

Article

Rapid Intensification of Typhoon Rammasun (2014) with Strong Vertical Wind Shear

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Abstract: From a traditional point of view, the growth of a tropical cyclone (TC) requires that the vertical wind shear (VWS) should be weak. However, Typhoon Rammasun (2014) underwent a rapid intensification (RI) even in the presence of a strong VWS background. This study investigates the counterintuitive phenomenon, using the multiscale window transform (MWT) and the theory of canonical transfer. For the first time, the diagnostic results show that the strong VWS provided additional available potential energy (APE) to the mid-to-upper troposphere through baroclinic instability. This APE was converted into kinetic energy (KE) via buoyancy conversion and transported to the lower troposphere by pressure gradient, increasing the lower-troposphere wind speed. The strong VWS facilitated the RI in two main ways. First, it was via baroclinic instability. Strong VWS facilitated the transfer of APE from the background flow window to the typhoon scale window, supplying additional APE to the mid-to-upper troposphere, hence enhancing the warm-core structure. Second, the VWS direction shifted from an east-west orientation to a north-south orientation. This directional change put the typhoon's vertical alignment from a westward tilt back to a straighter one. This effectively suppressed the destructive effects of the asymmetric circulation, and promoted the conversion of APE into KE via buoyancy conversion, hence contributed to the RI.

Keywords: rapid intensification (RI); strong vertical wind shear (VWS); baroclinic canonical transfer; multiscale window transform



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1. Introduction

The rapid intensification (RI) of a tropical cyclone (TC) is typically defined as an increase in the maximum sustained wind speed of at least 30 knots (approximately 15.4 m/s) within a 24 h period [1]. Offshore RI events have gained particular attention due to their potential for significant destruction. Despite considerable advancements in tropical cyclone track forecasting over recent decades, improvements in intensity forecasting remain limited [2]. In the South China Sea, for example, landfalling typhoons frequently experience offshore RI, complicating the prediction process. Prominent cases, such as Typhoon Rammasun (2014) and Typhoon Mangkhut (2018), highlight this challenge, underscoring the importance of studying RI events in this region to enhance the accuracy of intensity forecasts for the South China Sea TCs.

A crucial theoretical concept regarding RI is the maximum potential intensity (MPI). It represents the theoretical maximum intensity that a typhoon can achieve under specific

environmental conditions. This concept was introduced by Emanuel, who conceptualized a typhoon as an ideal Carnot heat engine and derived a series of equations to explain the MPI [3–5]. The theory is based on assumptions of slantwise moist neutrality, hydrostatic balance, and gradient wind balance. From this, previous studies have indicated that RI may be brought about by a variety of factors, including the oceanic state, the internal structure of the typhoon, and the surrounding synoptic-scale environment. The ocean state, particularly the state of the sea surface temperature (SST), plays a pivotal role in providing the energy required for intensification, directly influencing the MPI that a storm can achieve [3,6–8]. Research has demonstrated a strong correlation between the SST and the rate of TC intensification [9,10]. Additionally, the tropical cyclone heat potential (TCHP), especially the depth of the 26 °C isotherm, significantly impacts both the intensity and the likelihood of RI [11,12]. Moreover, mesoscale oceanic processes also have a significant impact on the RI of typhoons [13,14].

The internal structure of a TC also plays a crucial role in RI. The radius of maximum winds (RMW) and convective bursts are key factors influencing the onset of RI [15,16]. Moreover, the relative intensity of the TC and its translation speed can modulate the intensification process, with faster-moving storms exhibiting asymmetric core structures that inhibit intensification [17,18]. Guo and Tan [19] proposed the concept of tropical cyclone fullness and showed that higher fullness is associated with stronger cyclone intensities, suggesting that a fuller structure may correspond to a more likely intensification.

From a synoptic perspective, weak vertical wind shear (VWS) is typically conducive to RI, especially in the layer between 300 and 1000 hPa, whereas VWS exerts a more pronounced suppressive effect on TC intensification [1,20]. Additionally, high relative humidity in the middle to lower troposphere and elevated convective available potential energy further support intensification [21,22]. Strong upper-tropospheric divergence also accelerates the intensification process [1,21]. Therefore, RI results from a complex interaction of atmospheric and oceanic processes.

As a result, strong VWS is often viewed as a key inhibitor of TC intensification. Jones [23,24] demonstrated that VWS induces a tilt in the TC's vortex structure, reducing its vertical alignment and thereby weakening the storm [25–27]. Kaplan et al. [21] developed a prediction model for RI in tropical cyclones, attempting to identify favorable environmental conditions, such as a high SST, weak wind shear, and high middle- to lower-tropospheric moisture. However, some TCs have undergone RI even under conditions of strong VWS, as evidenced by several cases of coastal intensification.

Typhoon Rammasun (2014) is such an example, as observed in the South China Sea. An unexpected observation is that its significant intensification was accompanied by strong vertical wind shear (VWS) between 200 and 850 hPa (>11 m/s). Many studies have been conducted to explain this counterintuitive phenomenon. Qiu et al. [28] attributed it to the combined effects of the southwest monsoon and the South Asian High, which supplied abundant moisture to the middle and lower troposphere, triggering enhanced convection and leading to an asymmetric storm structure. Their research showed that Rammasun's RI was not severely inhibited by the high VWS; rather, the storm adjusted its vertical structure, reducing the vortex tilt through counterclockwise rotation, allowing the core to gradually symmetrize and maintain warm core intensification. Additionally, Zhu et al. [29] found similar results for Hurricane Bonnie (1998), where VWS did not inhibit intensification but instead facilitated deepening under specific conditions, such as strong updrafts along the eyewall and high SSTs. Reasor et al. [30] proposed an elastic mechanism that dampened the vortex tilt induced by VWS, preserving the upright vortex structure. Despite the negative structural effects of strong VWS, such asymmetries can be compensated for by convective bursts and enhanced warm-core processes, ultimately allowing RI to occur [15,31].

While these studies have provided valuable insights, they have not fully addressed the underlying energy transfer mechanisms that allow a typhoon to intensify under strong VWS. In fact, cyclogenesis is a typical atmospheric process that involves the interaction between different ranges of scales, which we hereafter will refer to as scale windows. Within this framework, RI can be understood as an energy burst process within the scale window where typhoons reside. How a typhoon acquires such a portion of energy is, therefore, of central importance and may be the key to the problem. This has been overlooked in essentially all the previous studies. During the past decades, a methodology based on a functional analysis apparatus, namely, the multiscale window transform (MWT), has been developed to address these multiscale interactions. We, in this study, then would like to apply it to examining Typhoon Rammasun, which, though with a strong VWS, still rapidly intensified, in the hope of gaining new insights into the complex tropical cyclone dynamics. The next section shows data and methods. Section 3 reviews the second rapid intensification (RI) of Rammasun and presents the multiscale setup and the reconstructions on the thus-set scale windows. In Section 4, we discuss the dynamical processes underlying the RI of Typhoon Rammasun. This study is concluded in Section 5.

2. Data and Method

2.1. Data

Grid data over the Western Pacific and the South China Sea were obtained from ERA5 [32], and the track and intensity data of Typhoon Rammasun were downloaded from the International Best Track Archive for Climate Stewardship (IBTrACS, [33,34]).

ERA5 reanalysis data, provided by the European Centre for Medium-Range Weather Forecasts (ECMWF), offer hourly estimates for a large number of atmospheric, ocean-wave, and land-surface quantities [32]. The data have a fine horizontal resolution of 0.25° by 0.25° . Vertically, they span 37 pressure levels, ranging from 1000 hPa to 1 hPa.

For the purposes of our experiments and analysis, the selected variables included the geopotential, temperature, and wind vector (u, v, ω), covering the horizontal range from 10° to 50° N and from 90° E to 180° and the vertical range from 1000 hPa to 50 hPa. Since it is required that the number of time steps is a power of 2, we chose the period from 14 May to 18 September 2014. This yielded a total of 1024 time points (sampled once every 3 h).

The best track of Typhoon Rammasun was provided by the China Meteorological Administration (CMA), which is downloadable from the IBTrACS website [33,34]. The typhoon paths shown in the ERA5 data, which we used, are consistent with these tracks (they can be observed in subsequent analyses and figures).

2.2. Methodology

The research methodology for this study is the multiscale window transform by Liang and Anderson [35], the MWT-based theory of localized multiscale energetics analysis by Liang and Robinson [36], and the theory of canonical transfer [37]. The remaining part of this section provides a brief introduction; for details, refer to Liang (2016) [37] and the papers as cited.

Multiscale window transform (MWT) is a functional analysis tool developed particularly for multiscale studies. It can orthogonally decompose a field into several parts (as many as desired), each residing in a subspace that includes a range of scales, or scale windows as called. Different from other methods, such as filters for multiscale decomposition, MWT provides, similar to Fourier analysis, transform coefficients, as well as reconstructions (filtered fields). This allows us to represent energy in terms of the transform coefficients, which is impossible with the classical filters. A detailed description is beyond the scope of this paper; details can be found in [35], or, for a concise introduction, see the second

section of [37]. In the following, we just write the MWT of a field, say T , symbolically written as $\widehat{T}_n^{\sim\omega}$, where n stands for the time step, and ω for the scale window. Given a series $\{T(t)\}$ with a duration τ , a scale level j corresponds to a period $2^{-j}\tau$. In this study, we will need a background flow window, a TC-scale window, and a high-frequency window. For convenience, we denote these windows as 0, 1, 2, respectively. These windows are bounded above by three scale levels, j_0 , j_1 , and j_2 , or by three scale bounds, $2^{-j_0}\tau$, $2^{-j_1}\tau$, and $2^{-j_2}\tau$. The demarcation of these windows, i.e., the determination of the scale levels, is shown later in Section 3.

Based on MWT, the kinetic energy (KE) and available potential energy (APE) in the three scale windows can henceforth been obtained. Please refer to [37]; here we just write out the result in a symbolically way:

$$\frac{\partial K^{\omega}}{\partial t} + \nabla \cdot \mathbf{Q}_K^{\omega} = F_K^{\omega} - \nabla \cdot \mathbf{Q}_P^{\omega} - b^{\omega} + \Gamma_{K,P}^{\omega} + \Gamma_{K,h}^{\omega} \quad (1)$$

$$\frac{\partial A^{\omega}}{\partial t} + \nabla \cdot \mathbf{Q}_A^{\omega} = \Gamma_A^{\omega} + b^{\omega} + S_A^{\omega} + F_A^{\omega} \quad (2)$$

For simplicity, the time step n is suppressed, and the explanations and expressions for these terms are listed in Table 1. Here, the Γ terms are of particular importance. They represent the transfers of energy (KE or APE) between the different scale windows. It is worth noting that they are very different from those in classical formalisms. Particularly, there is an interesting property, i.e.,

$$\sum_{\omega} (\sum_n \Gamma_n^{\omega}) = 0 \quad (3)$$

which was first shown in [36] and later rigorously proved by [37]. This means that, physically, energy transfers across different scale windows within a system will neither be created nor destroyed; in other words, energy should be conserved. It merits mentioning that this property, although it appears obvious, does not actually hold in previous Lorenz-type energetics formalisms. To distinguish, we term such a transfer as a “canonical transfer”. In [37], it is shown that a canonical transfer bears the form of a Lie bracket, reminiscent of the Poisson bracket in Hamiltonian dynamics, and satisfies the Jacobian identity. Most of all, the $0 \rightarrow 1$ component of Γ_A^1 is proven by [38] to correspond exactly to an important dynamical process in geophysical fluid dynamics, i.e., the baroclinic instability of the basic flow. Likewise, that component of Γ_K^1 is proven to correspond to barotropic instability. In this study, we pay particular attention to these terms.

Table 1. A list of the symbols in (1) and (2) from [37]. The notations are conventional.

Symbol	Expression	Physical Meaning
K^{ω}	$\frac{1}{2} \widehat{\mathbf{v}}_h^{\sim\omega} \cdot \widehat{\mathbf{v}}_h^{\sim\omega}$	KE in scale window ω
$\mathbf{Q}_K^{\omega}$	$\frac{1}{2} (\widehat{\mathbf{v}\mathbf{v}}_h)^{\sim\omega} \cdot \widehat{\mathbf{v}}_h^{\sim\omega}$	Flux of KE in window ω
Γ_K^{ω}	$\frac{1}{2} [(\widehat{\mathbf{v}\mathbf{v}}_h)^{\sim\omega} : \nabla \widehat{\mathbf{v}}_h^{\sim\omega} - \nabla \cdot (\widehat{\mathbf{v}\mathbf{v}}_h)^{\sim\omega} \cdot \widehat{\mathbf{v}}_h^{\sim\omega}]$	Canonical transfer of KE to window ω
$\mathbf{Q}_K^{\omega}$	$\widehat{\mathbf{v}}^{\sim\omega} \widehat{\Phi}^{\sim\omega}$	Pressure flux
b^{ω}	$\widehat{\omega}^{\sim\omega} \widehat{\alpha}^{\sim\omega}$	Buoyancy conversion
A^{ω}	$\frac{1}{2} c (\widehat{T}^{\sim\omega})^2, c = \frac{g}{\overline{T}(g/C_p - L)}$	APE in scale window ω
$\mathbf{Q}_A^{\omega}$	$\frac{1}{2} c \widehat{T}^{\sim\omega} (\widehat{\mathbf{v}T})^{\sim\omega}$	Flux of APE in window ω
Γ_A^{ω}	$\frac{c}{2} [(\widehat{\mathbf{v}T})^{\sim\omega} \cdot \nabla \widehat{T}^{\sim\omega} - \widehat{T}^{\sim\omega} \nabla \cdot (\widehat{\mathbf{v}T})^{\sim\omega}]$	Canonical transfer of APE to window ω
S_A^{ω}	$\frac{1}{2} \widehat{T}^{\sim\omega} (\widehat{\omega T})^{\sim\omega} \frac{\partial c}{\partial p} + \frac{1}{T} \widehat{\omega} \alpha^{\sim\omega}$	Apparent source/sink (usually negligible)

The theory of canonical transfer and MWT-based energetics analysis have been successfully applied in the study of various real atmosphere–ocean–climate problems, such as storm track [39], squall line [40], North Atlantic Oscillation [41], and cold wave [42], to name just a few. Specifically, in typhoon research, they have also been applied with interesting results (e.g., [43,44]).

3. Multiscale Features of the Second Rapid Intensification (RI) of Rammasun

Typhoon Rammasun (1409) originated in the Northwest Pacific and was named on 12 July 2014. It subsequently moved westward, undergoing continuous intensification. Upon making landfall in the Philippines, its intensity first increased and then decreased, reaching a peak at 1200 UTC 15 July. After entering the South China Sea, Rammasun experienced a second near-coastal RI. It henceforth made landfall as a super typhoon in Wenchang, Hainan Province, and Xuwen, Guangdong Province. Subsequently, it re-entered the sea and ultimately made a final landfall in Guangxi Province. The typhoon caused severe damage to South China, especially Hainan Province. The track and corresponding intensity of Rammasun are shown in Figures 1 and 2. This study primarily focuses on the second near-coastal RI that occurred in the South China Sea, as it occurred with strong VWS. Based on the definition of RI provided in Section 1 and the wind speed data from the China Meteorological Administration (CMA), the second RI event occurred at 0300 UTC 17 July, with the storm reaching its peak intensity at 0600 UTC 18 July. At this time, it reached a wind speed of 139 knots and a minimum central pressure of 888 hPa, as was, therefore, classified as a super typhoon. This period is indicated by the gray shaded area in Figure 2 and is the primary focus of this study.

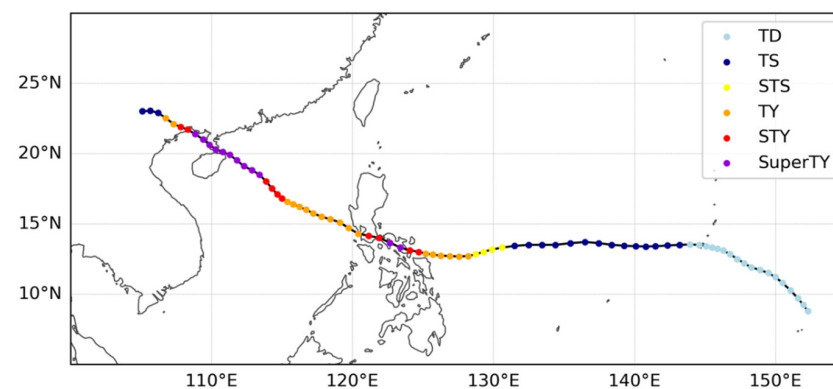


Figure 1. The track of Typhoon Rammasun. Each point represents the typhoon's position at different times, with different colors indicating its varying intensities: TD—tropical depression, TS—tropical storm, STS—super tropical storm, TY—typhoon, STY—severe typhoon, SuperTY—super typhoon.

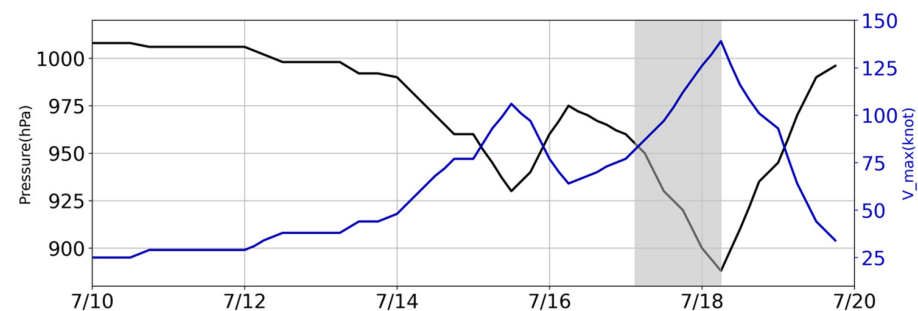


Figure 2. Intensity variations of Typhoon Rammasun over time. The black line denotes the minimum central atmospheric pressure, the blue line represents the maximum sustained wind speed, and the grey shaded area delineates the period from the onset of the second RI.

3.1. Scale Window Setup and Reconstruction Fields

The scale window, as introduced in Section 2, is set up with the aid of wavelet analysis. We select a point within the region where the rapid intensification (RI) occurred (19° N, 113.25° E) and analyze the geopotential time series at the 925 hPa level. As the entire time series consists of 1024 time points, $j_2 = 10$. Using wavelet analysis, we obtain a wavelet power spectrum, as shown in Figure 3. The spectrum clearly indicates that the typhoon-related process was concentrated in the 1–16 day window. This observation is consistent with the actual duration of the typhoon, which lasted for 8 days, corresponding to a trough in the geopotential field that lasted approximately 8 days, suggesting a full wave period of around 16 days. Based on these findings, we set $j_0 = 4$, and $j_1 = 8$, ensuring that this division captures the complete range of information in the time series. In this way, the whole functional space is decomposed into a background flow window, a TC-scale window, and a convection-scale window.

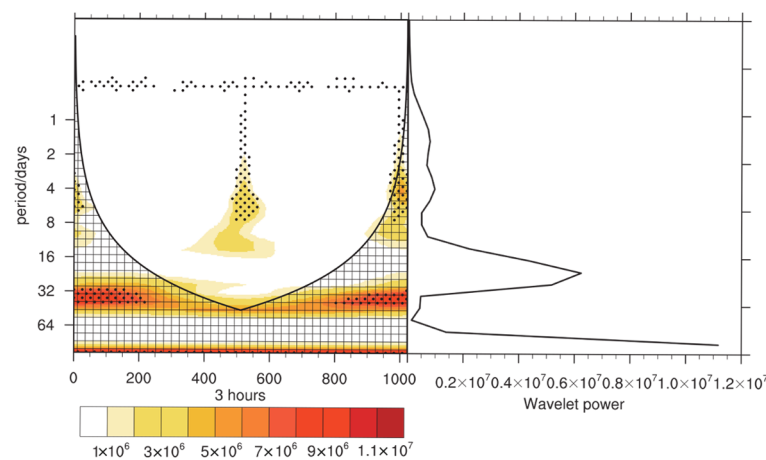


Figure 3. The wavelet power spectrum (m^2/s^2) and global wavelet spectrum (m^2/s^2) of the 925 hPa geopotential field time series at the selected point (19° N, 113.25° E). X-axis in the left panel: the time steps, with each step corresponding to a 3 h interval. X-axis in the right panel: energy/variance summed over time steps. Y-axis in both panels: the period (days). The dotted regions mark areas with statistically significant wavelet power at the 95% confidence level.

To evaluate the decomposition, Figure 4 shows the original field, the TC-scale feature, and the background flow at different altitudes at 0000 UTC 18 July. The results show that most of the typhoon-related information has been well captured in the TC-scale window, and the reconstructed typhoon track closely matches the track provided by the CMA. The large-scale background field remains stable with little variation, while the turbulent field successfully reproduced the convection associated with the typhoon. This validation confirms that the chosen method of window separation is scientifically sound and appropriate.

3.2. Vertical Wind Shear (VWS) in the Reconstructed Field

We follow the previous studies (e.g., [1,22]) to evaluate VWS in a TC. VWS is typically assessed within a radial distance of 220–880 km from the TC center and is categorized into upper-level shear (200–500 hPa), lower-level shear (500–850 hPa), and deep-layer shear (200–850 hPa). In this study, we focus on deep-layer shear, which is the most commonly used. Hereafter, VWS refers to deep-layer shear. The radius for this calculation is set at 500 km.

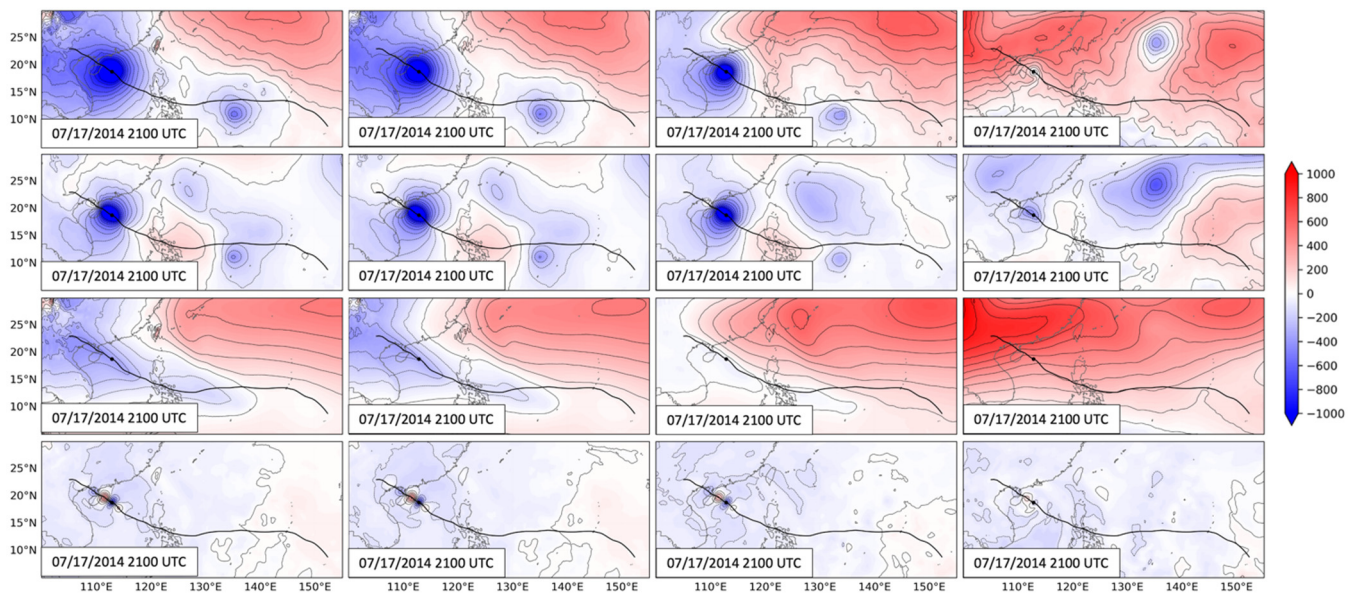


Figure 4. The geopotential anomaly fields of the original field, the TC-scale window, and the background flow window (from top to bottom) at different altitudes at 0000 UTC July 18 (from left to right: 925 hPa, 850 hPa, 500 hPa, 200 hPa; units: m^2/s^2). The black line represents the typhoon's track, with black dots indicating the typhoon's position at the corresponding time; the same applies to other figures below. The colorbar used in this figure is such that negative values are in blue and positive values are in red; the same applies to all figures hereafter.

Figure 5 illustrates the temporal evolution of the VWS magnitude associated with the TC, as well as the specific magnitude and direction of VWS across different spatial scales at four selected time points during RI. In the original ERA5 field, the VWS does not appear particularly strong and shows a noticeable decrease before the onset of RI. However, after scale separation, it is evident that the VWS of the background flow field remained consistently above 11 m/s; that is to say, it remained strong. A trend of significant decline is only observed after the RI ended. The initial decrease in VWS, as observed in the original field, was primarily due to the increased angle between the background VWS direction and the TC-scale window, which masked the actual strength of the background VWS. From another aspect, this verifies the importance of scale separation.

We also compute the wind field difference between 200 hPa and 850 hPa in the background field to observe the overall situation and variation across the entire ocean basin. Figure 6 shows that, during the cyclone's movement, the VWS within the background flow window remained relatively stable. The variations in the cyclone's VWS are more related to its geographical displacement rather than changes in its own circulation. As the cyclone moved from the Western Pacific into the South China Sea, particularly during its RI, the background VWS at the cyclone's location changed significantly, with the easterly component decreasing and the northerly component increasing. At its peak intensity, the VWS became almost purely northerly. Therefore, even though the magnitude of the background VWS remained relatively stable and consistently strong, its directional change across different ocean regions may have significant implications for the cyclone's structure and energy transfer processes. We discuss this in detail in the subsequent sections.

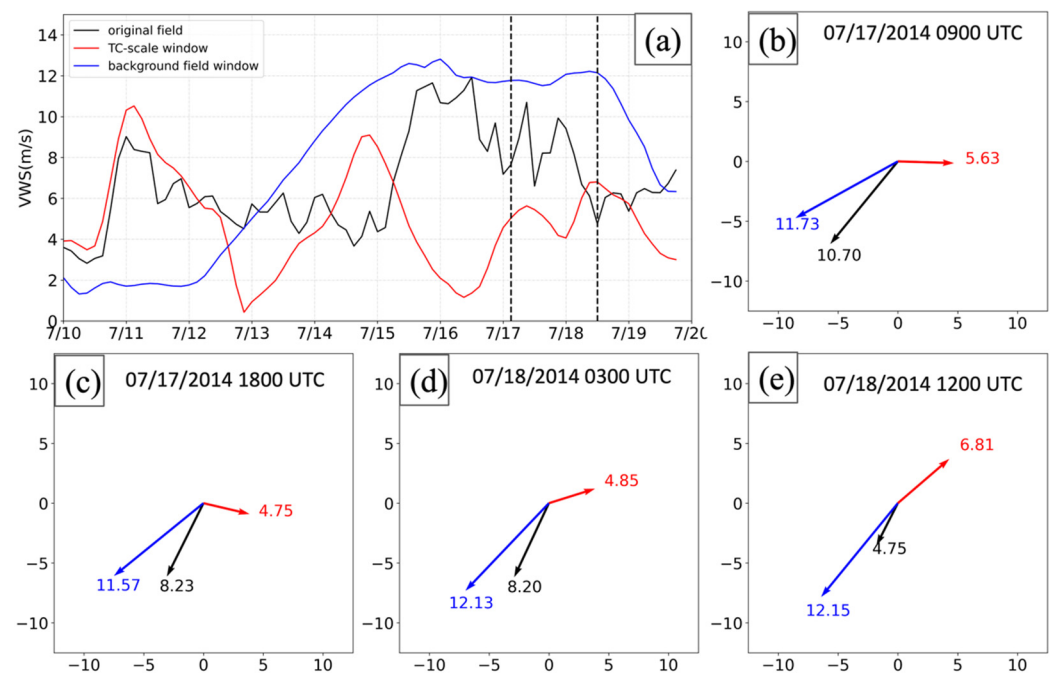


Figure 5. (a) Temporal evolution of the VWS magnitude for different fields, with the RI period indicated between the two black dashed lines. (b–e) The specific magnitude and direction of the VWS (m/s) for the three fields at different time points (black for the original field; blue for the background flow window; red for the TC-scale window). The time period between the two black dashed lines represents the RI period. This convention is applied in all subsequent figures.

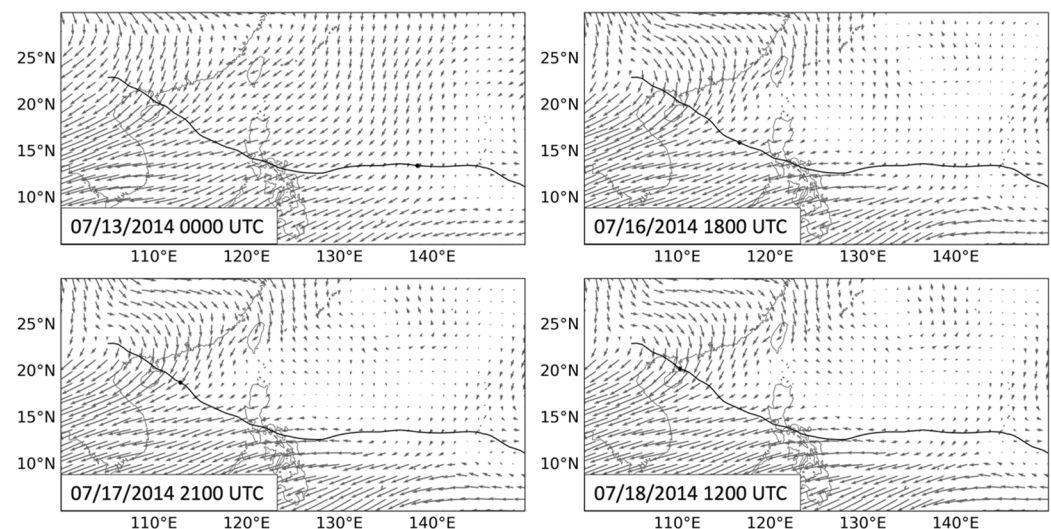


Figure 6. Spatial distribution of VWS at different times: 0000 UTC 13 July, 0000 UTC 15 July, 0000 UTC 17 July, and 1800 UTC 18 July (represented by arrows; the same in the figures hereafter).

4. Dynamical Processes Underlying the Rapid Intensification of Typhoon Rammasun

4.1. Local Energy Transitions Before and After Rapid Intensification

A typical TC features a warm-core structure around 200 hPa and strong winds in the lower troposphere, corresponding to high middle- to upper-tropospheric APE and lower-tropospheric KE. By tracking the evolution of these energy components, particularly the KE in the lower troposphere, we could assess changes in the TC intensity and wind speed. This allows us to identify the dynamic processes influencing the KE and quantify

their contributions, while recognizing the need for a comprehensive view of energy transfer and transport within the atmosphere.

Within the MS-EVA framework, as introduced in Section 2, we evaluate the energetics terms in Equations (1) and (2) in the TC-scale window at different layers. Here, the canonical transfers of APE (Γ_A) and KE (Γ_K) represent the key processes, which prove to be measures of baroclinic instability and barotropic instability, respectively. By definition, $\Gamma_A^{0 \rightarrow 1}$ and $\Gamma_K^{0 \rightarrow 1}$ stand for the energy transfer from the background flow window to the TC-scale window, providing a quantitative evaluation of the cross-scale energy supply for the TC in question. A positive value indicates an energy transfer from the background flow to the TC-scale window, while a negative value indicates the opposite. This study focuses more on the baroclinic instability process in the middle to upper troposphere, $\Gamma_A^{0 \rightarrow 1}$, due to its stronger connection to VWS. We analyze $\Gamma_A^{0 \rightarrow 1}$ to investigate whether strong VWS triggered baroclinic instability, enabling the TC-scale window to gain energy from the background flow. Additionally, the term F_A^1 primarily stems from latent heat release due to water vapor from the ocean, which serves as a fundamental source for the maintenance of the typhoon's structure and internal energy. Meanwhile, F_K^1 predominantly represents work done by frictional forces. By synthesizing and analyzing these figures, we aim to gain insight into the dynamic processes within Typhoon Rammasun, and the second RI in particular.

Based on the results, we calculate the mean values of various energy components within a 500 km radius of the TC center across different layers throughout the entire lifecycle of the typhoon. Figure 7 illustrates the temporal evolution of the TC-scale APE and KE at different layers, while Figure 8 shows the evolution of the corresponding time rates of change. During the RI, one conspicuous feature is the increase in the total APE within the warm core, accompanied by an expansion of the warm-core area and an increase in the lower-tropospheric KE (or wind speed). However, the peak values of APE and KE occur 6 h later than the maximum intensity time shown in IBTrACS, corresponding to two time steps in the MS-EVA experiment. This is normal, and the analysis and related figures in the following discussion reflect this adjustment (i.e., the maximum intensity time was set as 1200 UTC 18 July). During the RI, the magnitudes of all energy terms increase, and the associated energy processes become more intense: latent heat release intensifies, the thresholds for APE transport ($-dQ_A^1$) shift, and the areas influenced by buoyancy conversion and pressure-gradient force expand. Additionally, there is a notable cross-scale energy transfer. As mentioned in Section 2, the canonical transfers of APE ($\Gamma_A^{0 \rightarrow 1}$) and KE ($\Gamma_K^{0 \rightarrow 1}$) correspond to the fundamental concepts of baroclinic instability and barotropic instability, respectively.

For Rammasun, $\Gamma_A^{0 \rightarrow 1}$ shows a weak signal when the TC is in the Pacific but increases during the first RI. Before the second RI, $\Gamma_A^{0 \rightarrow 1}$ weakens but then steadily increases and then peaks around the time when Rammasun reaches its maximum intensity. Notably, $\Gamma_A^{0 \rightarrow 1}$ is concentrated within the 125–275 hPa layer, centered around 200 hPa. This region differed from the latent heat release zone but served as a critical outflow layer for the cyclone. Therefore, the energy contributions via baroclinic instability at this level were significant and cannot be ignored.

A question naturally arises: how does APE gained by the warm core in the upper troposphere contribute to the increase in kinetic energy (KE) in the lower troposphere? Figure 9 illustrates the energy pathway derived from the analysis presented in Figure 8. The two figures show comparisons of the local energy processes before and after the second RI of Rammasun. Before RI (serving as a comparison), APE in the 200–500 hPa layer is primarily gained through latent heat release. A portion of this APE is transported upward due to a divergence of APE flux (mainly caused by vertical flux divergence), both of which sustain the warm-core structure near 200 hPa. Buoyancy conversion leads to the conversion

of the middle- to upper-tropospheric APE into KE, which is then transported to the lower troposphere through pressure-gradient force, leading to high wind speed near the surface. A portion of the KE in the lower troposphere is also lost due to frictional dissipation (presented in Figure 8). Overall, APE and KE are closely connected within the TC, and variations in APE in the middle to upper troposphere significantly influenced the change in KE in the lower troposphere.

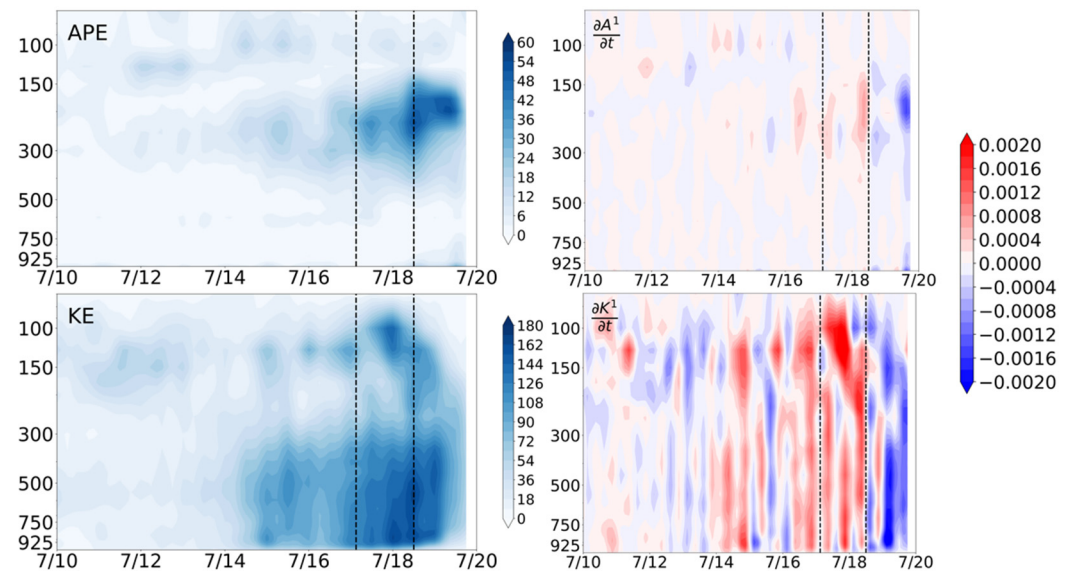


Figure 7. The time evolution of available potential energy (APE), kinetic energy (KE), and their rates of change at different vertical layers (specific values are indicated in the upper-left corner of the figure). The units for the energy terms are m^2/s ; for the rates of change, the units are m^2/s^2 .

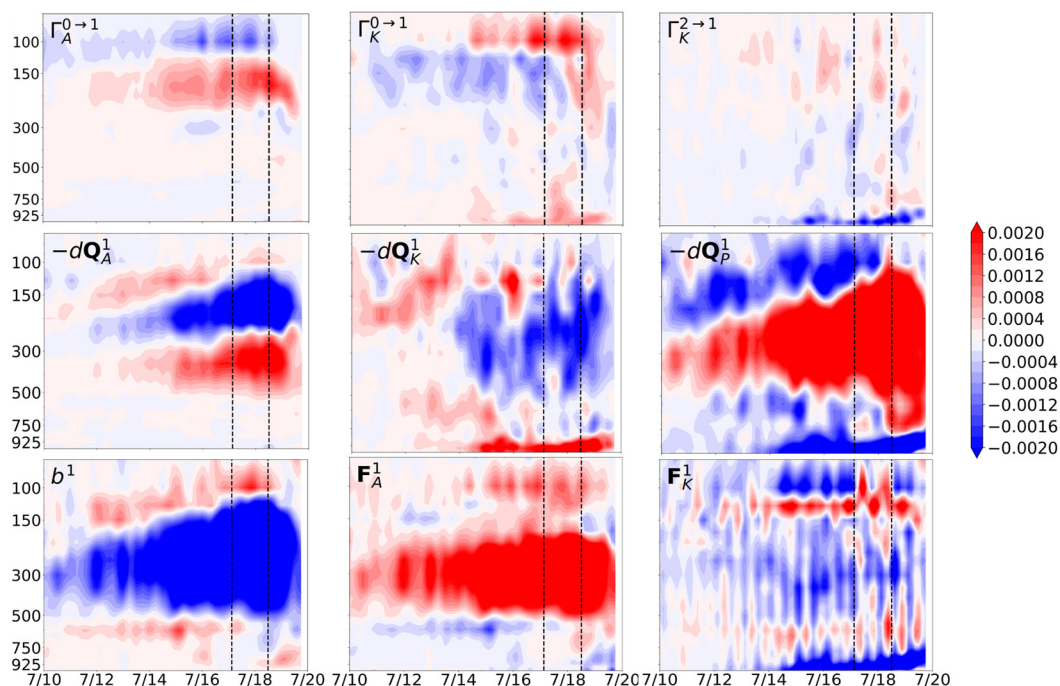


Figure 8. The time evolution of energy components, where positive values indicate energy gained in the TC-scale window, and negative values represent energy loss (averaged over grid points within a 500 km radius of the center; units: m^2/s^2).

where the overbar indicates averaging over the time interval of the TC-scale window in question, and the prime stands for the departure from the average. Note, this Reynolds decomposition is in general different in value from the corresponding MWT as applied (where $j_0 \neq 0$), though it may be true when the interval is small or when the process is stationary. Since here we are just for the purpose of interpretation, this will suffice.

Equation (4) can be further expanded to be

$$\Gamma_A^{0 \rightarrow 1} = \underbrace{\frac{\epsilon}{2} \overline{T} \frac{\partial \overline{u'T'}}{\partial x}}_{(1)} + \underbrace{\frac{\epsilon}{2} \overline{T} \frac{\partial \overline{v'T'}}{\partial y}}_{(2)} + \underbrace{\frac{\epsilon}{2} \overline{T} \frac{\partial \overline{\omega'T'}}{\partial p}}_{(3)} - \underbrace{\frac{\epsilon}{2} \overline{u'T'} \frac{\partial \overline{T}}{\partial x}}_{(4)} - \underbrace{\frac{\epsilon}{2} \overline{v'T'} \frac{\partial \overline{T}}{\partial y}}_{(5)} - \underbrace{\frac{\epsilon}{2} \overline{\omega'T'} \frac{\partial \overline{T}}{\partial p}}_{(6)} \quad (5)$$

Here, the first three terms are divergence terms, while the last three are eddy heat flux terms (each including contributions from the zonal, meridional, and vertical directions). We pay particular attention to the latter, as they are closely associated with strong VWS. When T' is replaced by the velocity perturbation, they are just the contributions from Reynolds stress. For simplicity, we refer to these six terms from left to right as “①” through “⑥” in subsequent discussions.

To determine which of these six terms—or, in other words, which physical processes—play a dominant role in baroclinic instability and their potential connection to strong VWS, we calculate the mean values of these terms within a 500 km radius of the TC center and examined their temporal evolution. As shown in Figure 11, the primary contributor is term ⑥, with terms ⑤ (the meridional eddy flux term) and ② (the meridional divergence term) also contributing to the initial stages. In contrast, term ④ (the zonal eddy flux term) and other divergence terms act as suppressors. Notably, the magnitude of term ⑥ is significantly larger than that of the other terms, making it the most substantial contributor to the baroclinic canonical transfer during the RI.

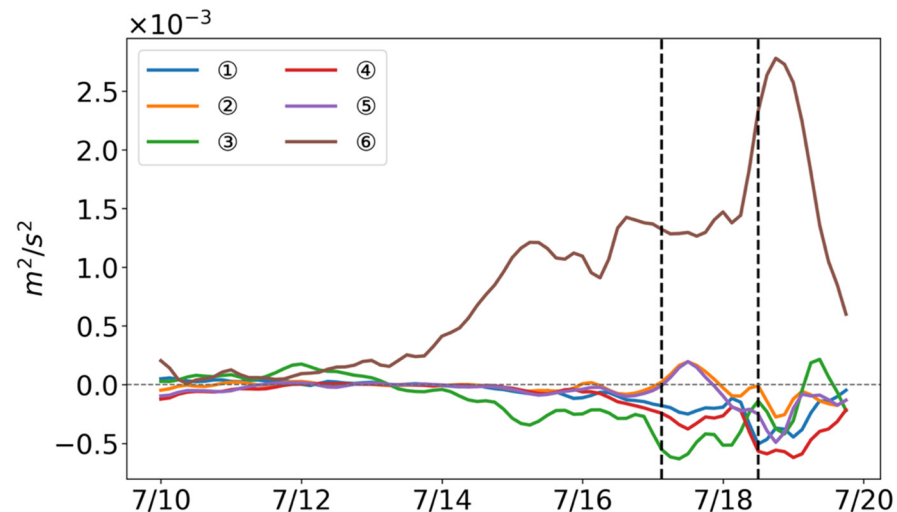


Figure 11. The temporal evolution of different components of $\Gamma_A^{0 \rightarrow 1}$ (as shown in Equation (5)) within a 500 km radius of the typhoon center as it moved. (The terms ①–⑥ are expressed as the respective terms of the right hand side of Equation (5)).

Term ⑥ is the largest contributor (its horizontal distribution is shown in Figure 12a–c and is highly consistent with the distribution of $\Gamma_A^{0 \rightarrow 1}$). From its expression, $-\frac{\epsilon}{2} \overline{\omega'T'} \frac{\partial \overline{T}}{\partial p}$, we can see that this term is associated with the vertical temperature gradient in the large-scale field and the coupling between the vertical velocity and temperature distribution within the TC-scale window. In previous studies, Frank and Ritchie [25] have already observed that under strong VWS, enhanced upward motion and concentrated precipitation tend to occur on the downshear-left side of the TC. They attributed such asymmetric circulation induced by strong VWS as a cause of TC weakening. Consistent with their findings, the

distribution of $\Gamma_A^{0 \rightarrow 1}$ and term ⑥ in this experiment is also concentrated on the downshear-left side of the VWS. Additionally, we observe that $\overline{\omega'T'}$, the coupling of vertical velocity and temperature within the TC-scale, is negative during the RI phase due to the warm-core structure and upward motion at 200 hPa. Simultaneously, $\partial\bar{T}/\partial p$ is positive throughout this region, indicating energy acquisition through the eddy heat flux. However, contrary to Frank and Ritchie's conclusions, our energy-based analysis suggests that this process enabled the TC-scale window to gain APE from the background flow window, contributing to consolidate the warm core and facilitate RI (at least for this specific case).

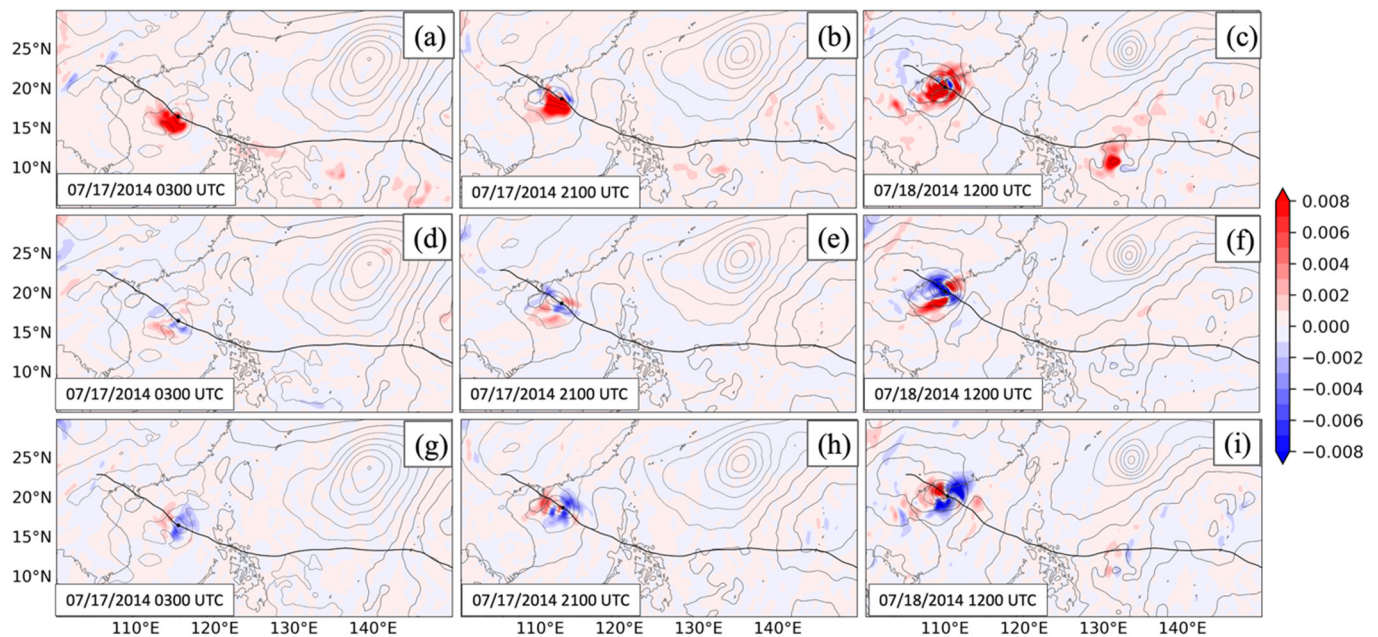


Figure 12. The horizontal distribution of the eddy heat flux terms in $\Gamma_A^{0 \rightarrow 1}$ during RI. (a–c) Term ③; (d–f) term ⑤; (g–i) term ④ (units: m^2/s^2).

Term ⑤ is another significant contributor (its horizontal distribution is shown in Figure 12d–f). Recall the thermal wind relation for geophysical flow:

$$\frac{\partial u_g}{\partial z} = -\frac{R}{f} \frac{\partial T}{\partial y} \quad (6)$$

$$\frac{\partial v_g}{\partial z} = \frac{R}{f} \frac{\partial T}{\partial x} \quad (7)$$

where the symbols are conventional. Thus, the magnitude of term ⑤ is essentially proportional to the vertical shear of the meridional wind (v) in the background flow window. Since the background field is relatively stable, this study presents Figure 13 using the snapshot at 1200 UTC, 18 July, as an example. As shown in Figure 13d, $v'T'$ is positive on the left side and negative on the right side (since T' is positive here, the sign depends on v' , consistent with cyclonic characteristics). Meanwhile, $\partial\bar{T}/\partial y$ is significantly larger on the left side of the typhoon track compared to the right side, which can be observed in Figures 6 and 13c,e. This results in stronger canonical transfer in regions with greater east–west VWS (regardless of the direction). Consequently, the horizontal distribution of term ⑤ exhibits an approximate quadrupole pattern, with its overall positive or negative sign depending on variations in the magnitude of the east–west VWS around the typhoon center. Overall, baroclinic canonical transfer contributes to energy gain during the early stages of RI but exhibited a suppressing effect in the later stages of RI.

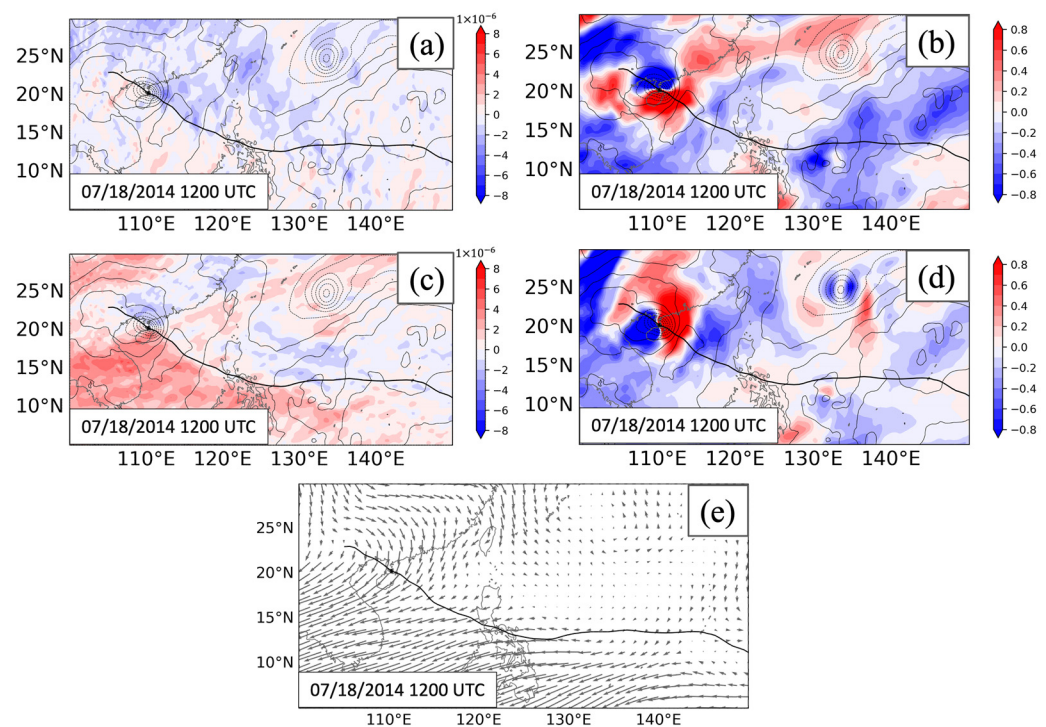


Figure 13. (a–d): Horizontal distribution of the physical quantities involved in Equation (5) at 200 hPa at 1200 UTC 18 July. (a) $\partial \bar{T} / \partial x$; (b) $u'T'$; (c) $\partial \bar{T} / \partial y$; (d) $v'T'$. (e): Horizontal distribution of VWS at the same time.

Similarly, term ④ (the zonal eddy heat flux term) is proportional to the vertical shear of meridional wind (v) based on the thermal wind relationship (its horizontal distribution is shown in Figure 12g–i). As the typhoon moves northward, the north–south VWS significantly intensifies, leading to a continuous increase in the negative value of term ④. As a result, the changing direction of VWS around the typhoon negatively impacts the RI process through the dynamics under terms ④ and ⑤. However, this effect is negligible compared to term ⑥.

In conclusion, while some energy dissipates to the surrounding region and is transported to layers near 200 hPa, the TC still gains energy through baroclinic canonical transfer, which is closely connected to the strong VWS.

4.3. Release of Available Potential Energy Through Buoyancy Conversion

As previously discussed, during RI, Rammasun enhanced its middle- to upper-tropospheric APE through latent heat release, baroclinic instability, and other processes, further deepening its warm-core structure. However, the aspect most relevant to daily life and of primary concern is the wind speed in the lower troposphere, which means an increase in KE. APE can be converted into KE through buoyancy conversion and transported to the lower troposphere by the strong vertical pressure gradient force. In fact, changes in the direction of strong VWS facilitate this process.

Previous studies have pointed out that Rammasun's rapid intensification (RI) was not significantly suppressed by strong VWS [28,45]. Instead, the cyclone adjusted its internal structure, particularly through counterclockwise rotation, which reduced the vertical tilt of the TC center and gradually symmetrized its inner core while maintaining the warm-core enhancement.

As mentioned above, VWS is chosen as the wind speed difference between 200 hPa and 850 hPa. The direction of strong VWS shifted from east–west to north–south during this period, which contributed to a straighter cyclone structure. At the same time,

convection developed to higher levels, further supporting this process. These processes not only facilitated buoyancy conversion, releasing APE into KE to further enhance the typhoon, but also made the upward motion region within the typhoon more symmetric and expanded its spatial extent (shown in Figure 14). Along with this, the enhanced symmetry effectively suppressed the asymmetric development within the TC and further strengthens the contributions of term ⑥, as discussed in Section 4.2.

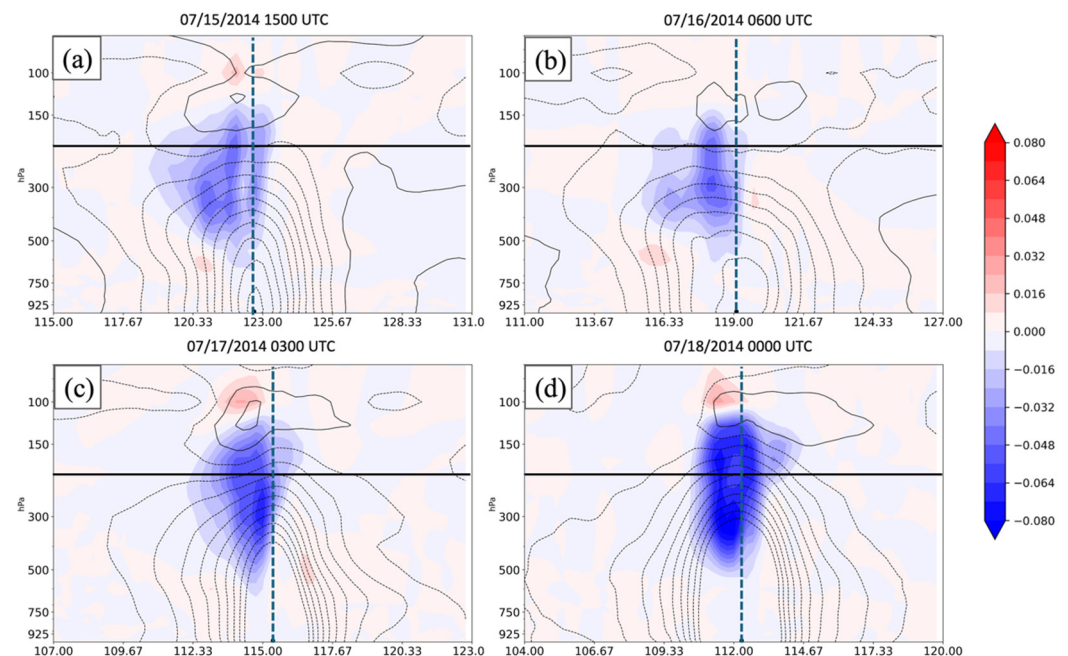


Figure 14. The vertical structure of the potential and the vertical distribution of buoyancy before and during the RI phase. (a–c) The state before RI, and (d) the state during RI. In the figure, the contour lines represent the potential, with the color-filled areas representing the buoyancy (units: m^2/s^2). The black horizontal lines mark the 200 hPa height level, and the dashed lines are the cyclone's center reference line.

Therefore, while changes in strong VWS did not directly increase APE through baroclinic canonical transfer via the thermal wind relationship, they significantly contributed to convection development and the maintenance of the cyclone's upright structure, so as to enhance the buoyancy conversion.

5. Conclusions

Traditional theories suggest that strong vertical wind shear (VWS) typically suppresses the structure of tropical cyclones, and thereby inhibits their intensification. This, however, is challenged by the offshore rapid intensification (RI) of Typhoon Rammasun in 2014, which has a strong VWS. Ever since this observation, many studies have been conducted, attempting to explain such a counterintuitive phenomenon. In all these studies, a possibility has been overlooked; that is, strong VWS might directly contribute to RI through baroclinic instability. Early work establishes that the interaction between temperature gradients and VWS is a critical driver of atmospheric instability [46,47]. VWS can modify the wind speed and direction at different atmospheric levels, increasing the efficiency of energy conversion in baroclinic systems, and facilitating the conversion of potential energy into kinetic energy, thus driving mesoscale weather systems, like cyclones or convective systems.

Motivated by this, we apply the functional analysis tool, multiscale window transform (MWT), and MWT-based energetics analysis, to reexamine Rammasun, and provide a new perspective on its RI. Our study reveals that, due to the change in the direction of the strong VWS of the background field and the typhoon's unique location, the strong VWS not only

failed to suppress Rammasun's RI, but, contrary to conventional understanding, facilitated its development through processes such as baroclinic canonical transfer. This is revealed in the pathways of energy on the typhoon scale, which is diagnosed using the multiscale window transform (MWT) and MWT-based energetics analysis [37]. More specifically, the mechanisms through which the strong VWS facilitated the designated RI of Rammasun can be summarized as the following two steps:

(1) Baroclinic energy transfer and warm-core enhancement. Under normal circumstances, VWS tilts a TC's vortex, causing asymmetries that suppress intensification. However, in the Rammasun case, strong VWS induced baroclinic instability, which triggered an efficient transfer of available potential energy (APE) from the large-scale background flow to the typhoon-scale circulation. This process fueled the middle to upper troposphere, strengthening the storm's warm core and enhancing upper-level divergence. As a result, convection became more concentrated, sustaining a more organized and intense eyewall structure, which is critical for RI.

(2) VWS directional shift and improved structural alignment. While the magnitude of VWS remained strong, its direction gradually shifted from an east–west to a north–south orientation during the RI period. This change reduced the destructive vortex tilt that would typically weaken a TC. Instead, it allowed the storm to regain a more vertical alignment, improving the efficiency of buoyancy conversion and pressure gradient force transfer. With a more symmetric structure, kinetic energy (KE) was effectively transported from the upper to the lower levels, leading to an acceleration of near-surface wind speeds and further intensification.

These findings demonstrate that VWS does not always inhibit RI; under certain conditions, it can enhance it by facilitating energy transfer and structural reorganization. This challenges the conventional assumption that strong VWS is universally detrimental to TC intensification and suggests that the interaction between VWS and baroclinic processes plays a more nuanced role than previously understood.

The multiscale energetics diagnosis, nonetheless, does not exclude the traditional viewpoint that latent heat release remains the most important energy source during RI. The role of baroclinic instability is complementary, though important. We just emphasize its importance for this special case, which is quite counterintuitive, and hope the finding may enrich our understanding of the mechanisms behind the RIs in tropical cyclones.

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Abbreviations

The following abbreviations are used in this manuscript:

TC	Tropical cyclone
VWS	Vertical wind shear
RI	Rapid intensification
MWT	Multiscale window transform
APE	Available potential energy
KE	Kinetic energy
SST	Sea surface temperature
MPI	Maximum potential intensity
TCHP	Tropical cyclone heat potential
RWM	Radius of maximum winds
IBTrACS	International Best Track Archive for Climate Stewardship
CMA	China Meteorological Administration

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