

1 **Warm-Sector Heavy Rainfall over the Eastern China Plains:**
2 **Identification, Characteristics, and Precursor Dynamic Signals**

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Abstract

Warm-sector heavy rainfall (WSHR) poses forecasting challenges, while its contribution to regional heavy rainfall and short-term precursors remain poorly quantified. Drawing on 2021–2025 warm-season (June–September) observations from 493 surface rain gauges, ERA5 reanalysis, and a radar wind profiler (RWP) mesonet across the eastern China plains, we build an objective identification framework integrating automated front detection, warm-sector mask, and station-specific 99th-percentile hourly rainfall thresholds to separate WSHR from other heavy rain events. We detect 477 station-based WSHR events alongside 6002 other heavy rainfall events. WSHR is spatially scattered, peaks in July–August, and preferentially begins at 2000–0300 Beijing Time. Accounting for merely 5–10% of annual heavy rainfall occurrences, WSHR has shorter durations yet comparable median rainfall accumulations and heavier extreme rainfall tails. Among the 297 stations recording at least one WSHR event, the mean station-level contribution of WSHR to heavy-rainfall accumulation is 13.8%, with a maximum of 48.1%. Composite ERA5 fields show that WSHR initiates within an 850-hPa moist tongue of high pseudo-equivalent potential temperature south of the front, superimposed on strengthening low-level moisture flux convergence. RWP composites of 62 WSHR events show increasing low-level jet (LLJ) occurrence and enhanced low-level ascent shortly before rainfall onset. Comparison with 271 warm-sector non-heavy rainfall events indicates a more pronounced late increase in LLJ occurrence and a stronger upper tail of ascent speeds at 1–2 km in WSHR, although the distributions overlap substantially. These observations identify candidate kinematic indicators whose event-level predictive skill remains to be evaluated.

1. Introduction

Warm-sector heavy rainfall (WSHR) occurs in the warm, moist air mass well ahead of a surface front, under weak synoptic-scale forcing (Huang et al., 1986; Sun et al., 2019; Wu et al., 2020). Because organized frontal lifting is absent, WSHR tends to be highly localized, to develop abruptly, to move slowly, and to concentrate intense rainfall over a confined area, producing disproportionate hazards through urban waterlogging, flash floods, and downstream river flooding (Zhang and Meng, 2019; Fu et al., 2023). As such, it is notoriously difficult to predict: operational global and regional models retain low quantitative precipitation forecast skill for WSHR events (Luo et al., 2017; Huang et al., 2025). Under weak synoptic-scale forcing, convective initiation is instead governed by mesoscale and boundary-layer processes, such as local convergence lines and boundary-layer jets, which operational models and conventional observations resolve only poorly (Wilson and Schreiber, 1986; Du and Chen, 2019). The regime therefore combines high societal impact with poor predictability.

Originally identified in South China during the pre-summer rainy season (Huang et al., 1986), WSHR has since been documented in several other regions of the country, including North China (Fu J. et al., 2020), the Yangtze–Huaihe coastal area (Yu et al., 2024), and the Sichuan Basin (Sun et al., 2019). The eastern China plains, which comprise the North China Plain and the middle-and-lower Yangtze Plain, are no exception. The region has experienced a number of WSHR events, including warm-shear-related events along the Yangtze–Huaihe coast during 2010–2017 (Yu et al., 2024), the localized rainstorm in Shijiazhuang on July 21, 2017 (He et al., 2021), and the extreme rainstorm over Beijing in late July 2025 (Wang et al., 2026). Mechanistic understanding has been built mainly from South China, where the low-level jet (LLJ), back-building mesoscale convective systems, and frontal–WS interactions have been identified as central controls on convection initiation and rainfall intensity (Wu and Luo, 2016; Du and Chen, 2018, 2019; Yang et al., 2024; Chen et al., 2025). In particular, double low-level jet structures have been shown to shape the regime there: the synoptic-system jet transports moisture, while the boundary-layer jet generates coastal

convergence and triggers convection near its terminus (Du and Chen, 2018, 2019). The nocturnal acceleration of the boundary-layer jet through the inertial oscillation (Blackadar, 1957) further explains the nighttime preference of warm-sector rainfall in that region. Over the eastern China plains, a smaller but growing body of work has characterized WSHR through object-oriented composite analyses in North China (Fu J. et al., 2020) and localized studies over the plains and the Yangtze–Huaihe coastal region (He et al., 2021; Yu et al., 2024).

Building on these advances, three directions deserve further investigation over the eastern China plains. The first is geographic. The mechanistic picture summarized above rests largely on South China, where coastline geometry, sea-breeze fronts, and warm marine moisture set the stage for the LLJ-driven coastal initiation that dominates the regime there. The eastern China plains operate under a different monsoonal and frontal climatology and lack much of that coastal forcing, so the controls established for the coast provide a strong starting point but need not transfer directly. Extending the analysis to this region therefore complements, rather than supplants, the South China work. The second concerns identification. Objective warm-sector criteria have been proposed for various regions of China (Wu et al., 2020; Li et al., 2021). Applying a consistent multi-year objective scheme of this kind to the eastern China plains would allow WSHR to be cleanly separated from other heavy rainfall, and its contribution to the regional total to be quantified on a directly comparable basis. The 200–400 km distance band follows prior studies, while the 99th percentile is chosen here; neither is specifically tuned for this region. The third concerns the pre-storm environment. Existing studies have established the broad synoptic and thermodynamic conditions that favor WSHR over the eastern China plains (Fu J. et al., 2020; Yu et al., 2024). However, the limited time resolution of reanalyses and the sparse sampling of operational radiosondes have led that work to focus on conditions at the time of, rather than in the minutes-to-hours leading up to, convective initiation. Radar wind profilers (RWPs) sample the low-level wind continuously at high vertical and temporal resolution and are well suited to extend this picture into the pre-onset window. Profiler records have

already yielded regional or nationwide LLJ climatologies (Du et al., 2012; Miao et al., 2018; Li et al., 2026), vertical divergence profiles of demonstrated value for nowcasting convective storms (Guo et al., 2023), and evidence that the LLJ modulates the onset of summertime nocturnal rainfall in Beijing (Li et al., 2024). Moreover, the regional RWP network over the eastern China plains has accumulated sufficient data to support such an analysis.

Building on these foundations, this study brings together five warm seasons (2021–2025) of minute-resolution nationwide rain-gauge records, ERA5 reanalysis, and an RWP mesonet over the eastern China plains to pursue three objectives. First, we develop an objective scheme that separates warm-sector from other rainfall and identifies heavy rainfall events of both types, on the basis of automated front detection, a warm-sector mask, and a station-specific hourly intensity threshold. Second, we characterize the spatiotemporal distribution of WSHR, compare its event-scale properties with those of other heavy rainfall, and quantify its contribution to the total heavy-rainfall accumulation. Third, by leveraging minute-scale rain gauge records to pinpoint the onset of rainfall and high-frequency RWP measurements to track the pre-onset environment, we quantify the precursor dynamic signals preceding local WSHR onset. Together, these objectives yield a consistent five-year, station-based picture of WSHR over the eastern China plains, and a set of profiler-observable precursor signals for this poorly predicted rainfall type. The remainder of the paper is organized as follows. Section 2 describes the data and the identification and diagnostic methods. Section 3 presents the spatiotemporal characteristics and the precursor signals. Section 4 discusses methodological strengths and limitations together with the physical mechanisms. Section 5 summarizes the main conclusions.

2. Data and methodology

2.1. Study region and period

The present study focuses on the eastern China plains, where heavy rainfall frequently occurs during the warm season (June–September) and poses substantial risks

to densely populated and economically developed regions. The study period covers the warm seasons of 2021–2025. This period corresponds to the primary season of heavy rainfall over eastern China, during which abundant moisture supply, active synoptic disturbances, and favorable mesoscale forcing frequently contribute to intense hourly rainfall.

Wang et al. (2023) documented a pronounced west-to-east intensification of extreme hourly precipitation across China, with the 95th-percentile hourly intensity attaining its annual high-value centers over the eastern China plains, including the North China Plain. Building on that result, here we examine the warm-season distribution and find that the high-value centers of the 99th-percentile hourly intensity ($R_{P_{99}}$) are likewise concentrated over the low-lying plains of eastern China. Figure 1a shows the spatial distribution of $R_{P_{99}}$ at 1708 national meteorological stations across central and eastern China during the warm seasons of 2021–2025, of which the rain gauges within our study region form a subset. The largest values are confined to the low-lying plains, indicating that these areas carry a high potential for intense hourly rainfall during the warm season. To focus on WSHR over relatively homogeneous low-terrain areas and to reduce the influence of complex terrain, the study region is defined as the portion of the eastern China plains bounded by 111.5–123.5°E and 30.5–40.7°N with terrain elevation at or below 200 m above sea level (dark blue contour in Fig. 1b).

Within this region, rain-gauge observations were used to identify rainfall events and to classify them into warm-sector rainfall and other rainfall categories. In addition, RWP observations at stations co-located with rain gauges were used to investigate the precursor dynamic signals associated with warm-sector rainfall events. As shown in Fig. 1b, the rain-gauge stations provide dense spatial coverage over the study region, while the co-located RWP and rain-gauge stations allow the pre-onset evolution of wind profiles to be aligned with local rainfall onset.

2.2. Rain gauge and RWP measurements

Rain gauge and RWP measurements were obtained from the National

Meteorological Information Center of the China Meteorological Administration (CMA), and strict quality control was applied. Additional data-availability criteria were further applied to ensure the reliability of the datasets.

The rain gauge observations have a temporal resolution of 1 min. To ensure data completeness, only stations with data availability of at least 95% during June–September of 2021–2025 were retained. A total of 493 rain gauges within the study region satisfied this criterion, and their spatial distribution is shown in Fig. 1b. The 1-min rain gauge measurements were accumulated into both hourly and 6-min rainfall totals. At each station, $R_{p_{99}}$ was calculated as the 99th percentile of non-zero hourly rainfall totals pooled over June–September of 2021–2025, with missing observations excluded. This threshold served as the rainfall-intensity criterion for identifying WSHR events. The 6-min totals were used to pinpoint the precise onset time of rainfall and, in conjunction with the RWP observations, to characterize the precursor dynamic signals.

The RWP observations have a temporal resolution of 6 min and a vertical resolution of 120 m. A total of 79 stations with co-located RWP and rain gauge measurements were available within the study region, as shown in Fig. 1b. The quality-control procedure followed the methods of Wei et al. (2014) and Miao et al. (2018). Observations between 0.3 and 5 km above ground level (AGL) were used, corresponding to 39 vertical levels for each profile. Each wind profile was discarded if more than 40% of the data within the 0.3–5 km AGL height range were missing. Missing values in the remaining profiles were then estimated using linear interpolation. The RWP provides the full horizontal and vertical wind, all retained for the analyses that follow; only the ascent diagnostics (Fig. 12) are restricted to ascending velocities ($-w > 0$). In the RWP product used here, w is positive downward. Ascending motion is therefore represented by $-w > 0$. Restricting the ascent diagnostics to negative reported vertical velocities reduces the contribution of downward hydrometeor motion but does not establish that the retained measurements are free of contamination. Hydrometeors may be present aloft before rainfall is detected at the surface gauge.

2.3. Identification of WSHR events

The identification of WSHR events draws on both ERA5 gridded fields and rain gauge observations. The ERA5 data used in this study cover 97–132°E and 22–57°N at a horizontal resolution of 0.25° and a temporal resolution of 1 h over the warm seasons of 2021–2025; this domain is deliberately larger than the study region so that fronts in the vicinity of the gauges can be captured. The 1-min gauge measurements were accumulated to hourly totals to match the temporal resolution of ERA5, yielding a rainfall matrix $R(T_i, S_j)$ of dimension 14640×493 , where T_i denotes the i -th hour ($i = 1, 2, \dots, 14640$) and S_j the j -th station ($j = 1, 2, \dots, 493$).

Synoptic fronts were identified objectively following an approach similar to that of Fu J. et al. (2020). The pseudo-equivalent potential temperature (θ_e) at 850 hPa was first computed from the ERA5 temperature and humidity fields (Bolton, 1980):

$$\theta_e = T_K \left(\frac{1000}{p} \right)^{0.2854(1-0.28 \times 10^{-3} r)} \times \exp \left[\left(\frac{3.376}{T_L} - 0.00254 \right) r (1 + 0.81 \times 10^{-3} r) \right] \quad (1)$$

where T_K is the air temperature at 850 hPa (K), p is the pressure (hPa), r is the water-vapor mixing ratio (g kg^{-1}), converted from the ERA5 specific humidity, and T_L is the lifting condensation temperature (K), obtained following Bolton (1980) as

$$T_L = \frac{1}{\frac{1}{T_K - 55} - \frac{\ln(U/100)}{2840}} + 55 \quad (2)$$

where U is the relative humidity (%). The thermal front parameter (TFP) was then applied to detect fronts and to retrieve their synoptic attributes, including frontal type (cold, warm, or quasi-stationary), frontal length, and frontal position; these attributes are hereafter referred to as the synoptic front features, $SFF[T_i]$ ($i = 1, 2, \dots, 14640$) (see Section 1 in the Supplement for the full procedure).

In accordance with the criteria of Zhang et al. (2022) and Song et al. (2023), the warm sector is defined as the region where warm advection prevails at both 925 and 1000 hPa and the meridional wind component (v) is southerly ($v > 0$); the set of grid points satisfying this definition at hour T_i constitutes the warm-sector mask,

WSM[T_i].

The warm-sector rainfall criteria in our study follow the conceptual definition of earlier work, in which WSHR develops within the warm, moist air mass on the warm side of, and well separated from, a surface front under weak synoptic-scale forcing (Huang et al., 1986; Chen et al., 2025). The 200–400 km distance band used below is likewise guided by previous studies: its lower bound keeps the samples clear of the frontal zone (Yang et al., 2024), and its upper bound confines them to the typical reach of the warm-sector ahead of the front (Chen et al., 2025). A rainy hour was classified as warm-sector rainfall only if the station simultaneously met both of the following:

- (i) it lay on the southern side of the front, between 200 and 400 km from it; and
- (ii) it fell within the warm-sector mask, $WSM[T_i]$.

Otherwise, it was classified as other rainfall. These two groups correspond to the warm- and cool-toned cells in Fig. 2, respectively, and each was further split by intensity into heavy ($R \geq R_{P_{99}}$; darker cells) and non-heavy ($0 < R < R_{P_{99}}$; lighter cells) rainfall, yielding the four labels. The rainfall label matrix $L(T_i, S_j)$, of the same dimension as $R(T_i, S_j)$, was initialized to N/A, and labels were assigned only to the station-hours with measurable rainfall ($R > 0$); all remaining entries were left as N/A. Accordingly, L is defined as

$$L(T_i, S_j) = \begin{cases} \text{N/A,} & \text{no rainfall} \\ 1, & \text{WSHR} \\ 2, & \text{warm-sector non-heavy rainfall} \\ 3, & \text{other heavy rainfall} \\ 4, & \text{other non-heavy rainfall} \end{cases} \quad (3)$$

Considering the intermittency of rainfall, the labeled hourly records were grouped into events following Zhang et al. (2023), with successive events separated by a dry period of at least 3 h. An event that reached $R_{P_{99}}$ in at least one hour was treated as a heavy rainfall event, and was designated a WSHR event or other heavy rainfall event depending on whether all of the hours from the onset of rainfall to the first hour reaching or exceeding the threshold lay within the warm sector (a WSHR case is illustrated in Fig. 3d). An event never reaching $R_{P_{99}}$ was classified from the station location at

onset, as a warm-sector non-heavy rainfall event or other non-heavy rainfall event according to whether the onset lay within the warm sector. Finally, any event influenced by a tropical cyclone (TC), defined as one with rainfall occurring within 500 km of a TC center according to the TC best-track dataset, was assigned to the other heavy or other non-heavy rainfall event category (depending on its intensity), even if it would otherwise have satisfied the warm-sector rainfall criteria. This treatment keeps the WSHR sample free of TC-related circulations while leaving the event totals complete.

2.4. Event compositing and RWP-based precursor diagnostics

The precursor dynamic signals were investigated at the stations where an RWP and a rain gauge are co-located. To match the 6-min resolution of the RWP, each event's onset time was determined from the 6-min accumulated rainfall and rounded to the nearest 6-min mark (Li et al., 2024). Here, onset refers to the first detected rainfall at the co-located gauge and does not necessarily coincide with convective initiation, because convection may develop upstream before reaching the station. The WSHR events occurring at these co-located stations were then used for the precursor analysis. A comparison sample consisted of 271 warm-sector non-heavy rainfall (WS non-HR) events from the same 37 co-located stations that contributed the 62 WSHR events. This sample was used to assess which features of the pre-onset kinematic evolution distinguish WSHR from non-heavy rainfall within the warm sector. WS non-HR events were defined as rainfall events that satisfied the warm-sector criteria at onset but never reached the station-specific heavy-rainfall threshold. The two samples were processed using the same RWP quality-control criteria, onset alignment and 180-min analysis window.

To characterize the environment in which WSHR develops, we additionally constructed station-relative composites of the ERA5 fields centered on each WSHR event. The target station, the rain-gauge station where each WSHR event occurred, was placed at the origin of a relative grid spanning $\pm 3.5^\circ$ in longitude and latitude. The 850-hPa specific humidity and θ_e , together with the low-level (1000–700 hPa) vertically

integrated moisture flux divergence (VIMFD), were averaged over all WSHR events at 6 h and 3 h before onset and at the onset hour. The ERA5 composites use the full sample of 477 station-based WSHR events, whereas the RWP composites use the subset with available profiler observations. The two datasets therefore characterize complementary aspects of the pre-onset environment and do not provide a fully matched thermodynamic–kinematic composite. Following the moisture-budget form of the water-vapor conservation equation (Banacos and Schultz, 2005) and the diagnostic procedure for reanalysis fields of Seager and Henderson (2013), VIMFD was obtained by first integrating the horizontal moisture flux over the 1000–700 hPa layer to form the vertically integrated moisture flux vector $Q = (Q_\lambda, Q_\varphi)$,

$$Q_\lambda = \frac{1}{g} \int_{p_t}^{p_s} qu \, dp, \quad Q_\varphi = \frac{1}{g} \int_{p_t}^{p_s} qv \, dp \quad (4)$$

and then taking its horizontal divergence on the sphere,

$$VIMFD = \nabla \cdot Q = \frac{1}{a \cos \varphi} \frac{\partial Q_\lambda}{\partial \lambda} + \frac{1}{a \cos \varphi} \frac{\partial (Q_\varphi \cos \varphi)}{\partial \varphi} \quad (5)$$

where g is the gravitational acceleration, q the specific humidity, u and v the zonal and meridional wind components, p the pressure with $p_s = 1000$ hPa and $p_t = 700$ hPa the base and top of the integration layer, a the Earth's radius, and λ and φ the longitude and latitude. By convention, positive VIMFD denotes moisture flux divergence and negative values denote convergence, so that the strengthening low-level moisture convergence preceding WSHR appears as increasingly negative VIMFD. These composites place the RWP-based kinematic diagnostics described below within their large-scale moisture and energy context.

For the WSHR events, the RWP signals were composited over the 180 min preceding the onset of rainfall. At each 6-min step and height level, we computed the mean wind profile, mean wind kinetic energy, mean vertical wind shear, the occurrence proportion and frequency of low-level jets (LLJs), the frequency of ascending motion, and the maximum ascent velocity. Tracking these signals through the pre-onset window characterizes how the low-level dynamic environment evolves as WSHR approaches onset.

Following the method proposed by Fu S. et al. (2020), the wind kinetic energy per unit mass is derived from the following equation:

$$KE = \frac{u^2 + v^2}{2} \quad (6)$$

where u and v are the zonal and meridional winds (m s^{-1}) and KE is in J kg^{-1} . The same KE diagnostic was calculated for both WSHR and WS non-HR events, with the corresponding time–height composites shown in Fig. 9c and Fig. S6, respectively. Following previous convective-environment and wind-profiler studies (Markowski and Richardson, 2006; Thompson et al., 2007; Li et al., 2013; Zhang et al., 2020), the vertical wind shear (VWS) over a layer is taken as the magnitude of the bulk vector wind difference (top minus bottom) divided by the layer depth:

$$VWS = \frac{\sqrt{(u_t - u_b)^2 + (v_t - v_b)^2}}{z_t - z_b}, \quad z_t > z_b \quad (7)$$

where z_t and z_b are the top and bottom heights of the layer, respectively, expressed in meters. The top-minus-bottom convention defines the orientation of the bulk-shear vector; reversing the order would reverse the vector but would not change its magnitude. The scalar VWS is therefore non-negative and has units of s^{-1} . For Fig. 10, 10^3 VWS is plotted, and the corresponding values are reported in 10^{-3} s^{-1} . To retain directional information, composite hodographs are also presented for WSHR and WS non-HR events in Figs. S8 and S9. The arrows represent the bulk wind difference $\Delta \mathbf{V} = \mathbf{V}_t - \mathbf{V}_b$, in m s^{-1} , whereas Eq. (7) gives its magnitude divided by the layer depth, in s^{-1} . The layer-endpoint heights are displayed in km AGL on the figure axes. The LLJ identification criteria are given in Section 2 in the Supplement. For the box plots in Fig. 11a, boxes span the 25th–75th percentiles. Whiskers extend to the most extreme observations within 1.5 times the interquartile range below the first quartile and above the third quartile. Observations beyond the whiskers are not shown.

3. Results

3.1. Spatiotemporal characteristics of WSHR

A total of 477 WSHR events were identified within the study region during the

warm seasons of 2021–2025. Their spatial distribution is markedly sparse and scattered (Fig. 4a). Of the 493 stations, 196 recorded no WSHR event and 297 recorded at least one. Among the latter, 180, 70, 32, 14 and 1 stations recorded one, two, three, four and five events, respectively (Fig. S4a). The higher counts show a weak preference for the southern and, in particular, the southeastern parts of the region, but they do not organize into a coherent spatial gradient. No clear relationship with local terrain elevation is apparent within the selected low-elevation domain. This lack of terrain dependence is unsurprising given the deliberately low and homogeneous relief (≤ 200 m) of the study region. This spatially scattered distribution is consistent with the weakly forced, locally triggered character of warm-sector convection, in which the absence of strong synoptic lifting leaves convective initiation to be governed by mesoscale and boundary-layer processes.

Temporally, WSHR is concentrated in midsummer, with the great majority of events occurring in July and August and very few in June or September (Fig. 4b). Its diurnal distribution is bimodal: a weaker afternoon mode around 1300–1600 Beijing Time (BJT) is accompanied by a dominant nocturnal-to-postmidnight mode spanning roughly 2000–0300 BJT, the latter containing the highest event counts of the entire distribution.

For comparison, 6002 other heavy rainfall events were identified over the study period. Their spatial distribution is both ubiquitous and well organized (Fig. 5a). Every station in the study region recorded at least five events. Event counts increase from north to south, with the maxima concentrated in the south and especially the southeastern corner of the region, decreasing northward to 6–15. Diurnally, other heavy rainfall displays a strong, sharply defined late-afternoon maximum around 1400–1800 BJT, most pronounced in July and characteristic of thermally driven convection (Fig. 5b); a secondary nocturnal signal (about 2200–0200 BJT) also emerges in July, but the afternoon peak clearly dominates.

The two categories therefore differ not only in frequency but also in spatial organization and diurnal phase. Both share the dual (afternoon and nocturnal) diurnal

character previously documented for the region between the Yellow and Yangtze rivers, where short-duration rainfall tends to peak in the late afternoon while longer-lived, nocturnal episodes constitute a second maximum (Yu et al., 2007a, b; He and Zhang, 2010). However, their relative emphasis is reversed: other heavy rainfall is afternoon-dominated, pointing to surface heating and the associated growth of conditional instability, whereas WSHR is comparatively more nocturnal. This nocturnal preference, together with the scattered and weakly organized spatial pattern, points to a forcing distinct from daytime thermal convection. A natural candidate is the nocturnal acceleration of the low-level jet through the inertial oscillation, which after sunset strengthens the warm, moist southerly flow and the convergence and conditional instability at its terminus, and which has been widely linked to warm-sector and nocturnal rainfall (Du and Chen, 2018, 2019). This mechanism is a plausible interpretation based on previous studies; the present diurnal statistics do not directly test inertial oscillations or jet-exit convergence. In this respect the eastern China plains echo South China, where warm-sector rainfall peaks at night to early morning under nocturnal low-level-jet forcing, in contrast to the afternoon peak of frontal rainfall. These contrasts motivate the analysis of the precursor dynamic environment presented in the following sections, which examines how the low-level environment evolves in the hours preceding the onset of WSHR events.

3.2. Comparison between WSHR and other heavy rainfall events

Summed over all stations in the study region, WSHR events accounted for roughly 5–10% of all heavy rainfall events in each year of the study period (Fig. 6a). Against annual totals of 1131–1541 events, WSHR events numbered 72–125 per year, with the remaining 1019–1416 classified as other heavy rainfall. The two categories differ systematically in their event-scale characteristics, as detailed below.

The probability density function (PDF) of event duration (Fig. 6b) is shifted toward shorter durations for WSHR events: their median duration is 5.7 h, compared with 6.8 h for other heavy rainfall, and their 10th–90th percentile range is

correspondingly narrower (2.1–13.4 h versus 2.5–17.6 h). The distribution of WSHR events also peaks higher and earlier, indicating a stronger concentration at short durations. The PDF of event cumulative rainfall (Fig. 6c) shows a complementary pattern. Although the median accumulations are comparable (70.3 mm for WSHR versus 56.2 mm for other heavy rainfall), the distribution of WSHR events is broader and possesses a markedly heavier upper tail, with a 90th percentile of 156.1 mm against 120.2 mm for other heavy rainfall. The other-category PDF, by contrast, is more sharply peaked at moderate amounts. Overall, the WSHR sample has a duration distribution shifted toward shorter events and a rainfall-accumulation distribution with a comparable median and a heavier upper tail than the other-heavy-rainfall sample. This combination of brief duration and large accumulation is precisely the behavior that makes WSHR hazardous and difficult to anticipate, consistent with the localized, sudden, and poorly predictable nature of warm-sector convection reported in previous studies (Du and Chen, 2018; Fu J. et al., 2020).

The accumulated rainfall at each station from WSHR events (Fig. 6d) exhibits a clear spatial structure that echoes the pattern of event frequency in Fig. 4a. Accumulations increase toward the south and, most prominently, the southeastern corner of the region, where many stations exceed 300–500 mm, whereas the northern stations generally remain below about 150 mm. To gauge the importance of this contribution, we computed at each station the ratio of WSHR to total (WSHR plus other heavy rainfall) heavy rainfall accumulation (Fig. 6e). Among the 297 stations with at least one WSHR event, the mean station-level contribution is 13.8% (median 12.4%), with a maximum of 48.1% (Fig. 6e). This statistic describes the average contribution at WSHR-affected stations, rather than the fraction obtained by pooling rainfall across the entire region. Although WSHR events are much less frequent than other heavy rainfall events, their contribution can be substantial at individual stations.

3.3. Pre-onset synoptic environment of WSHR

The station-relative composites introduced in Section 2.4 characterize the moisture

and energy conditions in which WSHR develops. Averaged over all WSHR events, the 850-hPa specific humidity and θ_e fields display a consistent thermodynamic setting from 6 h before onset through the onset hour (Fig. 7), described below at the onset hour (Fig. 7c, f).

The composite specific humidity field at onset (Fig. 7c) places the event station within a moist tongue. Specific humidity at the station reaches 14.8 g kg^{-1} , close to the composite maximum; the moist core extends toward the southwest, and values decrease sharply to about 10 g kg^{-1} roughly 3° to the north. The station therefore lies near the northern, leading edge of the moist tongue, immediately upstream of the strong meridional moisture gradient that separates the warm, humid air from the drier air to the north. The composite θ_e field (Fig. 7f) reinforces this picture energetically. The station sits on the warm side of a pronounced θ_e gradient, where θ_e is about 350–352 K and decreases northward to roughly 336–340 K across a band of densely packed contours; this concentrated θ_e gradient marks the frontal zone, with the front lying to the north of the station. WSHR is therefore consistently found within the warm, humid, high- θ_e tongue on the warm (southern) flank of the front. Such a moist, high- θ_e environment is consistent with the background conditions reported for WSHR over South China. In that region, moist low-level jets are associated with heavy rainfall and can support convective development through enhanced moistening and lifting (Du and Chen, 2018, 2019).

The composite VIMFD field adds a dynamic dimension to this thermodynamic setting (Fig. 8). Low-level (1000–700 hPa) moisture flux convergence, shown by negative VIMFD, is present near the event station throughout the pre-onset window and intensifies as onset approaches. Six hours before the onset of rainfall, the convergence is modest and displaced to the west of the station (Fig. 8a), and it deepens by three hours before onset (Fig. 8b). At the onset of rainfall, a pronounced convergence core sits essentially on the station, with values below about $-5 \times 10^{-4} \text{ kg m}^{-2} \text{ s}^{-1}$ (Fig. 8c). This progressive concentration of low-level moisture convergence at the station, superimposed on the warm, humid, high- θ_e tongue, preconditions the column for the

deep moist convection that produces WSHR. This evolution documents increasing low-level moisture flux convergence near the station as rainfall onset approaches, without establishing the process responsible for convective initiation.

These composite conditions establish that WSHR develops in an environment that is favorable in terms of moisture, energy, and low-level moisture convergence. These composites describe the environment associated with the sampled WSHR events but do not establish necessary or sufficient conditions for convective initiation. Under the weak synoptic-scale forcing that characterizes the warm sector, mesoscale and boundary-layer dynamic processes are likely to play an important, and perhaps decisive, role in the triggering and intensification of convection. To characterize the associated kinematic evolution, the following section examines the precursor dynamic signals recorded by the radar wind profilers co-located with WSHR stations (Fig. 9a), focusing on how the low-level dynamic environment evolves in the hours preceding WSHR onset.

3.4. RWP-derived precursor dynamic signals for WSHR

Following the procedure described in Section 2.4, 62 station-based WSHR events at 37 co-located stations were retained for the precursor analysis. Most stations contributed a single event (Fig. S4b). Using spatial and onset-time difference thresholds of 500 km and 24 h, respectively, these events were assigned to 47 episode clusters (Fig. S4d). These clusters are operational groupings of potentially related events; their statistical independence is not established. RWP profiles were aligned with each event's local rainfall onset and composited over the preceding 180 min.

In the mean wind profile (Fig. 9b), wind speed increases almost monotonically with height, from roughly 4 m s^{-1} near the surface to about 12 m s^{-1} at 5 km AGL, and the wind veers clockwise throughout the layer. This veering is a signature of low-level warm-air advection consistent with the warm, moist southerly flow and the high- θ_e tongue documented in Section 3.3. The composite hodograph also shows clockwise turning between the lowest and highest sampled levels, with a turning angle of about

80° for the preceding 3-h mean (Fig. S8a). Clockwise turning is also present in the WS non-HR composite (Fig. S9). The green bars show the mean 6-min rainfall accumulation during the first 30 min after local rainfall onset.

The wind kinetic energy generally increases with height in both the WSHR and WS non-HR composites (Fig. 9c and Fig. S6). The WSHR composite shows a more pronounced enhancement at approximately 1–3 km AGL during the final hour before local rainfall onset, whereas the corresponding evolution in WS non-HR is comparatively gradual. However, these differences in composite means do not establish that the KE evolution is unique to WSHR or that it provides event-level predictive skill.

The composite vertical wind shear (Fig. 10) is dominated by a shallow but intense layer of low-level shear. The strongest vertical wind shear (up to about $14 \times 10^{-3} \text{ s}^{-1}$) is concentrated in bulk layers confined to the lowest 1.5 km or so (lower left of each panel), while deep-layer shear spanning the mid-troposphere is comparatively weak. This low-level vertical wind shear intensifies as the onset of rainfall approaches. Its prior-3-h composite mean is shown in Fig. 10a, while the time windows in Fig. 10b–f resolve the evolution. The vertical wind shear is relatively steady in the earlier windows (Fig. 10b, c), strengthens over the final hour (Fig. 10d, e), and, at the onset of rainfall, is joined by an additional band of strong vertical wind shear near 2–3 km (Fig. 10f). This deepening and intensification of the shear indicate an evolving low-level wind structure but do not directly establish horizontal convergence or its role in organizing convection. Strong shallow-layer shear is also present in the WS non-HR composites (Fig. S7). The WSHR composites show a more pronounced enhancement near 2–3 km close to onset, whereas the WS non-HR evolution is comparatively steady. Thus, the presence of strong low-level shear alone does not distinguish the two rainfall categories.

Because the horizontal-wind precursor signals are concentrated below 3 km, the low-level wind was analyzed in greater detail (Fig. 11). The maximum wind speed below 3 km has a median of about 12 m s^{-1} that rises gradually toward onset, with upper whiskers reaching about 25 m s^{-1} in the final tens of minutes (Fig. 11a). An LLJ is identified at least once during the analysis window in 29 of the 62 WSHR events

(46.8%) and 107 of the 271 WS non-HR events (39.5%; Fig. 11b, c). The event-mean jet-core heights and speeds overlap substantially between the two groups, indicating that a low and strong jet core is not unique to WSHR. The LLJ occurrence frequency in WSHR is generally about 5–11% at earlier lead times, increases during the final hour, peaks at 21.0% at –36 min, and is 16.1% at rainfall onset (Fig. 11d). In comparison, the WS non-HR frequency remains relatively steady at approximately 9–13%. For WSHR, the event-mean LLJ occurrence proportion increases by 7.0 percentage points from the early window (–180 to –120 min) to the late window (–60 to 0 min). Both windows include their endpoints. The 95% episode-cluster bootstrap confidence interval is –0.25 to 14.18 percentage points (5000 resamples; Section 3 in the Supplement). Because this interval includes zero, the increase should be interpreted as a composite mean tendency with appreciable uncertainty. Its possible connection to convergence and convective triggering is a literature-based interpretation discussed in Section 4.2, rather than a causal relationship established by these profiles.

The strengthening low-level flow is accompanied by enhanced ascent. Throughout, only ascending velocities ($-w > 0$) are considered. Figure 12a shows, at each height and lead time, the frequency with which the events exhibit ascent (the fraction with $-w > 0$). In the WSHR composite, ascending motion is concentrated in the 1–3 km layer, with the highest frequencies, up to about 35%, centered near 1–2 km. Ascending motion occurs frequently in the lower troposphere in both samples (Fig. 12a, b). In WSHR, enhanced frequencies are evident near 1–2 km during part of the final hour, but the evolution is not monotonic and the frequencies decrease close to onset. The WS non-HR sample also exhibits frequent low-level ascent, indicating that its occurrence alone is not specific to WSHR.

Figure 12c–f summarizes the distribution across events of the maximum upward velocity within each height layer at each lead time. The clearest late enhancement in WSHR occurs at 1–2 km (Fig. 12e), where the mean and upper part of the distribution increase during approximately the final 30 min. The WS non-HR distribution changes less markedly, although the two samples overlap. Enhanced upper-tail ascent is also

evident at 0.3–1 km (Fig. 12f), while the evolution above 3 km is comparatively weak (Fig. 12c). These distributions indicate substantial event-to-event variability and do not establish a causal relationship between the LLJ and ascent.

The comparison with WS non-HR events indicates that low-level shear, clockwise wind turning and ascending motion occur in both rainfall categories. The more pronounced late increase in LLJ occurrence and upper-tail ascent at 1–2 km in WSHR provides partial evidence of differing pre-onset evolution. However, substantial overlap between the samples remains, and neither a causal jet–ascent pathway nor event-level predictive skill is established by these composites.

4. Discussion

4.1. Strengths and limitations of the identification and diagnostic framework

The objective scheme developed here separates WSHR from other heavy rainfall without manual synoptic typing, which is its main advance over the case-by-case or subjectively classified samples that dominate the warm-sector literature. Combining automated TFP detection (Hewson, 1998; Jenkner et al., 2010; Fu J. et al., 2020) with a warm-sector mask, a fixed 200–400 km distance band, and a locally defined 99th-percentile intensity threshold, the method yields a homogeneous five-year sample of 477 WSHR and 6002 other heavy rainfall events across 493 stations. This consistency makes the comparisons in Sections 3.1 and 3.2 reproducible and removes the subjectivity that complicates earlier intercomparisons.

Several limitations temper this strength. The front and warm-sector fields come from ERA5 at 0.25° and hourly resolution (Hersbach et al., 2020), which represents synoptic fronts well but smooths drylines, outflow boundaries, and other mesoscale features, so a minority of events near them may be misattributed. TFP detection places the front within the baroclinic zone rather than at its warm leading edge (Hewson, 1998), and the distance band and percentile thresholds, though physically motivated, were not tuned by sensitivity tests. Defining heavy relative to each station's climatology accommodates the south-to-north moisture gradient but mixes stations of very different

absolute intensity. These choices warrant sensitivity tests in future work.

The RWP diagnostics are a second contribution. Continuous 6-min wind profiling captures pre-onset low-level flow evolution that hourly reanalyses and soundings fail to resolve (Du et al., 2012; Guo et al., 2023), allowing us to characterize the pre-onset dynamic environment of WSHR within the broader warm-sector thermodynamic environment characterized by ERA5. The approach is nonetheless constrained by the single-point nature of each profiler. Vertical motion is inferred indirectly rather than measured as a divergence field, which would require profiler mesonets (Guo et al., 2023); the 0.3–5 km range omits the surface layer and the mid to upper troposphere. The hourly ERA5 fields cannot resolve thermodynamic changes on the same timescale as the 6-min RWP observations, particularly during the final 30 min before rainfall onset. The timing of moisture and stability changes relative to the observed kinematic intensification therefore remains unresolved. Finally, compositing 62 events smooths the large case-to-case variability seen in Figs. 11 and 12, so the composite signals are best interpreted as mean tendencies across the sampled events, rather than event-level certainties. The WS non-HR comparison provides a baseline within the warm sector, but non-raining warm-sector conditions and other heavy rainfall remain outside the present comparison. Event-level predictive skill therefore remains untested. Equal-episode composites retain the broad kinematic structures of the station-event composites (Fig. S5), while event and episode-cluster bootstrap intervals for LLJ frequency are generally similar under the selected grouping (Fig. S10). These checks assess sensitivity to event weighting and within-episode dependence; they do not establish independence between all episode clusters.

4.2. Physical interpretation and possible mechanisms

The results in Section 3 provide an observational basis for discussing possible mechanisms linking the warm-sector environment to heavy rainfall. The thermodynamic stage is set by the warm, humid, high- θ_e tongue on the warm side of the front (Fig. 7), which characterizes the moist, high- θ_e low-level background of the

sampled events. Such a thermodynamic background is favorable but, under weak synoptic-scale forcing, is not by itself sufficient to initiate WSHR. The strengthening low-level moisture flux convergence (Fig. 8) and the RWP kinematic evolution motivate consideration of mesoscale and boundary-layer processes in this transition.

The RWP composites show increasing LLJ occurrence during the final hour before rainfall onset (Fig. 11), together with enhanced low-level kinetic energy (Fig. 9) and ascent at 1–2 km (Fig. 12). Previous studies suggest that boundary-layer jets can promote convergence near their termini and facilitate convective initiation (Du and Chen, 2018, 2019). This provides a possible interpretation of the observed evolution, but the present analysis does not locate the jet terminus, resolve the associated horizontal convergence, or establish a causal link between the LLJ and ascent. Nocturnal inertial acceleration is another possible contribution (Blackadar, 1957), which is not tested here.

The shallow but intense low-level shear (Fig. 10) acquires significance in this context. Low-level (0–3 km) vertical wind shear is a primary control on the organization and longevity of moist convection (Weisman and Klemp, 1982; Rotunno et al., 1988). The observed shear evolution may be related to the changing low-level flow, including the LLJ. Its potential role in organizing or sustaining convection remains a literature-based hypothesis because convective organization is not resolved in the present analysis.

This possible pathway may help explain why WSHR events have shorter durations yet comparable median and heavier extreme accumulations. Following the ingredients framework of Doswell et al. (1996), heavy rainfall reflects sustained high rainfall rates, the product of strong ascent acting on moist, nearly saturated air at high precipitation efficiency. The enhanced upper tail of ascent speeds near 1–2 km in the WSHR composite, together with the moist, high- θ_e environment, is consistent with conditions favorable for high rainfall rates. However, these composites do not establish that the observed ascent caused the heavy rainfall. If the resulting convection is quasi-stationary or back-building in the manner of extreme-rain-producing mesoscale convective

systems elsewhere (Schumacher and Johnson, 2005), large totals can accumulate over a short interval. The present single-site analysis cannot resolve this organizational mode directly, so the inference remains a hypothesis to be tested with radar.

The principal observational result is the resolved pre-onset kinematic evolution of WSHR and its comparison with warm-sector non-heavy rainfall. The observed composite features include a buildup of low-level kinetic energy, low and strong LLJ cores in a subset of events, a late mean increase in LLJ occurrence frequency with substantial uncertainty, and enhanced upper-tail ascent speeds near 1–2 km shortly before onset. Together, they characterize the pre-onset evolution of WSHR. These low-level dynamic signals are not available from moisture and instability fields alone, and they are consistent with recent profiler-network evidence that low-level jets strengthen within roughly 30–120 min before nocturnal rainfall (Li et al., 2024).

4.3. Implications and directions for future work

These findings bear directly on the long-standing difficulty of forecasting WSHR, for which the quantitative precipitation forecast skill of operational global and regional models remains low (Luo et al., 2017). Because the kinematic features identified here can be monitored by operational radar wind profilers, they warrant evaluation as candidate short-range indicators. In particular, the late increase in LLJ occurrence and enhanced low-level ascent motivate further examination, although their specificity, useful lead time and false-alarm characteristics remain to be established through event-level validation. Guo et al. (2023) demonstrated that RWP mesonets can resolve lower-tropospheric convergence preceding heavy rainfall, supporting the value of high-frequency wind observations for nowcasting. A complementary application is data assimilation: using observing system simulation experiments for three heavy-rainfall cases over the Beijing–Tianjin–Hebei region, Zhao et al. (2025) showed that adding suitably located RWP observations can improve mesoscale wind analyses and short-term rainfall forecasts. Together, these studies motivate combining profiler-based monitoring with assimilation experiments to evaluate potential improvements in

WSHR prediction.

The statistical, composite nature of the analysis also defines its principal avenue for extension. Composites establish the mean precursor structure but average over the diversity of individual events, and they cannot confirm the causal chain or the convective organization inferred above. Detailed case studies of individual WSHR events are the necessary next step, combining the RWP records with weather-radar reflectivity and dual-polarization fields to test whether back-building or training organization underlies the large accumulations (Schumacher and Johnson, 2005; Wu and Luo, 2016). Convection-permitting simulations could then verify the jet–convergence–ascent pathway and isolate the contributions of the nocturnal inertial oscillation, cold pools, and terrain. Pairing profilers with ground-based microwave radiometers for thermodynamic profiling, and deploying profiler mesonets for direct horizontal-divergence retrieval (Guo et al., 2023), would remove two of the present observational limitations. Finally, applying the identification framework to other monsoon regions and seasons would test the generality of the precursor signals reported here.

5. Conclusions

Using five warm seasons (2021–2025) of minute-resolution rain-gauge records, ERA5 fields, and radar wind profiler (RWP) observations, we developed an objective scheme to separate warm-sector heavy rainfall (WSHR) from heavy rainfall not satisfying the warm-sector criteria over the eastern China plains. We characterized the two types and identified the pre-onset dynamic signals that precede WSHR.

WSHR is comparatively infrequent (477 station-based events, about 5–10% of the annual heavy-rainfall events, against 6002 other events) and weakly organized in space, scattered across the region without a clear terrain or latitudinal gradient, in contrast to the south-to-north decrease of other heavy rainfall. Its onset is preferentially nocturnal, with a 2000–0300 BJT maximum and a weaker afternoon mode, whereas other heavy rainfall peaks in the late afternoon. Their shorter durations and larger upper-tail

677 accumulations indicate a concentration of substantial rainfall into relatively short
678 events. Among stations recording at least one WSHR event, the mean station-level
679 contribution to heavy-rainfall accumulation is 13.8%, with a maximum of 48.1%.

680 Composite RWP observations indicate that, within the warm, humid, high- θ_e
681 environment where WSHR develops, the pre-onset low-level flow undergoes a marked
682 kinematic evolution. This includes enhanced kinetic energy below 3 km, a late mean
683 increase in LLJ occurrence and low, strong jet cores in a subset of events, and
684 intensified shallow vertical shear. Enhanced upper-tail ascending motion near 1–2 km
685 also occurs before onset, against a background of strengthening low-level moisture flux
686 convergence. The late increase in LLJ occurrence and upper-tail ascent is more
687 pronounced in WSHR than in WS non-HR, although the distributions overlap. These
688 composite tendencies describe the mean evolution across the sampled events and may
689 not occur in every individual event. These precursor dynamic signals are compatible
690 with a possible jet-related lifting pathway, but the causal links remain unverified. They
691 characterize the pre-onset evolution of WSHR and may serve as candidate reference
692 indicators, with their specificity and event-level predictive skill to be evaluated in future
693 work.

694 **Data availability**

696 The processed data derived from rain-gauge observations, wind profiler radar
697 measurements, and ERA5 reanalysis used in this study can be acquired from Deng and
698 Guo (2026). The ERA5 dataset can be acquired via Copernicus Climate Change Service
699 (C3S) CDS (Hersbach et al., 2023). The TC best-track dataset can be downloaded freely
700 from <https://tcdata.typhoon.org.cn/en/index.html>.

701 **Author contribution**

703 **Weilong Deng:** Conceptualization, Methodology, Visualization, Writing (original
704 draft preparation), Writing (review and editing). **Jianping Guo:** Conceptualization,
705 Data curation, Funding acquisition, Validation, Writing (review and editing). **Ning Li:**

Investigation, Methodology, Validation, Writing (review and editing). **Yi Yang**: Investigation, Methodology, Validation, Writing (review and editing). **Zhen Zhang**: Investigation, Validation. **Yifei Wang**: Methodology, Validation, Writing (review and editing).

Competing interests

The contact author has declared that none of the authors has any competing interests.

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Figures

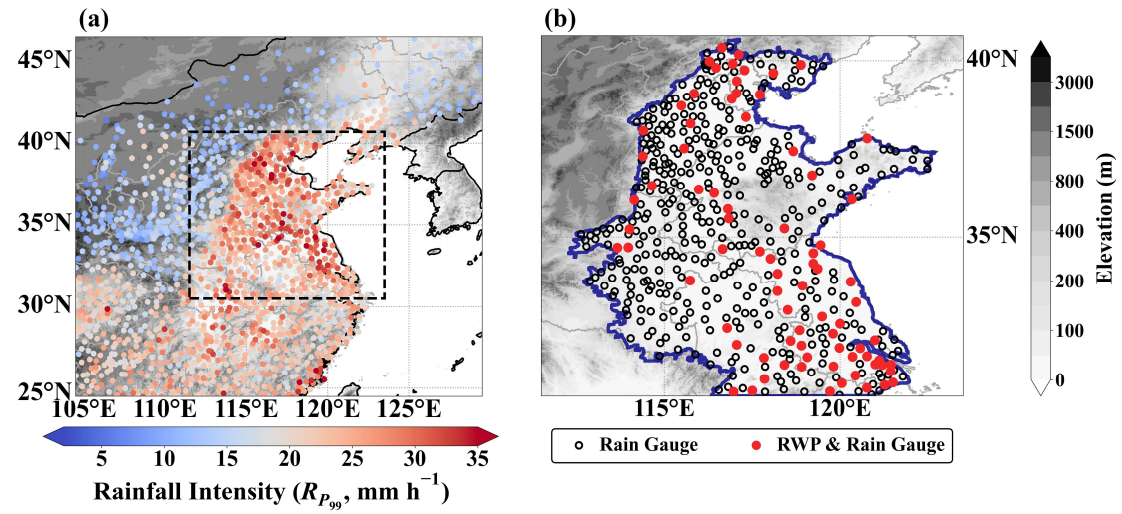


Figure 1. (a) Spatial distribution of 99th-percentile hourly rainfall intensity (R_{P99} ; mm h^{-1}) at rain gauges across eastern China during the warm seasons (June–September) of 2021–2025. Grayscale shading represents terrain elevation, and the dashed black rectangle marks the subdomain zoomed in panel (b). (b) Enlarged view of the study domain covering the eastern China plains. The dark blue contour denotes areas within 111.5–123.5°E and 30.5–40.7°N with terrain elevation at or below 200 m. Open black circles indicate standalone rain gauge sites, while red filled circles mark co-located radar wind profiler (RWP) and rain-gauge stations.

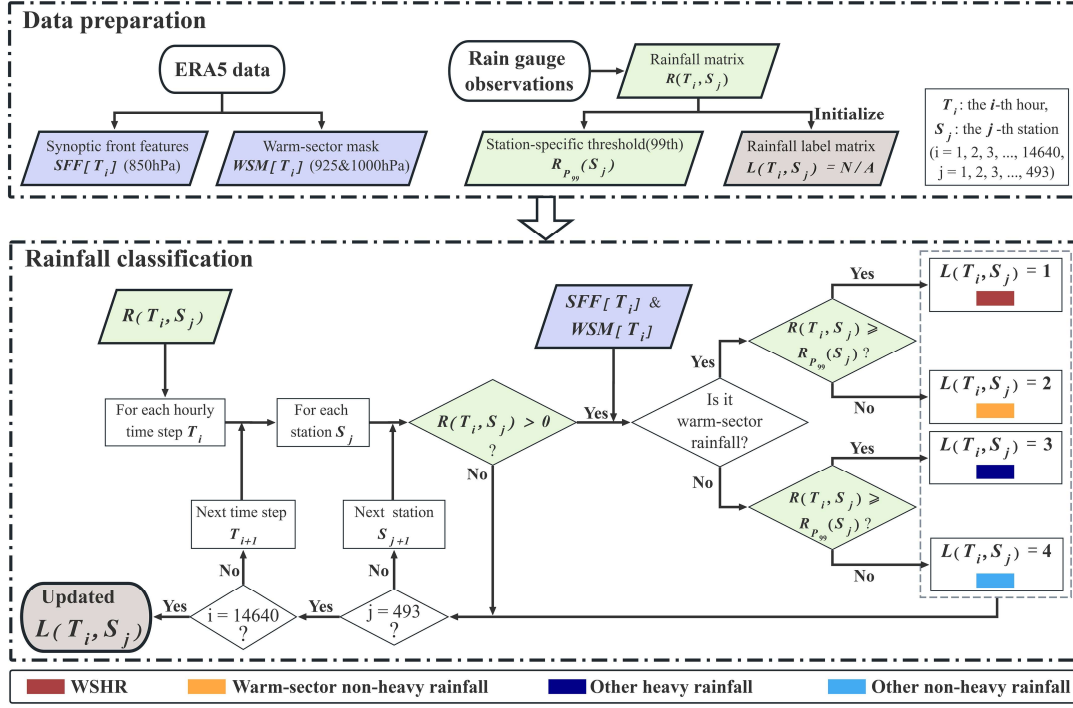


Figure 2. Flowchart illustrating the objective classification scheme for hourly rainfall observations at rain gauge stations. Four rainfall categories are defined: warm-sector heavy rainfall (WSHR, $L = 1$, dark red); warm-sector non-heavy rainfall ($L = 2$, orange); other heavy rainfall ($L = 3$, dark blue); and other non-heavy rainfall ($L = 4$, light blue).

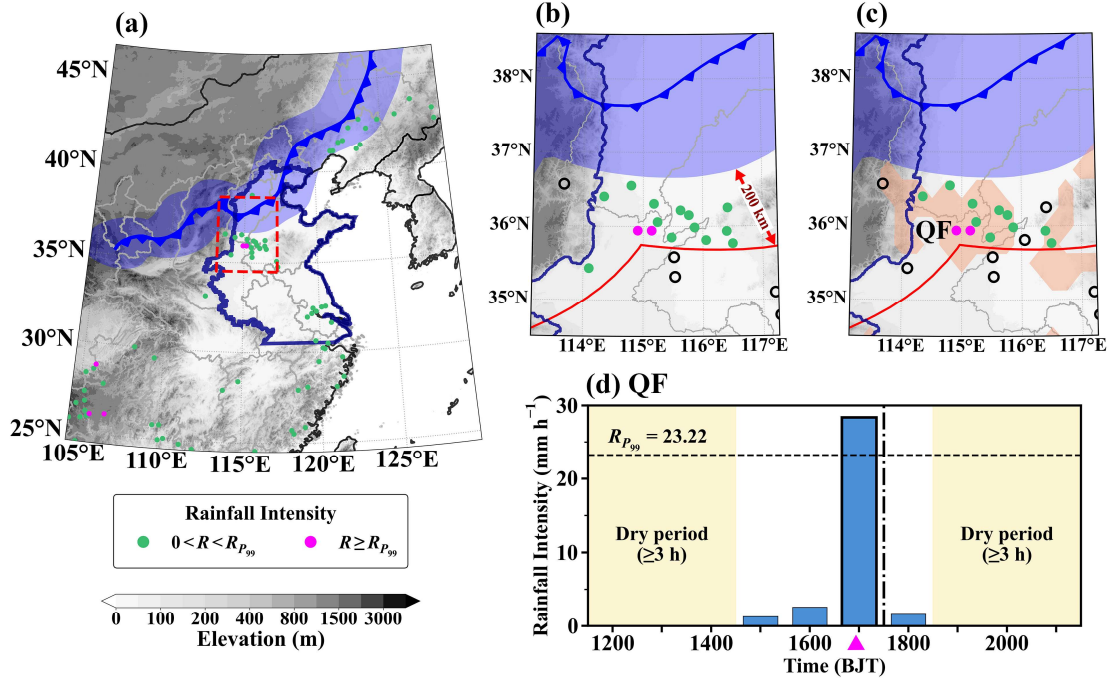


Figure 3. Example of an identified WSHR event at Qingfeng (QF) weather station on August 18, 2023. (a) Cold front (blue line) and its 200-km frontal zone (blue shading) at 1700 Beijing Time (BJT), with station rainfall overlaid (green dots: $0 < R < R_{P_{99}}$; magenta dots: $R \geq R_{P_{99}}$). (b) and (c) Enlarged views of the red dashed rectangle in (a): (b) shows rainfall at stations 200–400 km from the front (the warm-sector candidate region); (c) additionally shows the warm-sector mask (orange shading) and rainfall at stations falling within it. Open black circles denote stations with rainfall ($R > 0$) that fall outside the study region or fail the warm-sector criteria. (d) Hourly rainfall time series at QF station. The dashed line marks $R_{P_{99}} = 23.22 \text{ mm h}^{-1}$; the dash-dotted line marks the first hour reaching $R_{P_{99}}$, with all rainfall hours from onset to this hour required to lie within the warm sector. Yellow shading denotes dry periods ($\geq 3 \text{ h}$) bounding the event, and the magenta triangle marks the hour identified as WSHR.

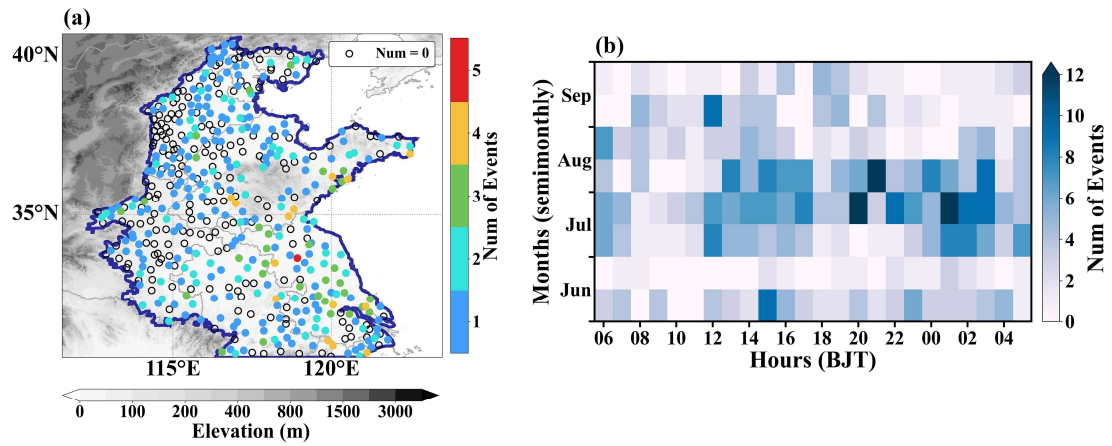


Figure 4. Spatiotemporal distribution of WSHR events over the eastern China plains during the warm seasons (June–September) of 2021–2025. (a) Station-wise counts of WSHR events; open black circles mark stations with zero detected events (Num = 0). Grayscale shading denotes terrain elevation. (b) WSHR event counts by semimonthly period and the hour of the onset of rainfall.

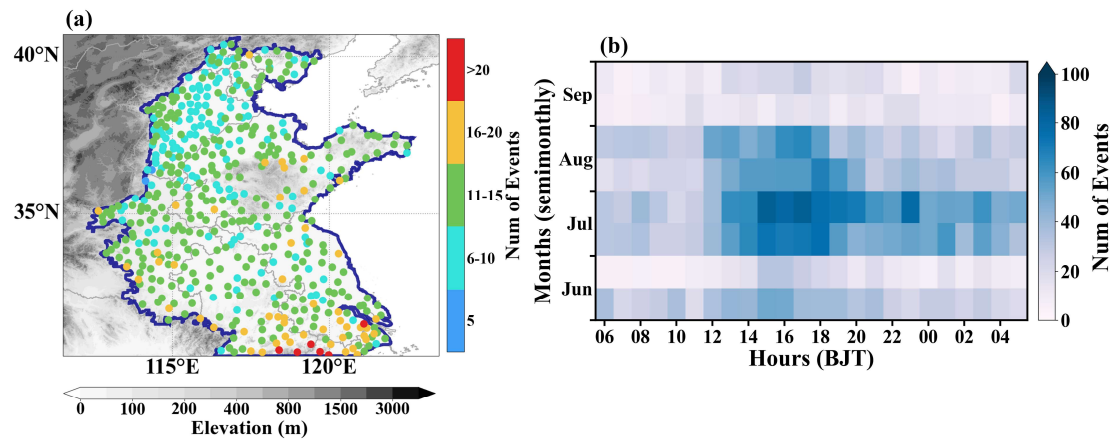


Figure 5. Same as Fig. 4, but for other heavy rainfall events.

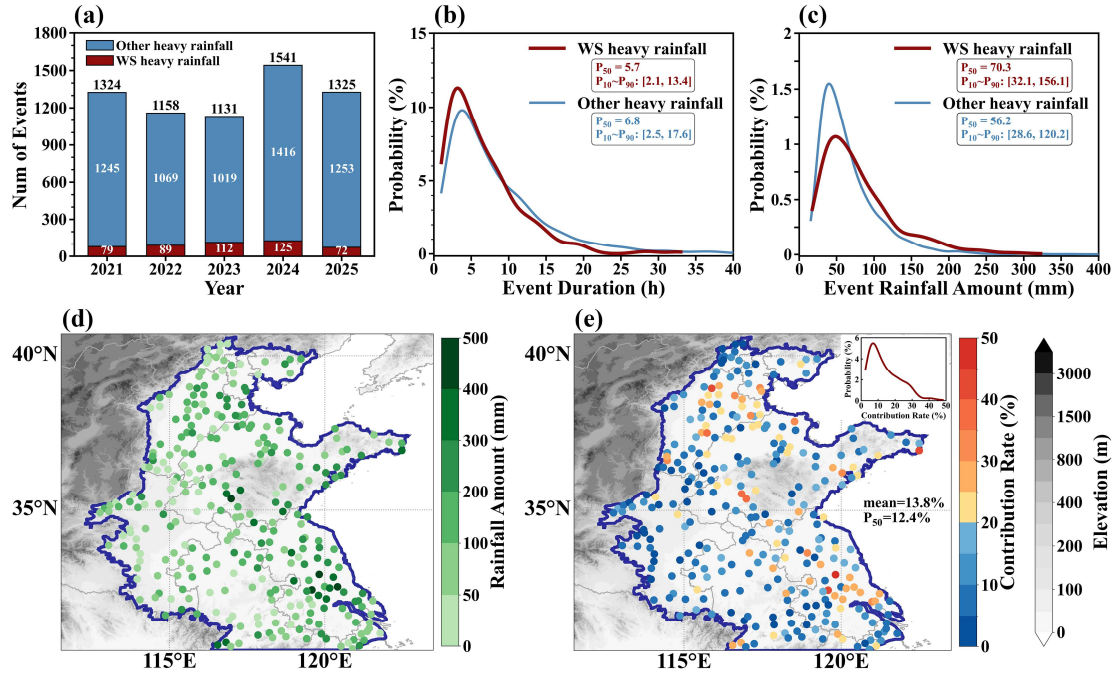


Figure 6. Comparison of WSHR and other heavy rainfall events during the warm seasons of 2021–2025. (a) Annual event counts (WS, dark red; other, light blue); numbers inside the bars are category counts, numbers atop are annual totals. (b) Probability density function (PDF) of event duration. (c) PDF of event cumulative rainfall amount. Legends in (b) and (c) give the median (P_{50}) and the 10th–90th percentile range (P_{10} – P_{90}). (d) Cumulative rainfall amount from WSHR events at each station in the study region. (e) Station-level contribution rate of WSHR to total heavy-rainfall accumulation; the inset shows the PDF of contribution rates among the 297 stations with at least one WSHR event, with mean and P_{50} annotated.

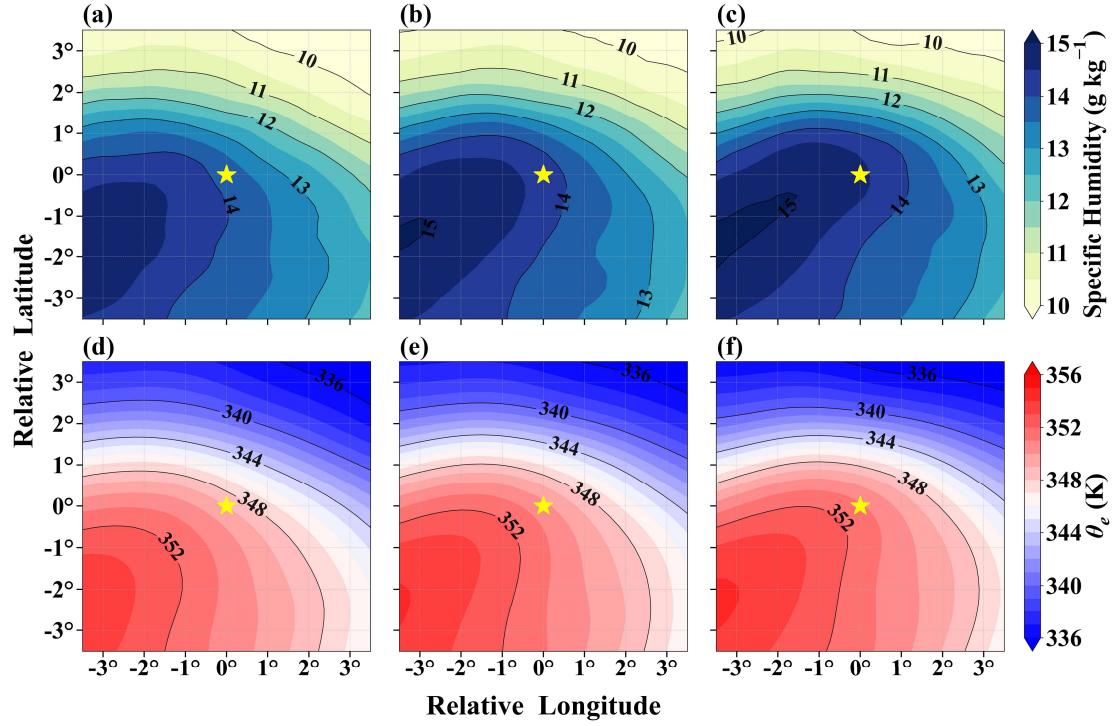


Figure 7. Composite 850-hPa specific humidity (a–c; shading, $g\ kg^{-1}$; contours every $1\ g\ kg^{-1}$) and pseudo-equivalent potential temperature θ_e (d–f; shading, K; contours every 4 K) fields for WSHR events, centered on the event target station (yellow star at origin). Columns correspond to 6 h before onset (a, d), 3 h before onset (b, e), and the onset hour (c, f).

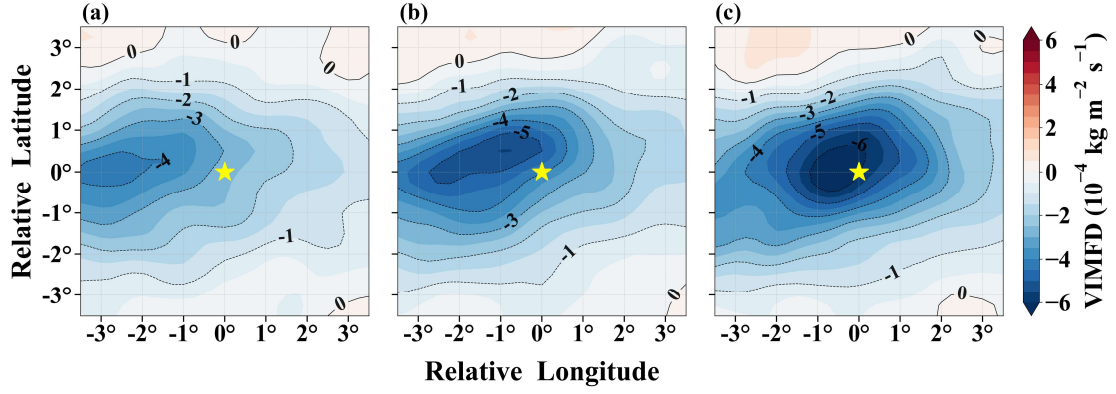


Figure 8. Composite low-level (1000–700 hPa) vertically integrated moisture flux divergence (VIMFD; shading, $10^{-4} \text{ kg m}^{-2} \text{ s}^{-1}$; contours every $1 \times 10^{-4} \text{ kg m}^{-2} \text{ s}^{-1}$) for WSHR events, centered on the event target station (yellow star at origin). Panels correspond to 6 h before onset (a), 3 h before onset (b), and the onset hour (c). Negative values (blue) indicate moisture flux convergence.

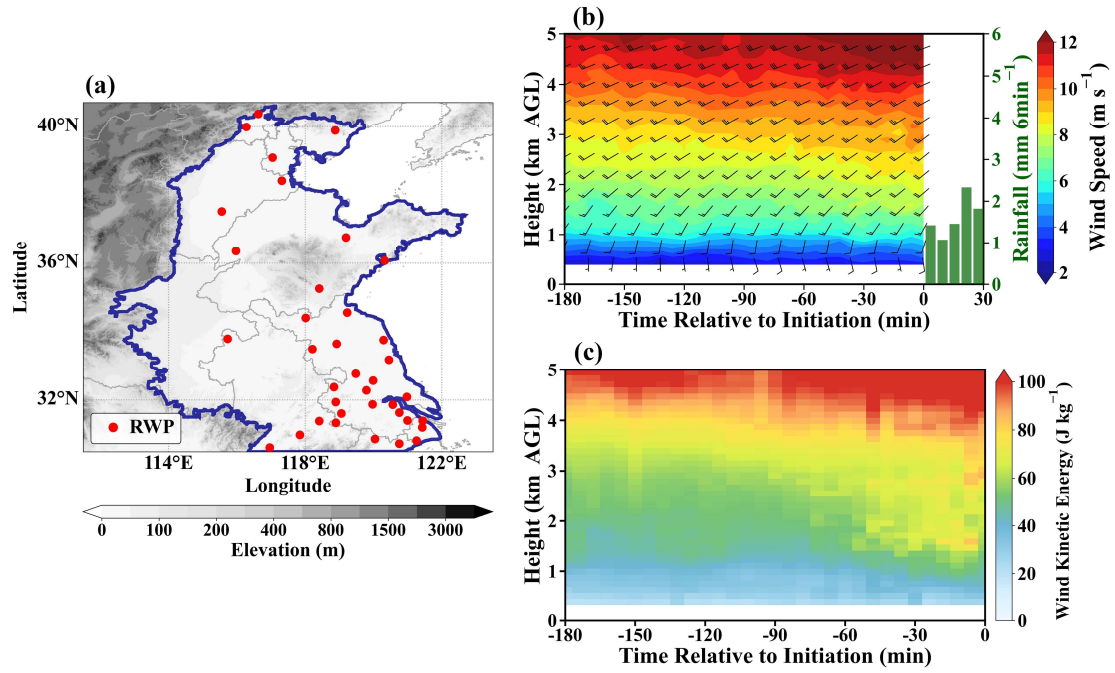


Figure 9. RWP observations and the composite time–height evolution of the low-level wind preceding the onset of WSHR events. (a) RWP stations with records of WSHR events (red dots) used to examine the precursor dynamic signals, overlaid on terrain elevation (grayscale, m). (b) Time–height composite of wind speed (shading, m s^{-1}) and wind barbs; green bars show composite rainfall (mm 6 min^{-1} , right axis). (c) Time–height composite of wind kinetic energy (shading, J kg^{-1}). In (b) and (c), the abscissa is time relative to the onset of rainfall (min) and the ordinate is height above ground level (km AGL).

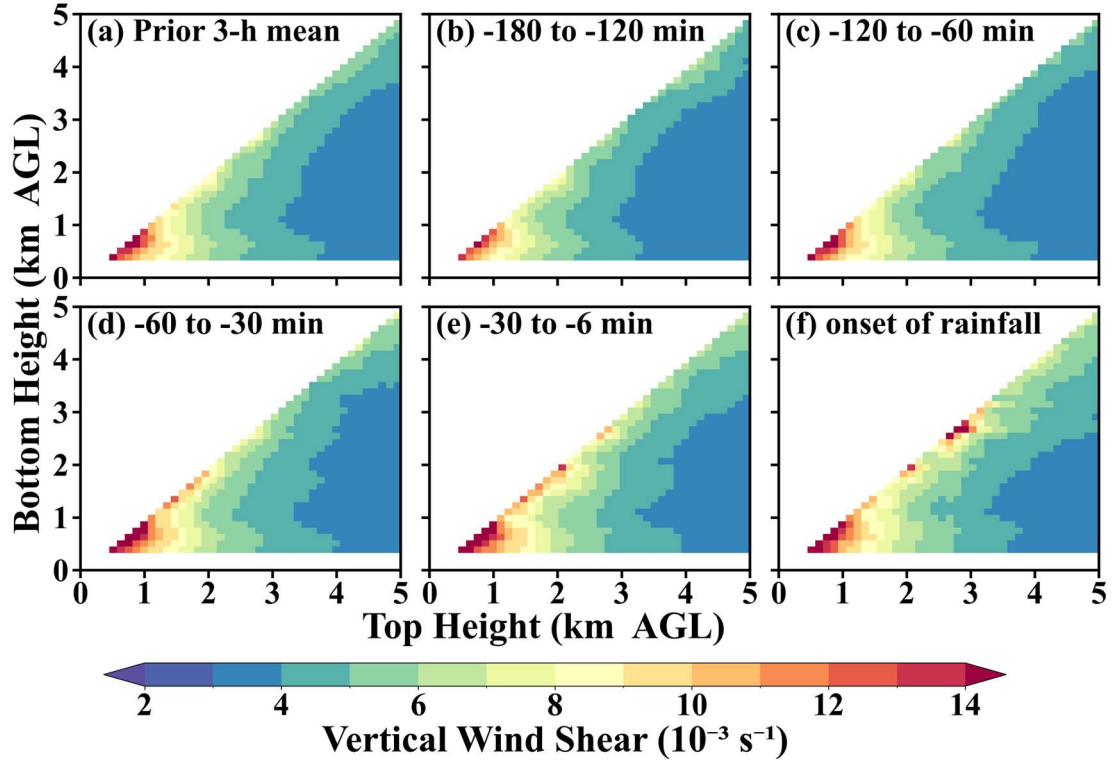


Figure 10. Composite vertical wind shear (shading, 10^{-3} s^{-1}) preceding the onset of WSHR events. (a) prior 3-h mean, (b) -180 to -120 min, (c) -120 to -60 min, (d) -60 to -30 min, (e) -30 to -6 min, and (f) the onset of rainfall. The abscissa and ordinate denote the top and bottom heights (km AGL), respectively, of the layer over which the bulk shear is computed.

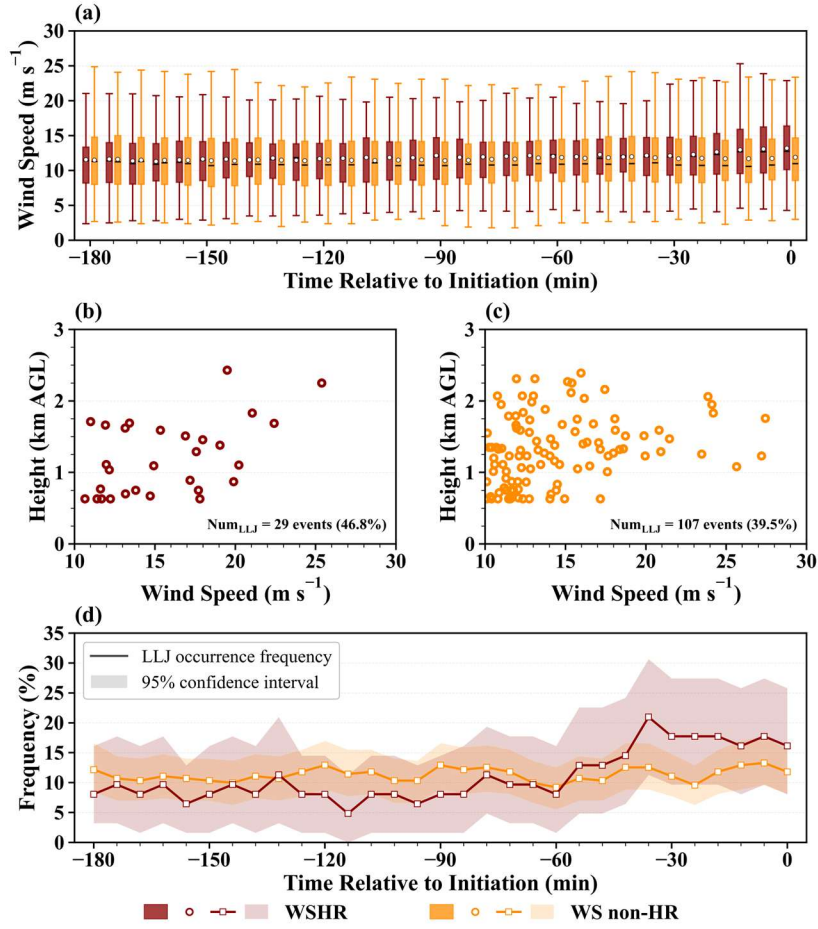


Figure 11. Low-level wind characteristics during -180 to 0 min relative to local rainfall onset for 62 WSHR (dark red) and 271 WS non-HR (orange) events. (a) Box plots of maximum horizontal wind speed below 3 km AGL; black lines and white circles indicate medians and means, respectively. (b, c) Event-mean LLJ core speed and height for (b) WSHR and (c) WS non-HR, averaged over LLJ-containing profiles. Annotations give the number and percentage of events containing an LLJ at least once during the window. (d) LLJ occurrence frequency at 6-min intervals, with shading indicating pointwise 95% event-bootstrap confidence intervals. Box-plot definitions are given in Section 2.4, and bootstrap procedures in Section 3 of the Supplement.

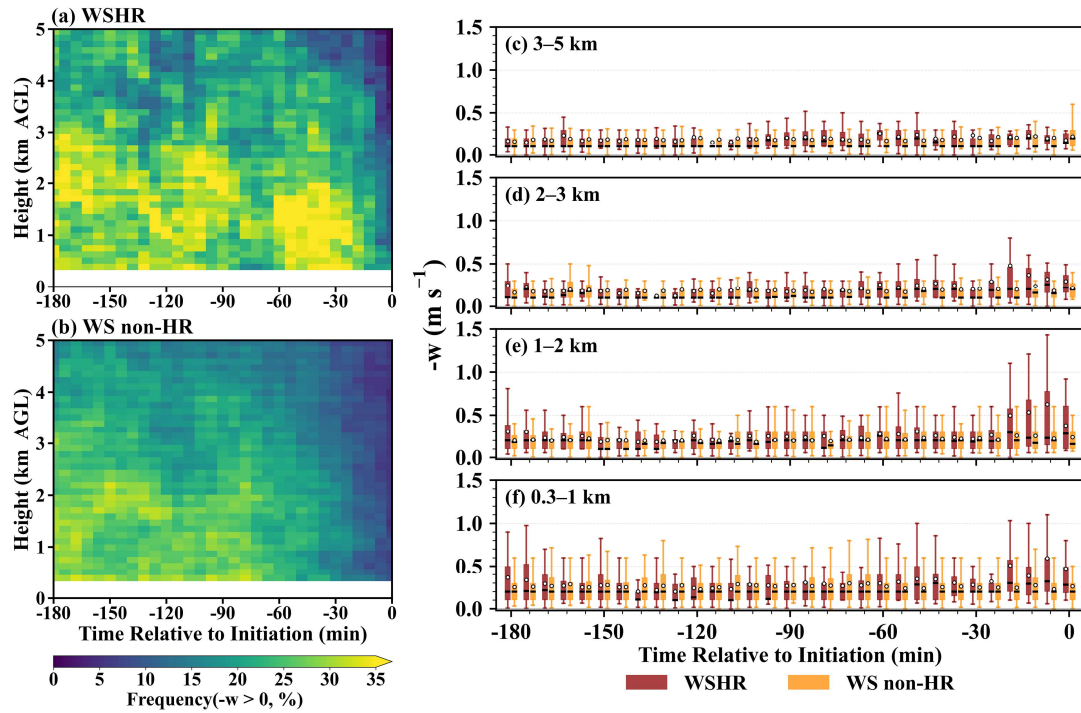


Figure 12. Composite ascending motion preceding the onset of WSHR events and WS non-HR events. (a, b) Frequency of ascending motion ($-w > 0$; %) as a function of height (km AGL) and time relative to the rainfall onset (min) for (a) WSHR events and (b) WS non-HR events. (c–f) Box plots of the across-event distributions of layer-maximum ascent speed ($-w$; m s^{-1}) at each 6-min lead time for the WSHR (dark red) and WS non-HR (orange) events in the (c) 3–5 km, (d) 2–3 km, (e) 1–2 km, and (f) 0.3–1 km AGL layers.