electrical and aerodynamic torques for thunderstorm ice particles and showed that, for electric fields of order 100 kV m^{-1}, electric-field alignment can occur for crystals with major dimensions from roughly 200 μm to 1 mm, depending on particle habit. The FSBM scheme used in this study does not include parameterizations of charging and discharging processes and therefore cannot explicitly simulate in-cloud electric fields. To evaluate the proposed mechanism within this limitation, small ice particles ($D \leq$ 263.22 μm, corresponding to the size threshold for ice crystals in FSBM) are assumed to be vertically oriented in the radar simulator as a sensitivity-test framework. A sensitivity test using a larger size threshold of $D \leq$ 823.88 μm produces a similar K_{DP} distribution (not shown).

Figure 3 decomposes the simulated K_{DP} field into hydrometeor contributions to identify the processes responsible for negative K_{DP} above the melting layer. In all experiments, drops produce strong positive K_{DP} below and near the melting layer (Figures 3a1–3f1), whereas the hail contribution is relatively weak and mostly positive (Figures 3a3–3f3). In contrast, pronounced negative K_{DP} (<−0.5° km^{-1}) appears in the snow contribution (including small ice crystals) only when FFD is included, as shown in the FFDOn and SIPsOn experiments (Figures 3c2, 3c4, 3f2, and 3f4). This negative snow K_{DP} signal extends through the upper mixed-phase and glaciated regions (above ∼7 km), where small ice particles are assumed to be vertically oriented in the radar simulator. The other experiments exhibit weaker negative snow and total K_{DP} (>−0.5° km^{-1}), but mainly within a limited region near 13–15 km. In FFDOn and SIPsOn, negative K_{DP} extends to altitudes of up to 16 km (Figures 3c2, 3c4, 3f2, and 3f4), whereas the observed signature is mainly confined below 15 km (Figures 1c1–1c4). This vertical-extent bias may result from stronger simulated updrafts or the simplified ice-particle orientation assumption used in the radar simulator. Quantifying the contributions of these factors is beyond the scope of this study and will be examined in future work. The more limited vertical extent and weaker magnitude of negative K_{DP} in SIPsOff, HMOn, SUBKOn, and FIICOn suggest that HM, SUBK, and FIIC alone are insufficient to fully reproduce the observed vertical structure and magnitude of negative K_{DP} signatures under the prescribed particle orientation.

For this simulated case, these results suggest that SIP processes, especially FFD, play an important role in enhancing concentrations of small ice particles and reproducing negative K_{DP} features above the melting layer. However, the magnitudes of simulated negative K_{DP} below ∼13 km in FFDOn and SIPsOn (Figures 3c2, 3c4, 3f2, and 3f4) remain smaller than observed (Figures 1c1 and 1c3). This underestimation may reflect biases in simulated ice particle number concentrations or SIP parameterizations, as well as uncertainties in the particle shape, density, and orientation assumptions used in the radar simulator (Hubbert et al., 2014). Future collocated in-situ measurements are needed to assess these potential biases and uncertainties.

3.3. SIP Impacts on Convective Clouds

To further examine how SIP processes influence convective cloud evolution, Figure 4 shows time series of ice water path (IWP), liquid water path (LWP), total water path (TWP), and mean accumulated precipitation for the SIP sensitivity experiments of the 24 August 2022 case. Figure 4 shows that SIP processes can modify the simulated bulk water budget and precipitation efficiency of the simulated convective cell. All experiments exhibit rapid cloud development during the first ∼30 min, with peak IWP of ∼0.12–0.16 kg m^{-2} (Figure 4a), peak LWP of ∼0.17–0.19 kg m^{-2} (Figure 4b), and peak TWP of ∼0.27 kg m^{-2} (Figure 4c). After this developing stage, the experiments including FFD, i.e., FFDOn and SIPsOn, maintain larger IWP, with values remaining near 0.05 kg m^{-2} by 120 min, compared with ∼0.02 kg m^{-2} in SIPsOff, HMOn, SUBKOn, and FIICOn (Figure 4a). In

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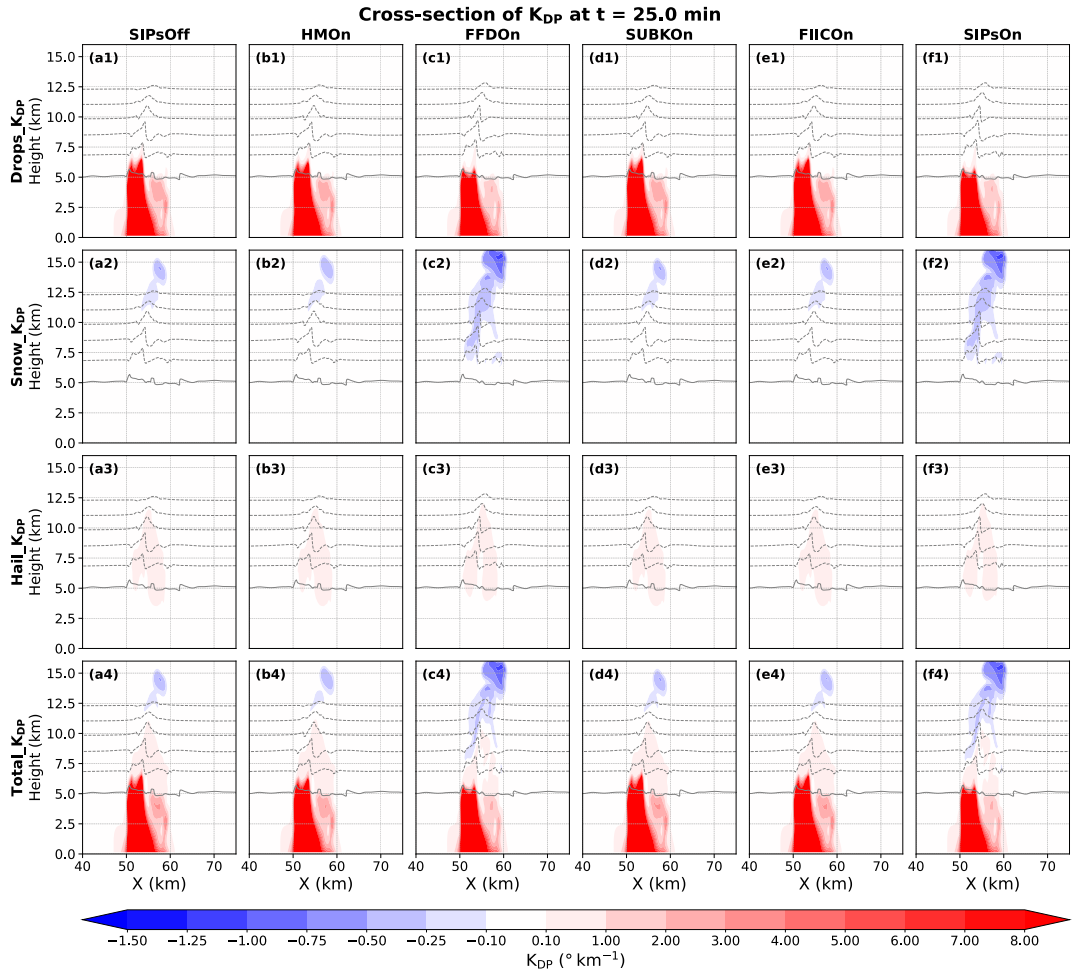

Figure 3. Simulated cross sections of hydrometeor-specific and total specific differential phase (K_{DP}; ° km^{-1}) through the convective cell at $t = 25$ min for the (a1–a4) SIPsOff, (b1–b4) HMOn, (c1–c4) FFDOn, (d1–d4) SUBKOn, (e1–e4) FIICOn, and (f1–f4) SIPsOn experiments. Rows show the contributions from drops, snow (including small ice crystals), hail, and total K_{DP}. Gray contours denote temperature isotherms from 0 to −50°C at 10°C intervals. The radar simulator assumes vertically oriented small ice particles ($D \leq 263.22$ μm).

contrast, LWP decreases rapidly after its peak and becomes very small (less than 0.01 kg m^{-2}) after ∼60 min in all experiments, with generally lower values when FFD is included (Figure 4b). The enhanced IWP and TWP, together with reduced LWP, suggest that FFD shifts condensate from liquid water toward ice-phase particles and helps maintain more condensate in the cloud column, with TWP increasing by up to ∼155% relative to SIPsOff during the later stage. The mean accumulated precipitation approaches a plateau after ∼60–70 min, reaching ∼0.24 mm in SIPsOn and FFDOn, compared with ∼0.27 mm in SIPsOff (Figure 4d). In SIPsOn, this corresponds to a reduction of ∼11.2% relative to SIPsOff. HMOn, SUBKOn, and FIICOn remain close to SIPsOff across all four quantities (Figure 4), indicating that HM, SUBK, and FIIC have much smaller impacts on the bulk water budget and accumulated precipitation than FFD. Therefore, these results suggest that FFD is the SIP pathway most responsible for shifting condensate from liquid water toward ice-phase particles and for reducing accumulated surface precipitation in this simulated convective cell. The weaker responses in HMOn, SUBKOn, and FIICOn suggest that these processes alone have limited impact on the domain-integrated water budget compared with FFD.

To further explain the changes in the bulk water budget shown in Figure 4, vertical profiles of hydrometeor water content (Figure S5 in Supporting Information S1) and major microphysical process rates (Figure S6 in Supporting Information S1) are compared between SIPsOn and SIPsOff. Compared with SIPsOff, SIPsOn produces more cloud ice ($D \leq 263.22$ μm) in the upper mixed-phase and glaciated regions above the −10°C level at

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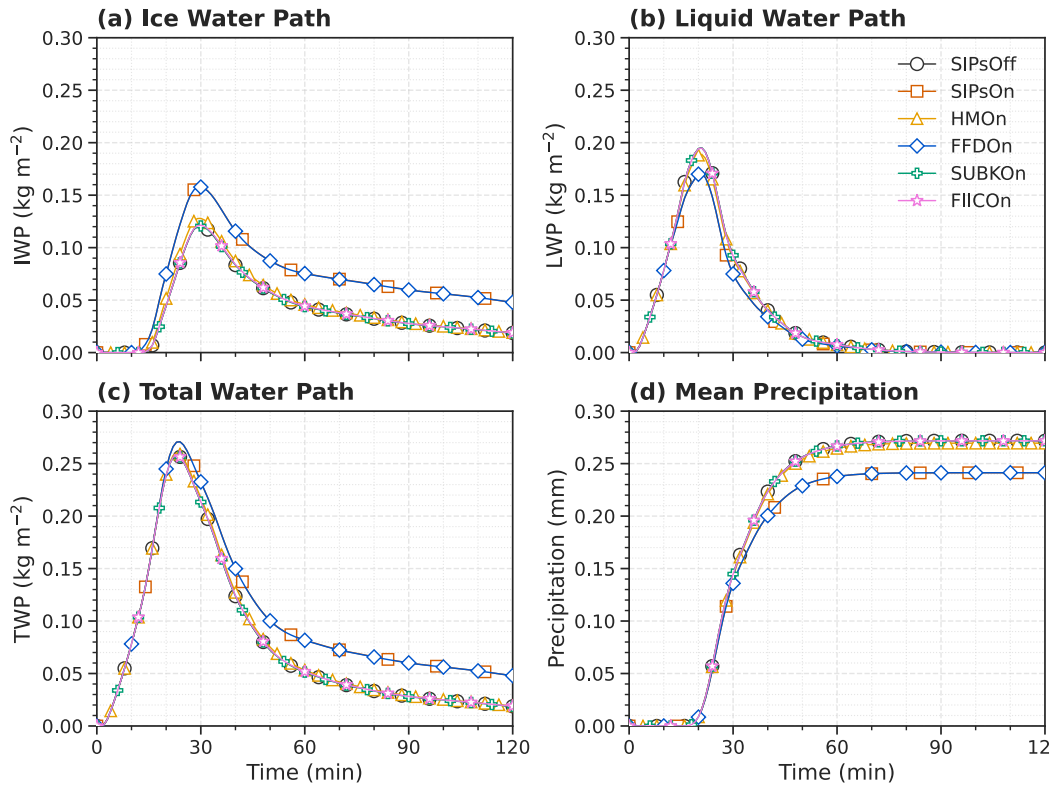

Figure 4. Time series of (a) ice water path (IWP, in kg m^{-2}), (b) liquid water path (LWP, in kg m^{-2}), (c) total water path (TWP, in kg m^{-2}), and (d) mean accumulated precipitation (mm) for the SIP sensitivity experiments, including SIPsOff, SIPsOn, HMOn, FFDOn, SUBKOn, and FIICOn.

t = 20–40 min, while cloud water (D < 101.59 μm), rain (D ≥ 101.59 μm), snow (D > 263.22 μm), and hail water contents are generally reduced or remain comparable after the developing stage (t = 30–40 min; Figure S5 in Supporting Information S1). This indicates that the larger IWP in SIPsOn is mainly associated with enhanced small-ice mass rather than increased precipitation-sized hydrometeors. The reduced rain and hail water contents are also consistent with the smaller accumulated precipitation in SIPsOn. During the active convective stage (t = 20–30 min), SIPsOn modifies ice-phase growth near and above the melting layer, with enhanced ice/snow depositional growth and reduced riming growth of hail aloft compared with SIPsOff (Figure S6 in Supporting Information S1). These changes suggest that SIP, especially FFD, increases the number of small ice particles, redistributes condensate from liquid water and large precipitating particles toward cloud ice (D ≤ 263.22 μm), and reduces precipitation efficiency. Therefore, the larger IWP, lower LWP, and ∼11.2% reduction in accumulated precipitation in SIPsOn are physically consistent with a microphysical shift toward smaller ice particles and weaker production or fallout of large precipitation-sized hydrometeors.

These results apply to a deep convective case with observed lightning activity and echo tops near 16 km. Han et al. (2024) reported surface precipitation reductions of up to 20% in simulated deep convection with cloud-top temperatures below −60°C, relative to a control with rime splintering as its only SIP mechanism. Lawson et al. (2015) initially selected developing cumulus clouds with estimated cloud-top temperatures higher than −12°C and documented FFD-related rapid glaciation between −8 and −15°C during subsequent growth. In contrast, Calderón et al. (2025) examined cumulus congestus sampled as deep as 300 m below cloud top, with a minimum sampled temperature near −18°C, and simulated a 26% increase in domain-mean cumulative precipitation when multiple SIP mechanisms were included, with enhancement occurring mainly later in development. These comparisons suggest that the precipitation response may depend on cloud conditions, developmental stage, and the represented SIP mechanisms and should not be generalized from this case.

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4. Summary and Conclusions

In this study, TRACER radar observations, lightning observations, and WRF-LES sensitivity simulations are used to investigate the occurrence of negative K_{DP} above the melting layer in deep convective clouds and to evaluate its possible microphysical origin. The major findings are summarized as follows:

1. Negative K_{DP} regions are frequently identified above the melting layer in convective clouds. Their consistent detection by both TRACER research C-band and operational S-band radars increases confidence in the observed signatures, although cross-radar agreement alone does not rule out biases arising from shared processing assumptions or other retrieval effects.
2. The negative K_{DP} regions are closely associated with lightning activity. Their spatial and temporal relationship with HLMA sources is consistent with electrical alignment of small nonspherical ice particles in the upper mixed-phase and glaciated regions of deep convection. However, particle habit and orientation remain inferred because collocated in-situ measurements are unavailable.
3. WRF-LES sensitivity simulations show that negative K_{DP} is reproduced with high likelihood only when SIP processes are included, especially FFD. For the examined deep convective case, including SIP processes increases ice water path, decreases liquid water path, and reduces accumulated surface precipitation by ∼11.2%, primarily through FFD. This response may vary with cloud conditions and model configuration and should not be interpreted as a universal precipitation response to FFD.

To the best of our knowledge, this study provides the first combined observational and modeling support for negative K_{DP} above the melting layer as a potential radar signature of electrically aligned nonspherical ice particles *associated with SIP* and storm electrification. Joint polarimetric radar and lightning observations offer a means of examining the spatial distribution and temporal evolution of these signatures in regions difficult to sample in situ. These findings can serve as observational benchmarks for future evaluations of ice microphysical processes in models that account for particle alignment and storm electrification. However, negative K_{DP} alone does not uniquely identify SIP, because it reflects the combined effects of particle properties and orientation. Despite several decades of research and the proposal of multiple SIP mechanisms, such as rime splintering, fragmentation of freezing droplets, and fragmentation during ice–ice collisions, large uncertainties remain in their quantitative rates, timing, and required environmental conditions in natural cloud systems, and their parameterizations in numerical models. Meanwhile, uncertainties in SIP process rates in numerical models arise not only from SIP parameterizations themselves, but also from the simulated hydrometeor populations produced by other microphysical processes and by different microphysics schemes. These uncertainties can further affect the simulated influence of SIP on surface precipitation, such as the contrasting precipitation responses reported in previous studies (Calderón et al., 2025; Georgakaki et al., 2022; Grzegorczyk, Wobrock, Canzi, et al., 2025a; Grzegorczyk, Wobrock, Dziduch, & Planche, 2025; Han et al., 2024; Sun et al., 2025). Although the synergistic ESCAPE field campaign collected in-situ observations of convective clouds, these measurements were not collocated with the convective cells sampled by CSAPR2 during TRACER. In addition, the ESCAPE research flights were conducted mainly below the −10°C level (Huang, McFarquhar, Patil, et al., 2025; Kollias et al., 2025), generally lower than the regions where negative K_{DP} signatures were observed. Future work should combine in-situ ice particle observations, in-cloud electric field measurements, laboratory experiments, and model developments that explicitly represent charging and discharging processes and better constrain SIP parameterizations using both field observations and laboratory measurements. Implementing such parameterizations in cloud-resolving models will be important for directly simulating in-cloud electric fields and further testing the proposed link among SIP, electrically aligned ice particles, negative K_{DP}, and lightning activity.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

TRACER CSAPR2 optimized convective cell tracking data set is available at https://doi.org/10.5439/1969992 (Oue et al., 2023). NEXRAD radar data set is available at the NOAA AWS S3 bucket (https://registry.opendata.aws/noaa-nexrad). The Houston Lightning Mapping Array data are available at https://doi.org/10.26023/GBKS-E7VT-HS11 (Logan et al., 2023). TRACER SMPS aerosol data are available at https://doi.org/10.5439/1476898

(Kuang et al., 2024). ERA5 reanalysis data set is available at https://doi.org/10.5065/XV5R-5344 (Hersbach et al., 2020). The WRF code is available at https://github.com/wrf-model/WRF (https://doi.org/10.5065/D6MK6B4K).

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