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**Sensitivity of Cross-equatorial Propagation of Cold Surge to Sea Surface Temperature  
in the South China Sea**

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**ABSTRACT**

East Asian cold surges occasionally propagate deep into the tropics and develop into cross-equatorial northerly surges (CENS) over the western Maritime Continent. Previous studies showed close association between the cross-equatorial propagation of cold surge and cold sea surface temperatures (SSTs) in the South China Sea (SCS). However, it remains unclear whether these cold SSTs are merely a response to cold surges or actively influence their propagation and evolution into CENS. This study addresses this question using numerical

experiments on several cold surges under cooler and warmer SST conditions. Our results show that decreasing (increasing) the preexisting SCS SST strengthens (weakens) northerlies south of the equator during cold surges. The atmospheric response is concentrated primarily below 1 km. SCS cooling induces high-pressure anomalies that enhance the pressure gradient force, favoring southward flow into the Southern Hemisphere, whereas warming produces the opposite response. The southern SCS plays the primary role in modulating the winds compared with adjacent seas. A 1°C SST cooling (warming) over the southern SCS strengthens (weakens) CENS winds by approximately 1.5 m/s. The resulting changes in cross-equatorial flow consequently alter rainfall responses to the surges. Cooler SSTs enhance southward moisture transport, increasing rainfall over northern Java while causing moisture divergence and drier conditions over the Karimata Strait. These findings demonstrate a direct influence of SCS SST on CENS evolution and highlight the importance of accurately representing oceanic conditions in numerical models to improve forecasts of CS-related rainfall.

## **1. Introduction**

A cold air outbreak from Siberia, widely known as an East Asian cold surge (CS) (Chang et al. 1979; Zhang et al. 1997), is a major synoptic phenomenon during the boreal winter that transports substantial moisture toward the tropics (Chang and Lau 1980; Johnson and Zimmerman 1986; Compo et al. 1999), causing heavy rainfall across the western Maritime Continent (Johnson and Priegnitz 1981; Chang et al. 2005; Lim et al. 2017; Narulita et al. 2023). The extratropical outbreaks of CS propagate to different pathways in mid and low latitudes (Abdillah et al. 2021). In the Maritime Continent, CS often propagates to the South China Sea (SCS), building up the climatology of tropical cold air in the western Maritime Continent seas (Tita et al. 2025). While entering the SCS, CS can further be divided into two types (Purwadani et al. 2026). The first type crosses the equator, triggering a cross-equatorial

northerly surge (CENS; hereafter CS-CENS) and bringing rainfall to northern Java (Suppiah and Wu 1998; Hattori et al. 2011; Yulihastin et al. 2020; Maulana et al. 2023; Hadi et al. 2025). The second type fails to cross the equator (CS-noCENS), instead triggering rainfall over Sumatra and the Malay Peninsula (Tan et al. 2023). These pathways are controlled by tropical variability, particularly pressure anomalies south of the Maritime Continent, where a low-pressure system supports cross-equatorial flow while a high-pressure system inhibits it (Hadi et al. 2025; Purwadani et al. 2026). Alongside these pressure anomalies, the sea surface temperature (SST) in the South China Sea (SCS) is hypothesized to be another crucial modulator (Yulihastin et al. 2020). While previous studies have found that during boreal winter, SCS SSTs influence rainfall and wind patterns over the western Maritime Continent on a seasonal timescale (Varikoden et al. 2010; Koseki et al. 2013), their direct causal impact on CS winds on a synoptic timescale remains poorly understood.

The SCS experiences cooling during the boreal winter monsoon, forming a region of relatively cold SST within the Indo-Pacific warm pool known as the cold tongue (Liu et al. 2004; Thompson et al. 2016). During this season, the variability of the SCS SST is primarily driven by changes in wind-related latent heat fluxes, shortwave radiation fluxes, and ocean horizontal advection (Thompson et al. 2016; Qin et al. 2025). These forcings are modulated by multi-scale phenomena, ranging from large-scale drivers like El Niño-Southern Oscillation and the East Asian winter monsoon (Wang and Wu 2021), to intraseasonal and synoptic phenomena such as the Madden-Julian Oscillation (Qin et al. 2025) and CS (Koseki et al. 2023). Furthermore, as a highly eddy-active region, the frequent occurrence of oceanic mesoscale eddies within the SCS also contributes to localized SST anomalies (Chen et al. 2011).

Climatologically, boreal winter seasons with a strong cold tongue are associated with a strengthening of the cross-equatorial flow through the Karimata Strait (Koseki et al. 2013). Numerical simulations by Koseki et al. (2013) suggest that cold SCS SSTs suppress local convection, thereby enhancing monsoon penetration toward northern Java. However, another previous study argued that Koseki's SST settings, specifically by applying October SSTs to November–March simulations, likely exaggerated the impact of the SCS SST (Seow et al. 2023). While Seow et al. (2023) concluded that the SCS SST does not significantly affect circulation across the Maritime Continent, their simulations failed to capture the enhanced monsoon penetration toward northern Java during strong cold tongue events. This inconsistency likely arises from their use of daily climatology data for lateral boundary conditions, which neglects the day-to-day variability of wind and SST fields over the SCS.

On a sub-seasonal scale, the SCS exhibits a distinct coupled relationship between the ocean and atmosphere, characterized by the southward propagation of SST anomalies, surface winds, and latent heat fluxes (Wu 2016). This interaction is also visible in how CS events interact with the SCS SST, with the CS driving southward expansion of oceanic cooling through strong surface winds (Koseki et al. 2023), especially CS with long duration (Yang et al. 2026). However, while the mechanism of wind-driven cooling is well-documented, the feedback effect of these cold SST anomalies on the CS winds remains poorly understood. Theoretically, SST can affect wind by modifying hydrostatic surface pressure (Lindzen and Nigam 1987) and/or altering the vertical wind profile through changes in atmospheric stability and vertical mixing within the planetary boundary layer (Wallace et al. 1989). Recent studies demonstrate that mesoscale SST anomalies in the SCS affect surface winds, where the vertical mixing mechanism dominates directly over the location of SST anomalies while pressure gradient and horizontal advection govern the wind response at the edge of SST anomalies (Liu et al. 2018).

Through these mechanisms, cold SSTs in the SCS can increase the high pressure north of the equator, generating a meridional pressure gradient that strengthens the cross-equatorial flow, while simultaneously stabilizing the lower atmosphere over the SCS, maintaining the wind intensity before it crosses the equator. Therefore, this study hypothesizes that during a CS event, cold SCS SSTs strengthen the cross-equatorial flow, whereas warm SSTs weaken it. To test this hypothesis, numerical simulations are utilized to isolate the oceanic feedback, specifically identifying how the SCS SST modulates the cross-equatorial propagation of the CS.

Thus, the present study investigates how SCS SST modulates the cross-equatorial flow of East Asian CS, thereby filling the knowledge gap regarding SCS SST and wind interactions on daily timescales. We conduct real-case numerical simulations focusing on CS-noCENS and CS-CENS events. To isolate the SST impact, sensitivity experiments are also carried out by replacing the real SST with a constant DJF climatological value. This approach allows us to clearly identify how SST conditions modulate the CS wind flow across the Maritime Continent. The remainder of this paper is structured as follows: Section 2 describes the data and methodology used, Section 3 presents the results and discussion, and Section 4 summarizes the important findings of this study and notes potential future studies.

## **2. Data and methods**

### **2.1. Case study selection**

This study selected six CS events from the CS events dataset previously identified by Purwadani et al. (2026) to investigate how SST conditions modulate cross-equatorial flow. These cases include three CS-noCENS events associated with warm SST anomalies and three CS-CENS events associated with cold SST anomalies, as detailed in Table 1. The combination

is justified by composite patterns showing that CS-CENS events are typically preceded by cold SST anomalies in the SCS, while CS-noCENS events are generally preceded by warm SST anomalies (Purwadani et al. 2026). Cases were specifically chosen based on two general criteria, the persistent presence of pre-conditioning SST anomalies and clear extratropical propagation signals of CS. For CS-CENS events, an additional criterion was applied, ensuring that the CENS peak does not precede the CS peak.

**Table 1.** List of CS event utilized in this study.

| No | Case      | Peak of CS |
|----|-----------|------------|
| 1  | CS-WARM-1 | 2006-12-21 |
| 2  | CS-WARM-2 | 2014-12-14 |
| 3  | CS-WARM-3 | 2016-02-07 |
| 4  | CS-COLD-1 | 2004-02-13 |
| 5  | CS-COLD-2 | 2005-01-15 |
| 6  | CS-COLD-3 | 2021-01-18 |

As illustrated in the spatial patterns in Figs. 1d-i, each case exhibits distinct atmospheric and oceanic anomalies. In the warm cases (CS-WARM), positive SST anomalies are observed across the SCS from day -7 to day 0, while the cold cases (CS-COLD) exhibit negative SST anomalies during the same period. All six cases are characterized by high mean sea level pressure (MSLP) over the Asian continent and strong low-level northerly wind across SCS, indicating typical features of a CS event (Supplementary Information Fig. S1). Simulations for each case were performed for a 16-day period, from day -8 to day +7 relative to the peak, including a 1-day spin-up period. This timeframe was chosen because the CENS typically begins to strengthen sharply eight days before the CS peak (Supplementary Information Fig. S2).

## 2.2. Model configuration

Numerical simulations in this study were conducted using the Weather Research and Forecasting (WRF) version 4.2 model (Skamarock et al. 2019). The model is configured with a single domain at a 9 km horizontal resolution with 40 vertical layers, covering the region of 20°N - 15°S and 95°E - 125°E as shown in Figure 1a. The atmospheric initial and lateral boundary conditions are provided by the ERA-5 hourly reanalysis data (Hersbach et al. 2020), while SST inputs are derived from the OISST v2.1 daily dataset (Huang et al. 2021), both with a 0.25° horizontal resolution. The physical parameterizations selected for the simulations include the Betts-Miller-Janjic (BMJ) cumulus scheme, WDM5 microphysics, RRTMG scheme for both shortwave and longwave radiation, and Noah Land Surface Model for land surface processes. This combination of parameterization schemes has been proven to perform well in simulating precipitation over the Maritime Continent (Fonseca et al. 2015), particularly during CS and CENS events (Yulihastin et al. 2025).

A sensitivity evaluation of the planetary boundary layer and surface layer schemes was performed, as the accurate representation of air-sea interactions is fundamentally linked to the simulation of boundary layer processes that govern the exchange of energy and momentum between the ocean surface and the atmosphere. This evaluation compared the YSU, QNSE, and MYNN schemes using two representatives from the newest selected cases, namely CS-WARM-3 and CS-COLD-3, selected based on the availability of radiosonde data from day -7 to day +7. According to the evaluation against radiosonde observations in the 1000–850 hPa layer, the QNSE scheme was selected for all simulations (Supplementary Information Table S1). Furthermore, GPM IMERG V07 data (Huffman et al. 2023) was used to evaluate the model's performance in capturing the spatial rainfall patterns.

### 2.3. Experiment using SST modification

To identify the impact of specific SST anomalies on CS winds, we performed sensitivity experiments by modifying the SST within the model. For each case, two simulations were conducted: a control (CTR) run using the daily-updated real-case SST, and a climatology (CLM) run. In the CLM run, the SST over a designated region extending from 20°N to 8°S and 103°E to 121°E (Fig. 1b) was replaced with a constant DJF (December–February) long-term climatological mean (1981/1982–2020/2021). This approach means that the resulting SST differences represent not only the specific event's anomalies but also include a seasonal deviation (e.g., December SSTs are climatologically warmer than the DJF mean). By utilizing a constant climatological mean over this region, the simulations ensure that the SST inputs remain within a realistic seasonal range. This replacement effectively removes the specific warm or cold SST anomalies of the case, revealing how the circulation responds when these oceanic forcing signals are absent.

The geographical extent of the SST modification, which expands beyond the SCS to include the Java Sea, is justified by the broad spatial differences in SST anomalies between CS-CENS and CS-noCENS events (Fig. 1b). In CS-noCENS cases, warm anomalies often cover the entire SCS down to the Java Sea. Conversely, CS-CENS events are associated with cold anomalies that expand and intensify southward as they are driven by the strong CS winds. Therefore, this broad modification domain is designed to comprehensively capture the oceanic influence around the CENS formation region. Nevertheless, to determine which specific sub-region exerts the most dominant influence, further sensitivity experiments dividing this modification area are discussed in Section 3.3. Finally, while the WRF model outputs were generated hourly,

a 24-hour running mean was applied to the data for all subsequent analyses to remove diurnal fluctuations.

### **3. Results and discussion**

#### **3.1. CENS response to changes in SST boundary conditions**

The impact of SST anomalies on the intensity and vertical structure of the CENS is analyzed by comparing the CTR and CLM simulations. The CENS index is calculated using the 10-m meridional wind over the CENS region ( $0^{\circ}$ – $5^{\circ}$ S,  $105^{\circ}$ E– $110^{\circ}$ E), as shown in Figure 1a. In the CS-WARM simulations (Fig. 2a), the CENS index produced by the CLM run is generally more negative (stronger northerly flow) than that of the CTR run. This result suggests that the presence of warm SST anomalies in the SCS during these cases acts to weaken the CENS flow. Thus, removing these warm anomalies in the CLM run allows the CENS to strengthen. Conversely, for CS-COLD events, the CLM run produces a weaker CENS index compared to the CTR run (Fig. 2b). This indicates that the cold SST anomalies present in the real cases actively modulate the strength of the CENS winds. Then, once these cold anomalies are replaced by warmer climatological values, the CENS intensity decreases.

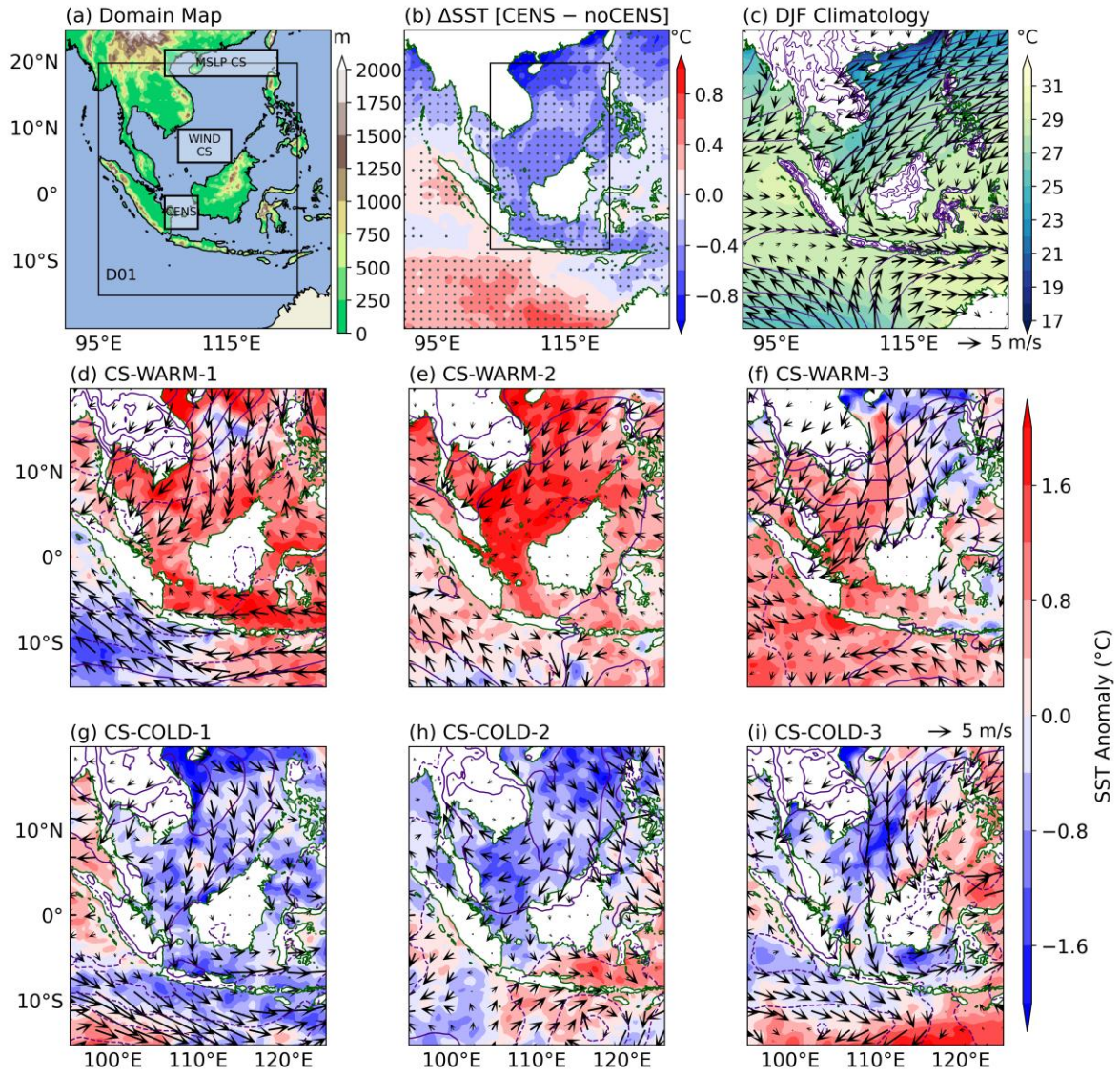
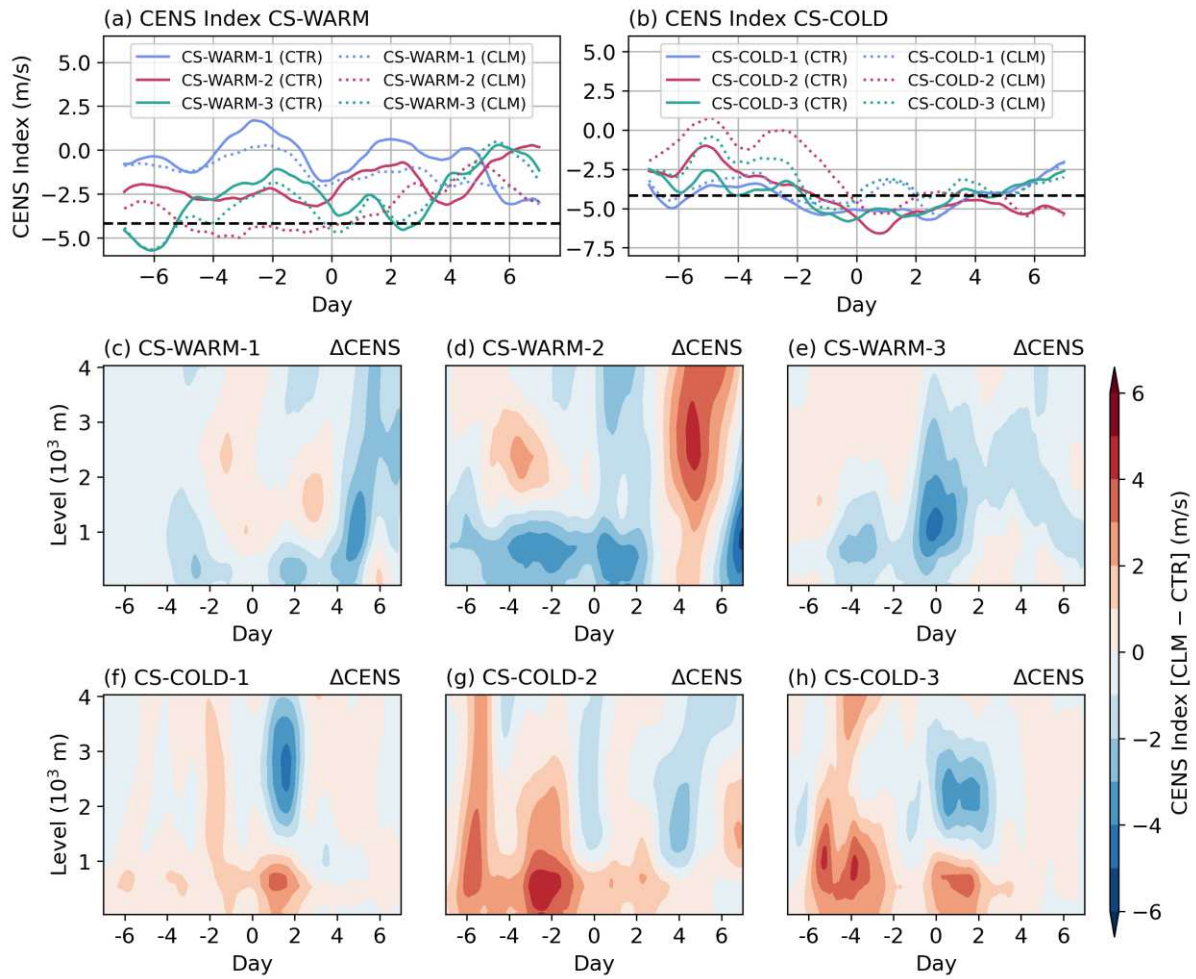


**Fig. 1.** (a) WRF model single domain (D01) with topographic height (m, shading), showing the regions of the CS and CENS indices for CS event classification from Purwadani et al. (2026). (b) Mean SST differences (°C, shading) between CS-CENS and CS-noCENS averaged from day -7 to day 0 of the CS peak. Black boxes indicate the region used for SST modification. (c) DJF climatological mean of SST (shading), MSLP (contours), and 10 m winds (vectors). (d–f) Spatial anomalies of warm cases (CS-WARM 1–3), where SST (shading) are averaged from day -7 to day 0, while MSLP (contours) and 10 m winds (vectors) are plotted at the day 0. (g–i) Same as (d–f), but for cold cases (CS-COLD 1–3). All SST fields (b–i) are derived from OISST, while MSLP and wind fields are from ERA5.



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**Fig. 2.** Temporal evolution of the CENS index for the (a) CS-WARM and (b) CS-COLD cases from the CTR and CLM simulations. (c-h) Vertical profile differences between the CLM and CTR runs in meridional wind speed over the CENS region from day -7 to day +7 relative to the CS peak for each case study.

Northerly winds in the CENS region during CS events typically reach 2000 meters or approximately 800 hPa (Supplementary Information Fig. S3), consistent with the typical vertical profile of CS in the SCS (Compo et al. 1999; Wang et al. 2024) and CENS phenomena (Moteki 2024). However, the strengthening of CENS winds induced by SST cooling and their weakening by SST warming are predominantly concentrated in the lower troposphere below

1000 meters. Individual cases exhibit unique vertical characteristics and temporal peaks. For instance, CS-WARM-3 shows a peak intensification at higher altitudes between 1000 and 2000 meters. Furthermore, while most cases maintain their response through day +2, the intensification in CS-WARM-3 is short-lived and peaks at day 0. Within the cold cases, CS-COLD-2 displays a relatively weak signal during the day 0 to day +2 interval, with its strongest wind reduction occurring earlier at day -3. Additionally, reversed wind anomaly patterns are occasionally observed in the upper levels for cases such as CS-WARM-1, CS-COLD-1, and CS-COLD-3. Based on the temporal evolution observed in Figure 2, the influence of SST modification on CENS winds is most pronounced during two specific intervals. The pre-peak period, from day -5 to day -1, represents the background influence where SST modulates the environment before the CS peak. The peak period, from day 0 to day +2, represents the direct impact as the CS propagates across the Maritime Continent.

To examine the spatial pattern of the oceanic and atmospheric changes associated with the strengthening or weakening of CENS, we analyze the differences in SST, MSLP, and wind fields between the CLM and CTR runs. The negative SST differences in the CLM-minus-CTR outputs in the CS-WARM cases represent the neutralization of positive SST anomalies. During the pre-peak period, the cooling signal extends from the northern SCS to the Java Sea, reflecting the widespread warm anomalies present in the CTR runs (Fig. 3a). This SST change is associated with an increase in MSLP over the SCS, which enhances the northerly flow in the western Maritime Continent before the CS reaches the tropics. By the peak period, the negative SST signal becomes concentrated in the southern SCS (Fig. 3b). This southward shift reflects the typical thermal response of the SCS to CS where strong northerly winds progressively cool the sea surface from north to south (Yang et al. 2026), leaving the remaining warm anomalies localized in the southern region during the peak phase. Although our prescribed SST setup does

not explicitly simulate the ocean's dynamic response to the atmosphere, this wind-driven cooling process is inherently captured by the real-case daily SST evolution used in the CTR simulation. Consequently, the removal of these southern SCS warm anomalies strengthens the MSLP in that region and drives a more intense cross-equatorial flow from the CS as it reaches the tropics (Fig 3c,d).

In CS-COLD cases, the spatial difference shows a warming signal as the CLM runs neutralize the cold SST anomalies in the CTR runs. During the pre-peak period, a widespread warming signal is observed across the SCS, indicating that strong cold anomalies were already well-established in the CTR simulations as a pre-conditioning state (Fig. 4a). This SST warming is associated with a decrease in MSLP, which reduces the meridional pressure gradient and weakens the northerly background flow before the surge reaches the tropics (Fig. 4c). By the peak period, the warming signal in CS-COLD-1 and CS-COLD-3 strengthens and shifts southward (Fig. 4b). This occurs because the intensifying CS winds create stronger cold SST anomalies in the CTR runs, which are then neutralized in the CLM experiments. However, CS-COLD-2 exhibits a unique evolution where the warming signal in the southern SCS is less intense during the peak period compared to its pre-peak phase. Such variability may reflect complex oceanic dynamics, though those processes are outside our scope, which only focuses on atmospheric responses to SST.

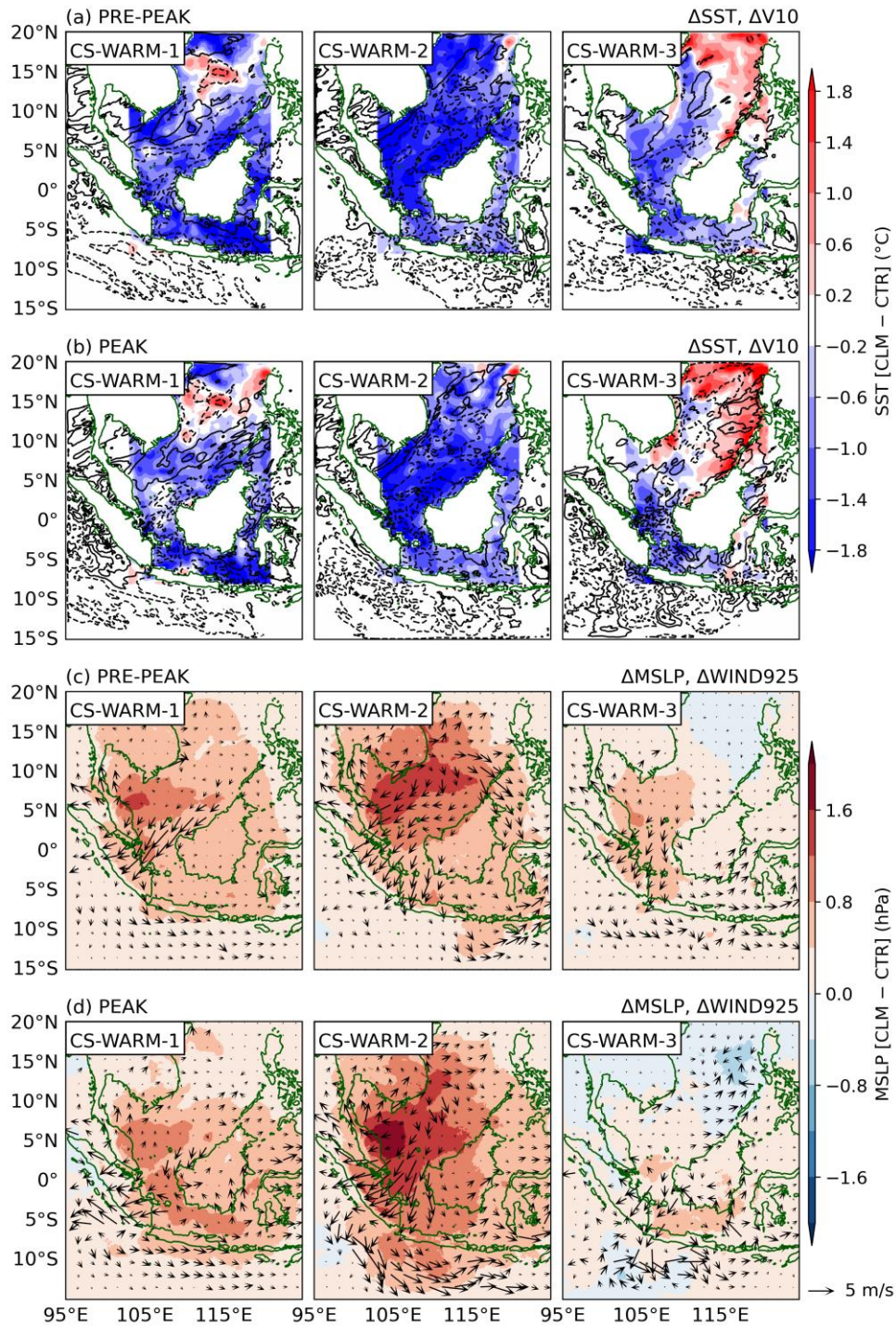


**Fig. 3.** (a, b) Spatial distribution of differences between the CLM and CTR simulations in SST (shaded) and 10-m meridional winds (contours), alongside (c, d) differences in MSLP (shaded) and 925 hPa winds (vectors) for the CS-WARM cases during the pre-peak and peak periods.

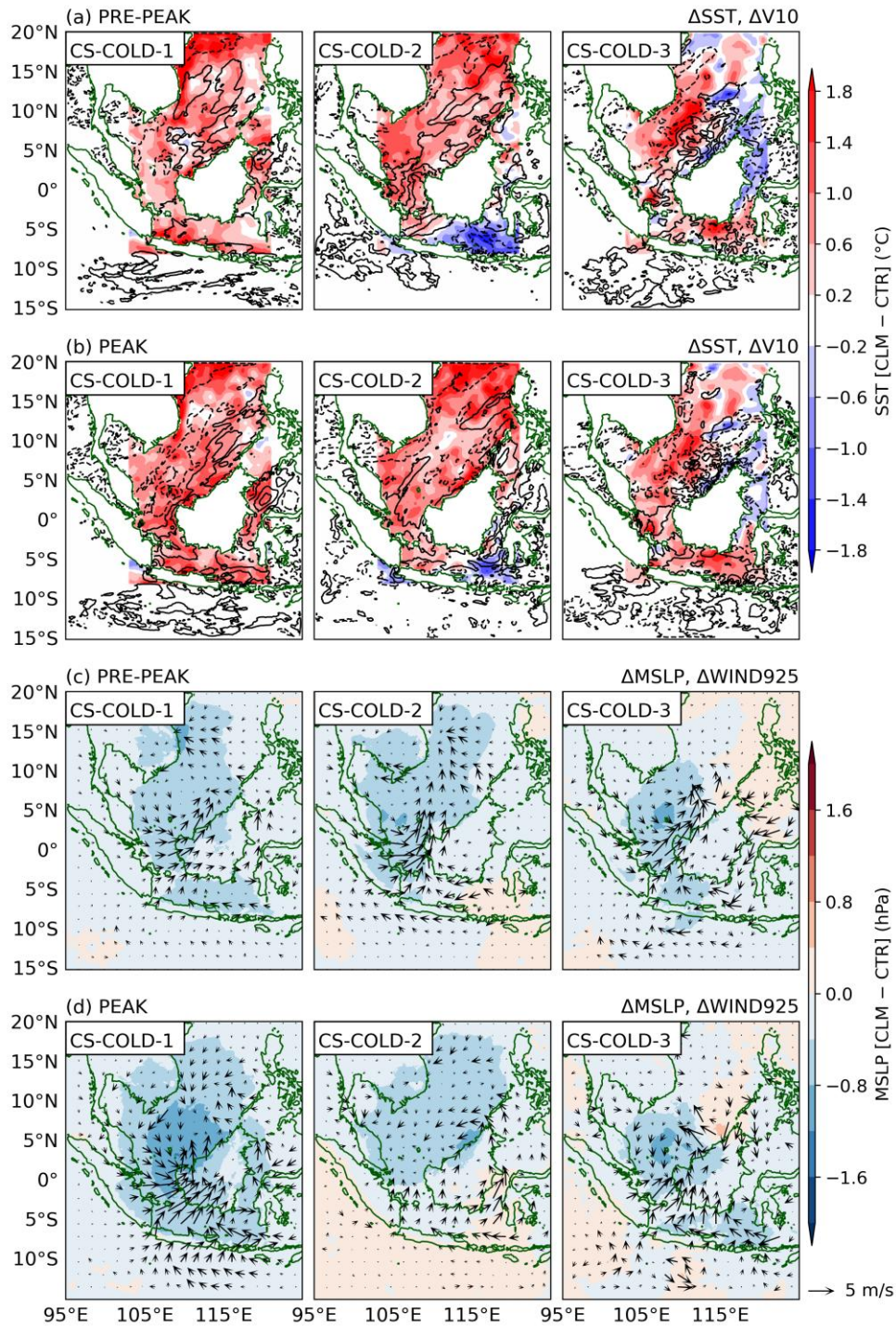


**Fig. 4.** Same with Fig. 3 but for CS-COLD cases.

To further examine the temporal and latitudinal evolution of these changes, we analyzed the Hovmöller diagrams of the CLM-minus-CTR differences (Fig. 5). The SST differences (cooling in CS-WARM and warming in CS-COLD) are evident from day -7 and show a

southward expansion signal, although the signal is somewhat mixed and potentially influenced by local oceanic dynamics in each case. The MSLP changes follow a similar pattern to the SST, where cooling (warming) causes an increase (decrease) in pressure with similar timing and location. While the timing of the meridional wind response aligns with the MSLP changes, the peak wind anomalies are located south of the MSLP centers. This latitudinal shift occurs because the wind response is likely driven by the meridional pressure gradient. Consequently, the strongest meridional wind changes are located on the southern flank of the MSLP centers, where the pressure slope is steepest.

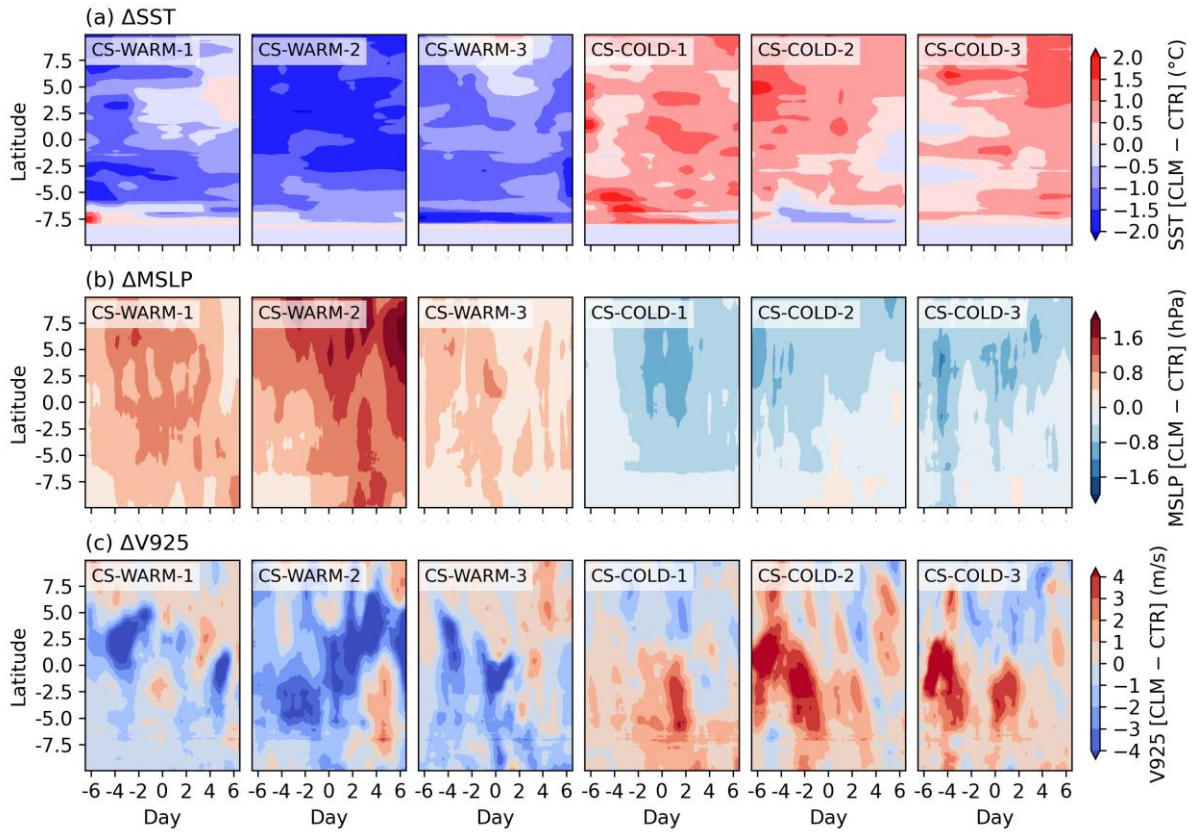


**Fig. 5.** Hovmöller diagrams averaged between  $105^{\circ}\text{E}$  and  $110^{\circ}\text{E}$  showing the temporal evolution of differences between the CLM and CTR simulations in (a) SST, (b) MSLP, and (c) 925 hPa meridional wind speed from day -7 to day +7 of the CS peak for the six case studies.

Overall, despite spatial and temporal variations in SST anomaly intensity, the atmospheric response remains consistent. Specifically, the location and magnitude of the MSLP changes correspond to the location and intensity of the SST changes, a stronger SST modification leads to a more pronounced atmospheric response. This explains the relatively weak CENS index reduction in CS-COLD-2 during the peak period (Fig. 2g), which is consistent with the less intense SST and MSLP changes in that case. These MSLP changes subsequently trigger wind responses south of the MSLP anomaly centers, indicating that the strengthening or weakening of the CENS winds is linked to the SST and MSLP conditions to its north over the SCS. This also indicates that the wind variations over the CENS region are likely driven by the underlying MSLP gradients, which are typically strongest around the edge of the SST anomalies (Lindzen and Nigam 1987). The detailed physical mechanisms linking these SST variations to the CENS wind response are examined in the next section.

### **3.2. Mechanisms of SST influence on the CS-to-CENS transition**

The ocean affects the atmosphere through surface heat transfer, which links SST conditions to atmospheric processes. The spatial-temporal pattern of the combined latent and sensible heat flux anomalies (CLM minus CTR) generally mirrors the underlying SST anomaly pattern (Fig. 6 and Fig. 5a). Negative values indicate a reduction in ocean-to-atmosphere heat transfer, while positive values indicate an enhancement. Areas with colder SST signals typically suppress surface heat fluxes, while warmer signals enhance them. However, a distinct reversal occurs around latitude 5°N–2°N in CS-WARM-1 (Fig. 6a). In this region, the magnitude of the negative SST difference is weak (Fig. 5a), allowing the mechanical effect of wind strengthening (Fig. 2c) to override the weak SST thermodynamic effect. The enhanced wind speed drives stronger upward evaporation, making the heat flux difference positive despite the slightly cooler SST. A similar mechanism explains the opposite pattern in CS-COLD-3 around

3°N–0°N (Fig. 6f). Because the positive SST difference in this zone is weak, the reduction in wind speed suppresses mechanical evaporation, resulting in a negative heat flux difference despite the warmer ocean surface.

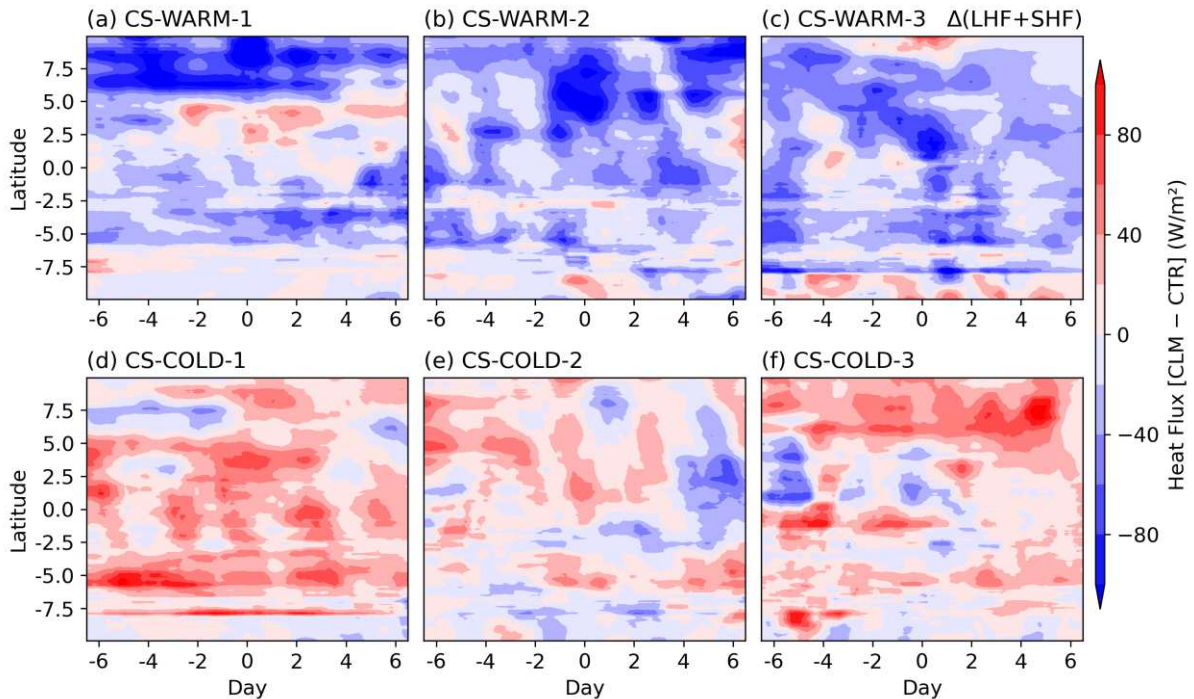


**Fig. 6.** Hovmöller diagrams averaged between 105°E and 110°E showing the evolution of differences between the CLM and CTR simulations in total surface heat flux from day -7 to day +7 of the CS peak. Negative values indicate reduced ocean-to-atmosphere heat transfer, whereas positive values represent enhanced transfer.

The differences in surface heat flux affect the vertical profile of virtual potential temperature ( $\theta_v$ ), pressure, and meridional-vertical circulation ( $v$ ,  $w$ ), as shown in Figure 7. The cross-sections indicate that the source of cooling or warming in the atmosphere originates from the lower levels, which are directly affected by the underlying SST conditions. Furthermore, the pressure contours directly mirror the underlying  $\theta_v$  differences, confirming that these temperature changes effectively modify air density and pressure. In the CS-WARM cases, a

clear negative  $\theta_v$  difference occurs at the lower levels. During the pre-peak phase (Fig. 7a), this cold anomaly can extend southward near the surface until around 6°S, with the strongest cooling centered in the northern region. This condition makes the highest increment of surface pressure located north of the equator, creating a north-to-south pressure gradient that drives the cross-equatorial flow. This structure persists into the peak phase (Fig. 7b), maintaining the north of equator high-pressure anomaly that drives the strong northerly winds in the CENS region. However, a distinct variation occurs during the peak phase of CS-WARM-3. In this case, the lower-level negative  $\theta_v$  difference is very weak and localized because the underlying SST difference is confined only to the southern SCS and Java Sea (Fig. 3b). Interestingly, two high-pressure peaks are formed near 2°N and 4°S. Consequently, the strong northerly wind core in CS-WARM-3 is shifted upward to the 1–2 km level because the lower-level flow is blocked by the localized high-pressure peak near 4°S, which extends up to an altitude of about 1.5 km.

Conversely, the CS-COLD cases exhibit an opposite thermodynamic response. During the pre-peak phase (Fig. 7c), positive  $\theta_v$  differences occur at the lower levels, with the largest differences located north of the equator. This reduces the north-to-south pressure gradient over the CENS region. During the peak phase (Fig. 7d), the vertical structure remains consistent with the pre-peak phase. A unique feature during the peak phase is observed in CS-COLD-2, where a positive  $\theta_v$  difference appears at higher levels around 1.5 km, separated from the surface. This upper-level warming is likely driven by non-local processes, such as atmospheric advection, rather than direct heat transfer from the ocean surface below. Nevertheless, despite this upper-level anomaly, the low-level wind field in CS-COLD-2 still follows the typical boundary layer response to SST warming, maintaining the overall weakening of the cross-equatorial wind.

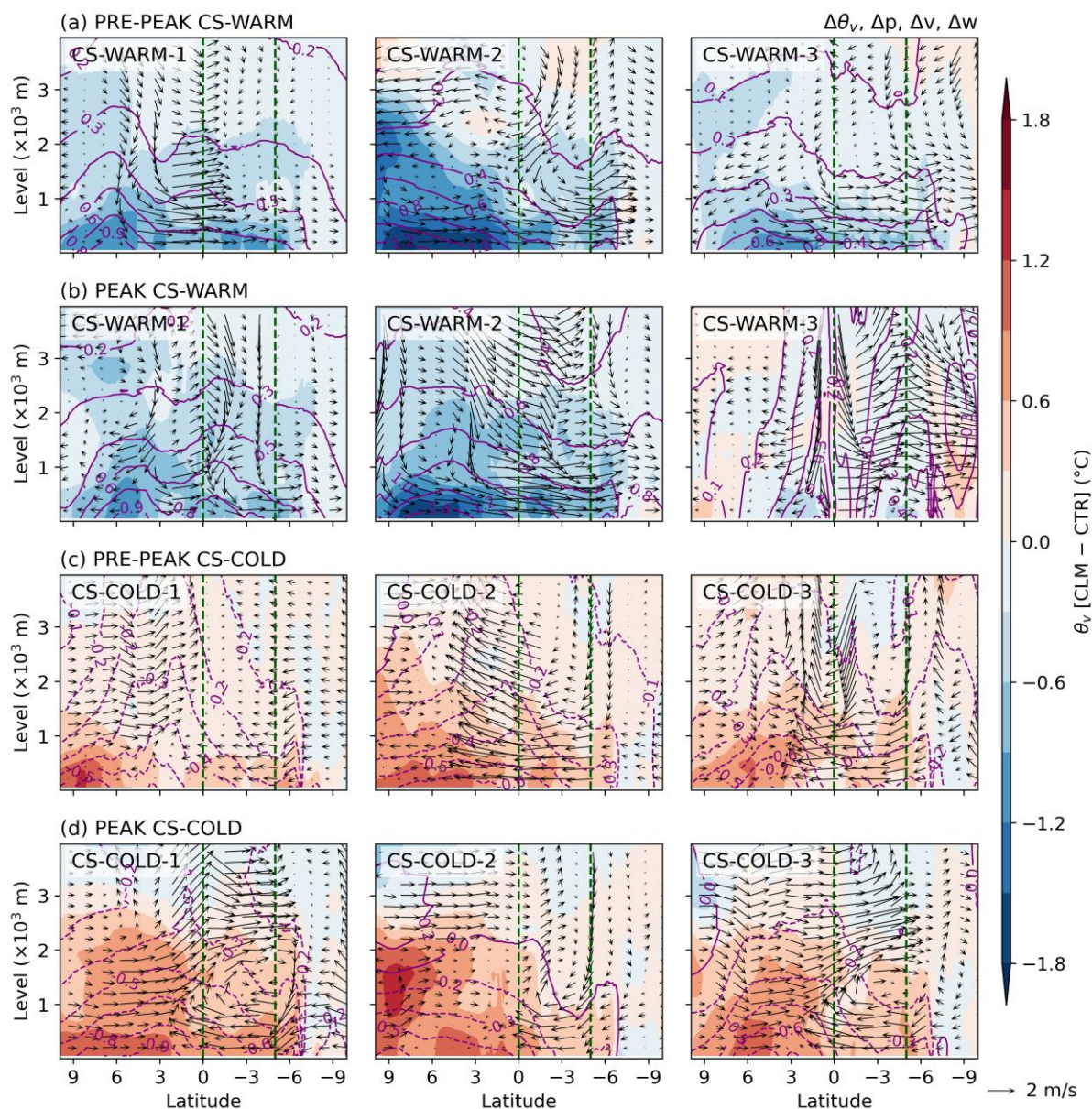


**Fig. 7.** Latitude-height cross-sections averaged between 105°E and 110°E showing the differences between the CLM and CTR simulations in virtual potential temperature (shaded), pressure (contours), and meridional-vertical circulation (vectors) during the pre-peak and peak periods for the (a, b) CS-WARM and (c, d) CS-COLD cases. Vertical wind components are scaled by a factor of 100 to enhance visibility.

Figure 8 and 9 presents the vertical structure of the meridional momentum budget terms averaged over the CENS region (Fig. 1a) to explain the physical drivers of CENS fluctuations. The budget is calculated based on the meridional momentum equation, where the local acceleration ( $\partial v/\partial t$ ) is determined by the sum of horizontal advection (ADV), the Coriolis force (CORI), the pressure gradient force (PGF), and residual terms (RES) (Eq. 1).

$$\frac{\partial \bar{v}}{\partial t} = -\left(\bar{u} \frac{\partial \bar{v}}{\partial x} + \bar{v} \frac{\partial \bar{v}}{\partial y} + \bar{w} \frac{\partial \bar{v}}{\partial z}\right) - f\bar{u} - \frac{1}{\rho} \frac{\partial \bar{p}}{\partial y} - \left(\frac{\partial \bar{u}\bar{v}}{\partial x} + \frac{\partial \bar{v}\bar{v}}{\partial y} + \frac{\partial \bar{w}\bar{v}}{\partial z}\right) \quad (1)$$

To calculate the momentum budget, the WRF output with 40 vertical levels was interpolated every 50 meters. To remove numerical noise resulting from derivative calculations between successive points within the same vertical grid cell, Gaussian smoothing was applied to the vertical profiles of each momentum budget term. In this analysis, negative values indicate an acceleration of northerly winds or CENS strengthening, while positive values indicate a deceleration. For the CS-WARM cases, the plots illustrate the difference between the CLM and CTR simulations, highlighting how the colder climatological SSTs modify the atmospheric response compared to the warmer SSTs in the control cases. Although the temporal pattern of  $\partial v/\partial t$  is fluctuating, the negative peaks occur at vertical locations and times that match the strongest CENS wind enhancement observed in the vertical profiles (Fig. 2). In CS-WARM-1 and CS-WARM-2, the PGF is the primary term supporting the low-level wind acceleration. The ADV term acts as an opposing force to balance the PGF (Liu et al. 2018). The RES term also opposes the wind enhancement, though its influence is limited to the lower levels as it represents the frictional drag force that increases when surface winds get stronger.

CS-WARM-3 shows a unique pattern where the negative  $\partial v/\partial t$  peak is located at an altitude of 1–2.5 km. This matches the vertical profile of CENS enhancement for this case, which

specifically peaks at a higher level compared to other cases. At the lower levels, the PGF becomes positive and inhibits wind acceleration, a result of the localized high-pressure center near 4°S that weakens the surface northerly PGF. This occurs because the cold SST anomaly during the peak phase of this case is narrow and localized around the southern SCS to the Java Sea (Fig. 3b). According to previous studies, around a cold SST anomaly, the advection term accelerates the wind on the upstream side and decelerates it on the downstream side, whereas the PGF acts in the opposite manner (Liu et al. 2018). In this case, the southern SCS acts as an upstream area where the advection term plays a strong role in driving the northerly wind acceleration. Consequently, in CS-WARM-3, the advection term serves as the main driver for strengthening the CENS wind at that altitude.

In the CS-COLD cases, the use of warmer SSTs in the CLM simulation produces a similar response across all events. Although the  $\partial v / \partial t$  pattern remains fluctuating, the timing and location of the wind weakening match the peak CENS weakening in the vertical profiles (Fig. 2). Around day 0, positive changes in  $\partial v / \partial t$  are observed in the near surface level, indicating a weakening of the CENS flow. The momentum budget shows that this northerly CENS wind deceleration is driven by a positive PGF term. The ADV term balances this PGF, particularly at altitudes above 1 km. The RES term shows negative values because the northerly wind weakens, which reduces the frictional drag at the lower levels.

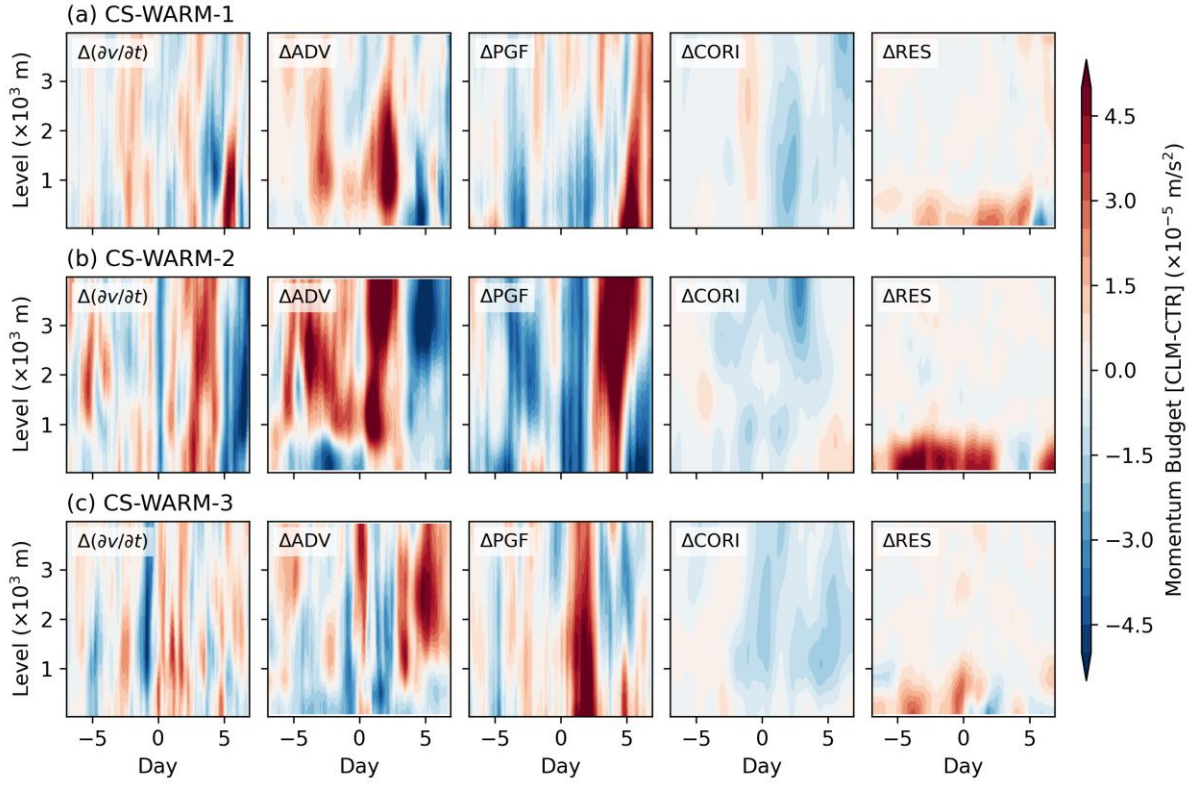


**Fig. 8.** Vertical structure of differences between the CLM and CTR simulations in meridional momentum budget components (local acceleration ( $\partial v/\partial t$ ), horizontal advection (ADV), pressure gradient force (PGF), Coriolis force (CORI), and residuals (RES)) averaged over the CENS region (Fig. 1a) from day -7 to day +7 of the CS peak for the CS-WARM cases. Negative values denote northerly wind acceleration, while positive values indicate deceleration.

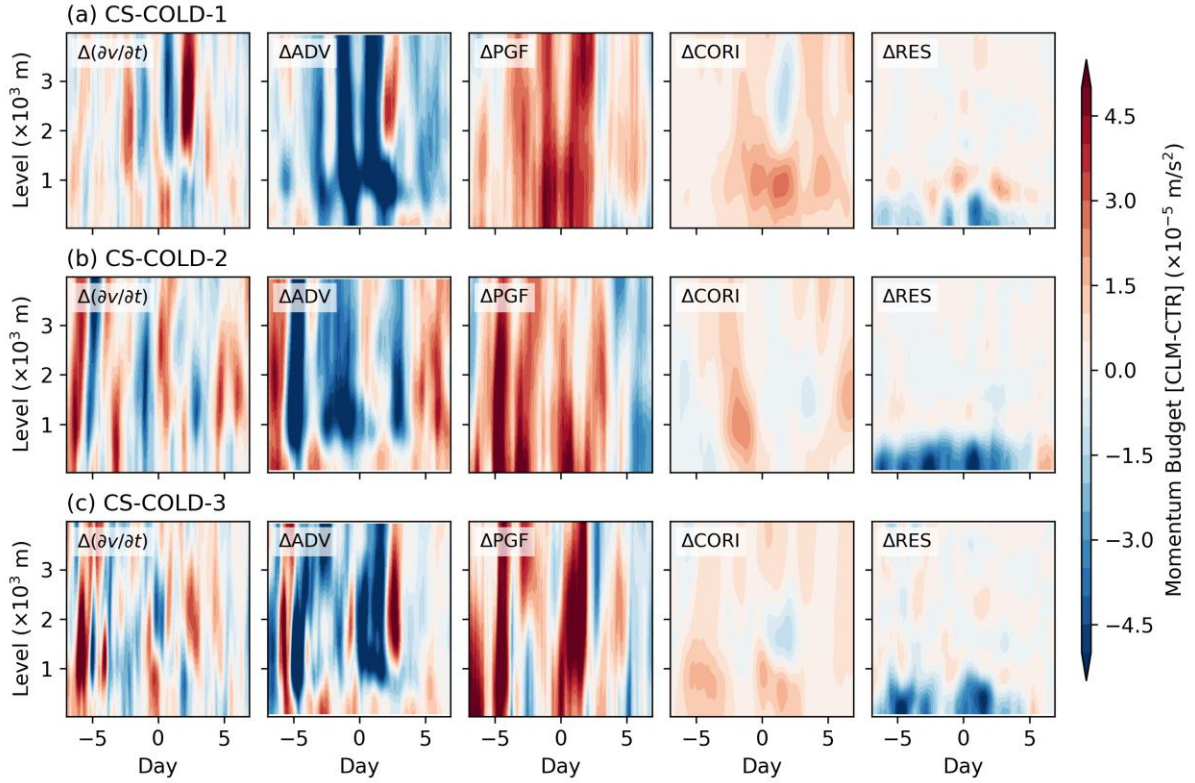


**Fig. 9.** Same with Fig. 8 but for CS-COLD cases.

### 3.3. Sensitivity to specific SST sub-regions and the quantitative impact on CENS

In the previous analyses, the modification of SST to DJF climatological value covered a broad domain, spanning from the northern SCS down to the Java Sea. To determine which specific region exerts the strongest influence on the CENS, sensitivity experiments were conducted by dividing the SST modification domain into three latitudinal sub-regions (using the same longitudinal boundaries as the SST modification experiment defined in Section 2.3): the northern SCS (NSCS; 10°N–20°N), the southern SCS (SSCS; 0°–10°N), and the Karimata Strait to the Java Sea (JVS; 8°S–0°). These regional sensitivity experiments were performed specifically for the CS-WARM-2 and CS-COLD-1 cases. As shown in the SST anomaly maps (Figs. 10a and 10c), these two cases were selected because they exhibit relatively uniform warm or cold SST anomalies extending across the entire domain.

Figure 10b shows the temporal evolution of the CENS index for the CS-WARM-2 case. When the warm SST anomaly is removed in the SSCS region (CLM\_SSCS), the resulting CENS strengthening closely matches the results of the CLM simulation. Conversely, when the SST anomalies are removed only in the NSCS or JVS regions, the CENS index remains similar to the CTR simulation. This clearly indicates that the cold SST conditions in the SSCS are the primary driver for the CENS enhancement. A consistent response is also observed in the CS-COLD-1 case (Fig. 10d). The removal of the cold SST anomaly specifically in the SSCS region produces a weakening of the CENS that closely aligns with the CLM simulation. Therefore, both experiments confirm that the SST variations in the SSCS play the most dominant role in modulating the strength of the CENS.

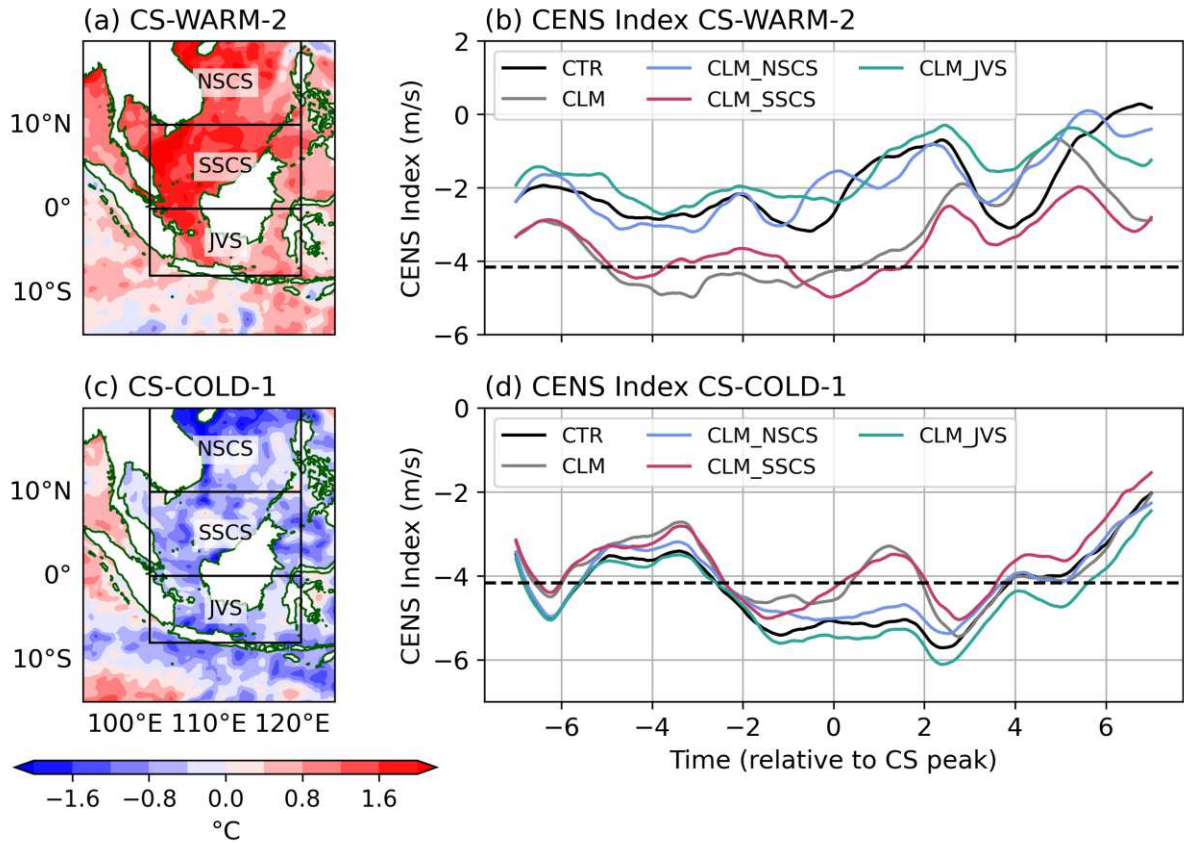


**Fig. 10.** Temporal evolution of the CENS index for the (b) CS-WARM-2 and (d) CS-COLD-1 cases based on sensitivity experiments modifying the SST climatology in three specific sub-regions: the northern SCS (NSCS), southern SCS (SSCS), and Java Sea (JVS), compared against the prior CTR and CLM simulations. Panels (a) and (c) display the SST anomalies for each case and define the sub-regions used for the regional SST modification.

To quantify the impact of SSCS SST on CENS winds, scatter plots are used to illustrate the relationship between changes in SSCS SST ( $\Delta$ SST) and the CENS index ( $\Delta$ CENS), derived from the differences between the CLM and CTR simulations (Fig. 11). A strong positive linear relationship is evident across the peak and pre-peak periods, where the regression slopes physically indicate that a 1°C cooling (warming) of the SSCS SST leads to a 1.52 m/s strengthening (weakening) of the CENS winds during the pre-peak phase, and a 1.54 m/s strengthening (weakening) during the peak phase. It is important to note that because the CENS

flow is characterized by northerly winds (negative meridional winds), a negative  $\Delta\text{CENS}$  value indicates an intensified cross-equatorial flow, whereas a positive  $\Delta\text{CENS}$  indicates its weakening. Consequently, this robust relationship demonstrates that CENS is strongly modulated by SSCS SST conditions, with SST cooling driving the cross-equatorial flow while SST warming suppresses it.

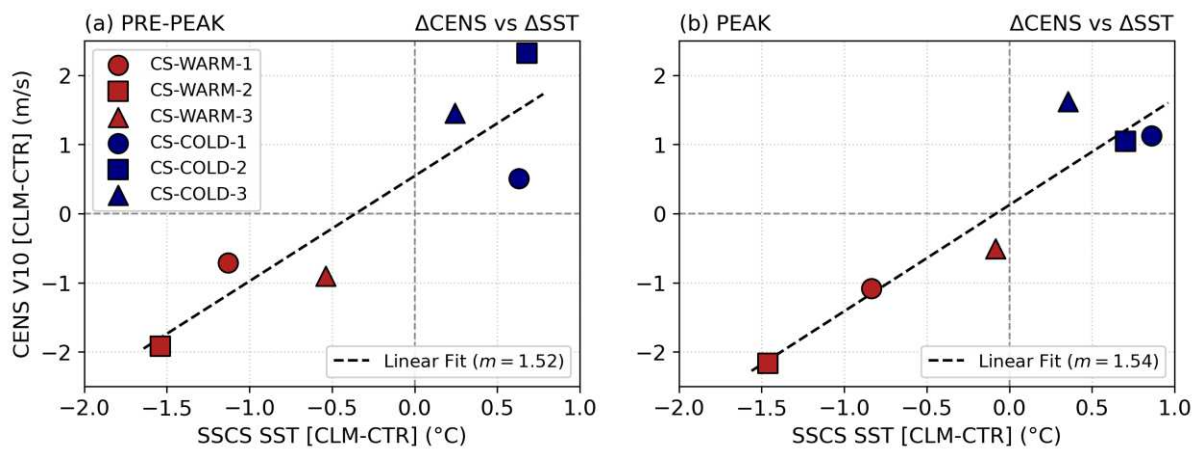


**Fig. 11.** Relationship between the changes in SSCS SST and the CENS index derived from the difference between the CLM and CTR simulations. Values are shown for the (a) pre-peak and (b) peak periods. Red markers represent CS-WARM cases, while blue markers represent CS-COLD cases. The dashed line indicates the linear fit for all plotted cases, with the slope ( $m$ ) provided in the bottom-right corner of each panel.

### 3.4. Impact on the western Maritime Continent rainfall

The rainfall impacts are evaluated on day +1 after the CS peak because the peak of the CENS generally occurs on this day. Based on an evaluation with GPM IMERG data, the WRF CTR simulations reasonably reproduce the spatial rainfall distribution for both CS-WARM and CS-COLD cases, thereby validating the model's performance (Supplementary Information Fig. S4). Fig. 12 illustrates the rainfall differences and the moisture flux differences between the CLM and CTR simulations, highlighting how the rainfall and moisture respond when the event-

specific SST anomalies are replaced with climatological conditions. In the CS-WARM cases, applying the climatological SST removes the warm SST anomaly, effectively cooling the SST. In CS-WARM-1 and CS-WARM-2, this modification produces a dry anomaly around the Karimata Strait and a wet anomaly north of Java Island, which is similar to the rainfall impact of CS-CENS (Hattori et al. 2011; Hadi et al. 2025; Purwadani et al. 2026). This indicates that for CS-noCENS cases in CS-WARM, removing the warm anomaly results in the strengthening of the CENS winds, which enhances the southward moisture transport and produces a pattern similar to a CS-CENS event. Specifically in CS-WARM-3, the removal of the warm anomaly drives a southward shift in the rainfall distribution, marked by a dry anomaly north of the Karimata Strait and a wet anomaly to its south. The removal of the warm SST anomaly induces moisture divergence over the Karimata Strait while strengthening the southward moisture transport from the strait toward the Java Sea.

Conversely, in the CS-COLD cases, applying the climatological SST means removing the cold SST anomaly. This SST warming results in a wet anomaly around the Karimata Strait and northern Borneo. It also creates a dry anomaly north of Java, which is located over the northwestern Java Sea in CS-COLD-1 and CS-COLD-2, and over the northeastern Java Sea in CS-COLD-3. These anomalies are similar to the rainfall impact of CS-noCENS (Tan et al. 2023; Purwadani et al. 2026), indicating that for CS-CENS cases in CS-COLD, removing the cold anomaly results in the weakening of the CENS winds and produces a pattern similar to a CS-noCENS event. The removal of the cold SST anomaly generates moisture convergence around the southern SCS and the Karimata Strait. As a result, the southward moisture transport toward the Java Sea weakens, which subsequently leads to moisture divergence over the Java Sea.

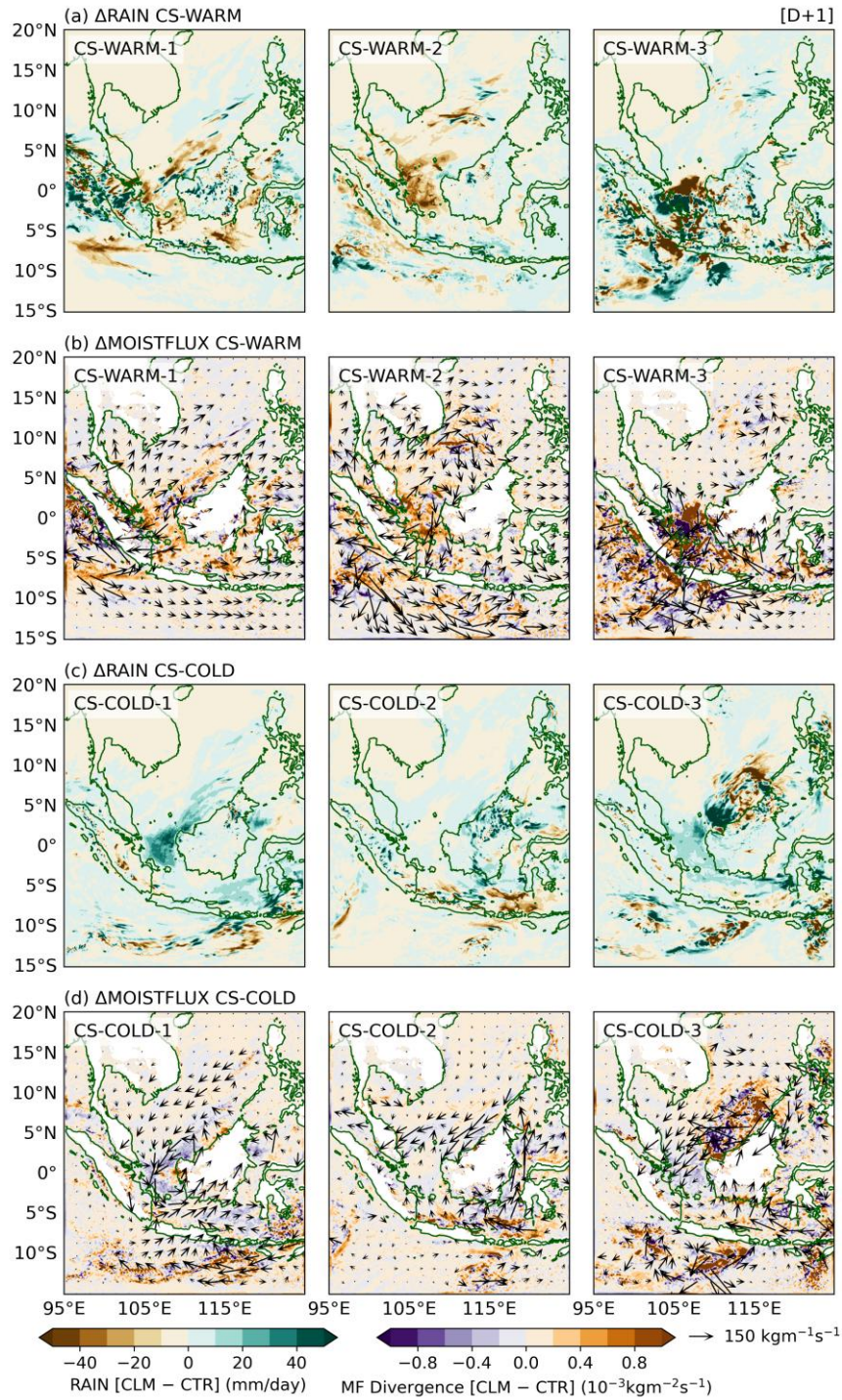


**Fig. 12.** Spatial patterns of differences between the CLM and CTR simulations in (a, c) daily rainfall and (b, d) vertically integrated moisture transport from 1000-700 hPa (vectors) alongside its divergence (shaded) at day +1 following the CS peak across all six case studies.

#### 4. Concluding remarks

This study investigates how sea surface temperature in the South China Sea modulates the cross-equatorial flow during East Asian cold surge events using numerical simulations. The results show that cooling the sea surface temperature in the South China Sea strengthens CENS winds while warming weakens them, which generally occurs in the lower atmosphere below 1000 meters. Although the location and timing of the SST anomalies vary across different events, the atmospheric response remains robust and consistent. Cold SST anomalies induce high-pressure anomalies over the SCS, creating a north-to-south pressure gradient that strengthens the cross-equatorial flow. Conversely, warm SST anomalies reverse this gradient and weaken the CENS wind. CS-WARM-3 exhibits a distinct behavior because its SST anomaly is spatially confined to the southern part of the southern SCS and the Java Sea. The resulting high-pressure system is localized in a way that blocks lower-level wind flow. Consequently, wind strengthening in this specific case occurs at higher altitudes between 1000 and 2000 meters rather than near the surface, and it is driven by the advection term instead of the pressure gradient. This result is consistent with a previous study showing that advection accelerates winds on the upstream side of a cold SST anomaly and decelerates them on the downstream side, while the pressure gradient force behaves in the opposite manner (Liu et al. 2018). Regional sensitivity experiments further confirm that SST variations in the southern SCS exert the most dominant influence on CENS wind modulation compared to the northern SCS or the Java Sea. Quantitatively, a 1°C SST cooling (warming) in the southern SCS strengthens (weakens) CENS winds by approximately 1.5 m/s.

The modulation of CENS by SCS SST directly influences the regional rainfall over the western Maritime Continent. When cold SSTs strengthen the CENS, southward moisture transport increases, leading to more rainfall over northern Java but less over the Karimata Strait due to

moisture divergence. In contrast, warm SSTs weaken the southward moisture transport, causing moisture convergence to occur over the Karimata Strait, while northern Java becomes drier. These findings highlight the importance of accurately representing oceanic conditions in numerical models to improve the prediction of rainfall associated with CS and CENS events. This aligns with recent research demonstrating that air-sea interactions enhance surface evaporation, which helps reproduce more realistic moisture supplies and convective activities during CENS (Zhao et al. 2025). While these results establish a clear link between SCS SST, CENS intensity, and regional rainfall, it is important to note that this study provides only a broad overview of daily rainfall impacts. Future research should conduct a more detailed investigation into the diurnal rainfall patterns and the shifts in moisture sources resulting from these SST changes, particularly for extreme precipitation events in the western Maritime Continent linked to CENS activity.

This study demonstrates that the standard 40 vertical levels are sufficient to capture the synoptic dynamics of the CENS and confirm that cold SCS SSTs can enhance cross-equatorial flow via the pressure gradient mechanism. However, future studies using higher horizontal and vertical resolutions are recommended to better resolve the inner boundary layer processes and convective initiation mechanisms over the Karimata Strait and northern Java. Finally, this study focuses only on the cross-equatorial flow of CS events. The role of SCS SST in modulating other regional phenomena that interact with CS, such as the Borneo Vortex, remains an open question. For instance, future research could explore whether warm SCS SSTs allow the Borneo Vortex to move eastward, while cold SSTs confine it on the western side of the Borneo Island.

The current results are based on an atmospheric model forced by constant SST values. Because strengthened winds over the SCS may further lower the preexisting SST through surface heat fluxes and ocean currents, the simulated northerly winds may be underestimated in the present simulations. Future studies using a coupled atmosphere–ocean model could address this limitation and capture these atmosphere–ocean processes more comprehensively.

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## Statements & Declarations

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**Author contributions** All authors contributed to the study. (NNP): conceptualization, methodology, software, investigation, formal analysis, writing—original draft, visualization, (MRA): conceptualization, methodology, formal analysis, writing—review & editing, supervision, (EY, NSN): conceptualization, formal analysis, supervision, (TI): formal analysis, writing—review & editing. All authors read and approved the final manuscript.

**Data availability statement** This study utilizes the ERA5 reanalysis data available from the Copernicus Climate Change Service (<https://doi.org/10.24381/cds.adbb2d47>; <https://doi.org/10.24381/cds.bd0915c6>) and OISST v2.1 (<https://psl.noaa.gov/data/gridded/data.noaa.oisst.v2.highres.html>). For model evaluation, radiosonde data were obtained from <https://weather.uwyo.edu/upperair/sounding.shtml> and GPM IMERG V07 precipitation data were acquired from <https://doi.org/10.5067/GPM/IMERGDF/DAY/07>. Data and programming codes that are used to produce the figures are available on <https://doi.org/10.5281/zenodo.21803756>.

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