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Responses of the East Asian summer monsoon to mid-latitude
forcings: effects of Indian summer monsoon heating and Tibetan
Plateau land surface processes

(中緯度の強制に対する東アジア夏季モンスーンの応答： イ
ンド夏季モンスーンの対流加熱とチベット高原の陸面過程の
効果)

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Abstract

The East Asian summer monsoon (EASM) stands out as a significant component within the Asian summer monsoon system. It is characterized as a northeastward extension of a rainband, leading to spatiotemporally persistent rainfall affecting East China, the Korean Peninsula, and Japan in summer. As a subtropical hybrid-type monsoon, large scale circulation of EASM is marked by multiple meridional circulation cells extending from tropical to mid-latitude regions. This feature distinguishes the EASM from the classical-type monsoon (e.g., the South Asian monsoon) with a single meridional cell confined to the tropics, which is largely determined by the land-sea thermal contrast. This feature allows the EASM to interact with various influences of remote factors originating from tropical, mid-latitude, and even polar regions. The complex influencing factors of EASM make its forecast continues to present as a challenging issue, even with increasing attention being paid to extreme weather events such as droughts and floods. For instance, during the abnormal EASM in 2020, operational forecasts started one month ahead are unsatisfactory in predicting its intensity, being 3 to 4 times lower than the observed values. Therefore, it is still motivated to further explore and expand our understanding of the EASM, as the effectiveness of mitigation strategies is strongly influenced by the accuracy of operational forecasts. To achieve this, two mid-latitude forcings and their potential interaction with EASM are presented in this dissertation, which are less understood previously.

The circumglobal teleconnection (CGT) is one of the forcings considered, which is recognized as the dominant summer circulation pattern as a wave train across the Eurasian continent, originating in Western Europe, moving to the northern Indian subcontinent, and reaching East Asia. The evidence from previous studies showed that Indian summer monsoon (ISM) is the primary trigger of the CGT wave train at interannual time scale, the consequent influence of CGT wave train acts as a bridge between the ISM and the EASM. This process, if effective and robust, has the potential to increase the predictability of the EASM. Nonetheless, the monthly and seasonal mean of summer precipitation in the ISM and EASM do not exhibit a long-term coherence during the past few decades; instead, their relationship is non-stationary. This non-stationary might primarily lie on that the monthly to seasonal mean precipitation hinder the signal of the ISM–EASM connection, and therefore this dissertation aims to investigate it at a subseasonal time scale in early summer (June).

The subseasonal lead–lag composite analysis relative to the abnormal ISM heating suggests a possible synchronization between ISM and EASM, namely an EASM rain band appears and grows rapidly after the anomalous ISM heating. The upper-level circulation pattern denotes a potential interaction between CGT wave train and ISM behind this synchronization. The diabatic heating released from ISM precipitation perturbs the air near the westerly jet stream, triggering a Rossby wave trapped within it. This Rossby wave travels along the westerly jet stream to the East Asian sector, with the wave phase orthogonal to the CGT wave train initially, and gradually reverts to CGT pattern in the following days. These features are seen in observation, 100-member ensemble climate model simulation output, as well as a simplified model simulation with imposed ISM heating. Furthermore, the prevailing poleward flow in the East Asian troposphere due to the ISM-induced Rossby wave leads to enhanced mid-tropospheric warm advection and substantial lower-tropospheric moisture transport to the EASM, contributing to the formation of the rain band. Further analysis suggests that ISM

heating may act as a trigger or reinforcer of the EASM rain band, contingent on structure of the westerly jet stream. These findings highlight the importance of accurately predicting the ISM to help improve the prediction of EASM.

The investigation on another mid-latitude forcing relates to the interannual variation of soil enthalpy over the Tibetan Plateau (TP), which reflects thermal conditions at the land surface, including soil temperature and soil moisture. In recent years, the study of land surface processes in the climate system has grown in popularity due to advances in modelling capabilities, improved observations, and a better understanding of the role of land in climate variability. In particular, previous findings have emphasized the critical effect of accurate representation of land characteristics for improved seasonal forecasting. Here, singular value decomposition (SVD) is used to identify a delayed influence of TP land surface processes on EASM. Through homogenous regression, the first SVD mode (36.87 %) highlights the ISM–EASM connection, while the second SVD mode (32.13 %) shows a coupling of the May cold TP with a June horizontal wet-dry-wet East Asian precipitation pattern. The observations from in situ meteorological stations based on second SVD mode suggest that the TP soil enthalpy has the ability to retain the sign of the heat capacity in response to the anomalous atmospheric temperature, which potentially be one of the mechanisms involved in the delayed impact. Soil memory appears to be particularly significant in the central western TP, which accompanies the initiation zone of a Rossby wave pattern propagating along the westerly jet stream. The Rossby wave affects the upper-level circulation pattern in East Asia, potentially modulating the EASM precipitation. The climate model simulations also show that the central western TP has a more pronounced soil memory effect compared to other regions within the TP. These results illustrate the possibility of the delayed influence of anomalous TP soil conditions on the EASM, which provides value-added insights for improving seasonal forecast initialization strategies.

In summary, this dissertation explored two mid-latitude forcings and their potential impacts on the EASM activity: the subseasonal ISM heating and the interannual TP land surface processes. The current study at subseasonal timescale revealed that the ISM–CGT interaction serves as a bridge between the ISM and EASM in early summer, and it contributes to rain band formation over EASM region. Therefore, the results suggest the subseasonal ISM heating could be one of the optimal predictors of the EASM onset and intensity, as well as the associated drought and floods event for East Asian countries. Furthermore, the research proposed the delayed impact of TP soil enthalpy on EASM, which is found to exert a significant soil memory effect particularly in the central western TP. This effect includes the storage of the heat anomalies given from the atmosphere, and it can further interact with the atmosphere in the subsequent month. The results indicate the importance of accurately representing TP soil processes for initializing seasonal forecasting. Overall, this dissertation suggests the improved representations of ISM precipitation and TP soil processes are essential for EASM prediction capabilities. The incorporation of these processes contributes to better mitigation strategies for the socio-economic of East Asian countries under the complex EASM variability. The future works should focus their interactions with the processes arisen from lower atmosphere (e.g., air-sea interaction) and the optimal way to represent them in the numerical simulations.

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List of abbreviations

AGCM	Atmospheric general circulation model
APHRODITE	Daily in situ gridded precipitation data (Yatagai et al., 2012)
BSISO	Boreal summer intraseasonal oscillation
CGT	Circumglobal teleconnection
CGTI-PC1	CGT index based on leading mode of June to August interannual 200-hPa meridional wind anomalies over 20°–60°N, 0°–150°E
CGTI-DW	CGT index based on 200-hPa geopotential height anomalies over 35°–40°N, 60°–70°E, following Ding and Wang (2005).
d4PDF	Database for policy decision making for future climate change, which includes AGCM ensemble output (Mizuta et al., 2017)
EAPR-PC1	Leading mode of East Asian precipitation over 20°–45°N, 100°–140°E
EASM	East Asian Summer Monsoon
ENSO	El Niño-Southern Oscillation
EOF	Empirical Orthogonal Function
ERA5-Land	The fifth-generation reanalysis data of land components from European Centre for Medium-Range Weather Forecasts (Sabater et al., 2021)
GPCP	Global precipitation climatology project monthly precipitation analysis (Adler et al., 2003)
HIST	Historical climate AGCM experiment of d4PDF database
IOD	Indian Ocean dipole
ISM	Indian summer monsoon
ISO	Intraseasonal oscillation
JRA-55	Japanese 55-year Reanalysis (Kobayashi et al., 2015)
LBM	Linear baroclinic model (Watanabe & Kimoto, 2000)
MJO	Madden-Julian oscillation
OLR	Outgoing longwave radiation

List of abbreviation (continued)

TP	Tibetan Plateau
S2S	Subseasonal to seasonal
SVD	Singular value decomposition
WAF	Wave activity flux

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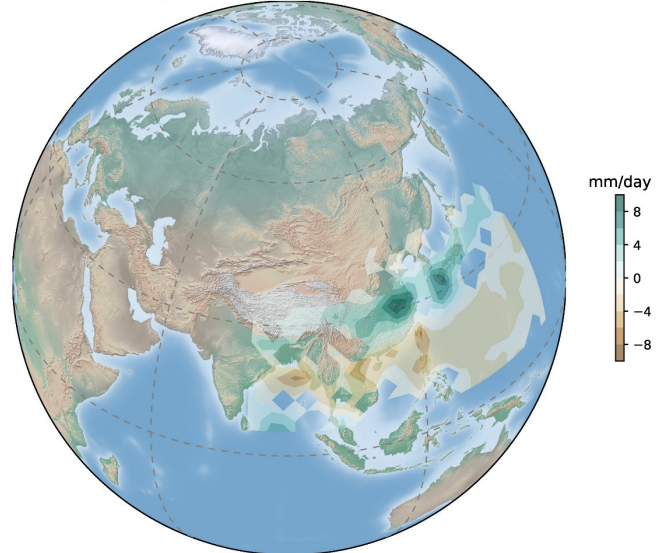
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1

2020 June-July abnormal EASM



Introduction of the East Asian summer monsoon, the circumglobal teleconnection, and the Tibetan Plateau thermal forcing

Brief summary

The East Asian summer monsoon (EASM) results in spatiotemporally continuous rainfall, affecting East China, the Korean Peninsula, and Japan during summer. Despite its interaction with oceanic forcing such as El Niño Southern Oscillation and Indian ocean dipole have been long investigated and have already proved to be important in the prediction, the forecast model still shows limitation. For instance, the predicted intensity of for the extreme 2020 EASM precipitation in operational forecast model was unsatisfactory. This deficiency led to severe socioeconomic damage in East Asian countries, primarily due to the absence of early warning and mitigation strategies. Therefore, delving deeper into our understanding of EASM beyond current knowledge is necessary. This dissertation focuses in particular on two mid-latitude forcings. One is the circumglobal teleconnection, whose interaction with the Indian summer monsoon and the EASM is still a topic of debate. The other is related to the coupling of EASM and Tibetan Plateau land surface processes, a relatively new concept that has emerged in recent years. The review of these two aspects is presented in this chapter.

1. Introduction

1.1. The East Asian summer monsoon

1.1.1. General characteristics

The East Asian summer monsoon (EASM) is one of the most distinct components of the Asian summer monsoon system (**Figure 1-1**), and together with the Indian summer monsoon (ISM; located near south Asia) and the western North Pacific summer monsoon (located near southeast Asia), it forms the entire Asian monsoon system.

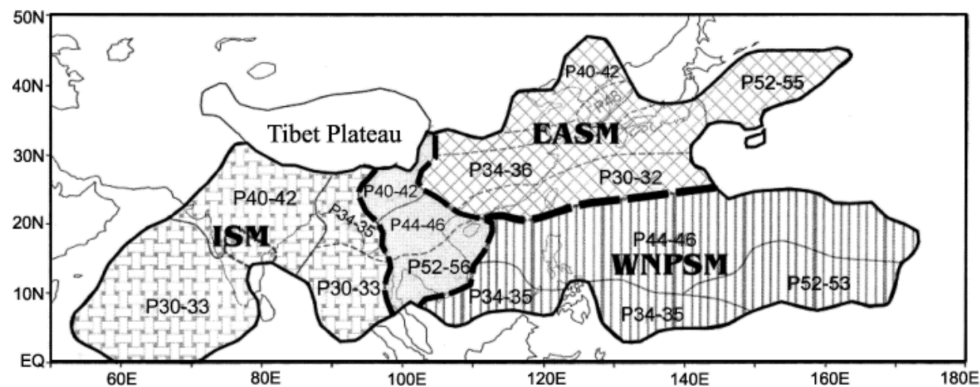


Figure 1-1 The Asian monsoon map. The map shows the three sub-monsoon systems in the Asian monsoon system, including the East Asian summer monsoon (EASM), the Indian summer monsoon (ISM), and the western North Pacific summer monsoon (WNPSM). This figure is taken from Ding & Chan (2005).

The EASM has been characterized and described as a “hybrid” monsoon system within the Asian monsoon system, primarily due to its large-scale circulation, which is influenced by multiple meridional cells over East to Southeast Asia (Lau et al., 2000). These cells span from the deep tropical ocean to mid-latitude regions (**Figure 1-2b**). This is in contrast with the more conventional “classic” South Asian monsoon (e.g., ISM), where the meridional circulation is only confined to the tropics (**Figure 1-2a**). Therefore, the EASM exhibits unique characteristics and dynamics, has strong interaction with the extratropical disturbances from westerly jet stream and the tropical convection owing to this wide latitudinal extension (**Figure 1-2b**).

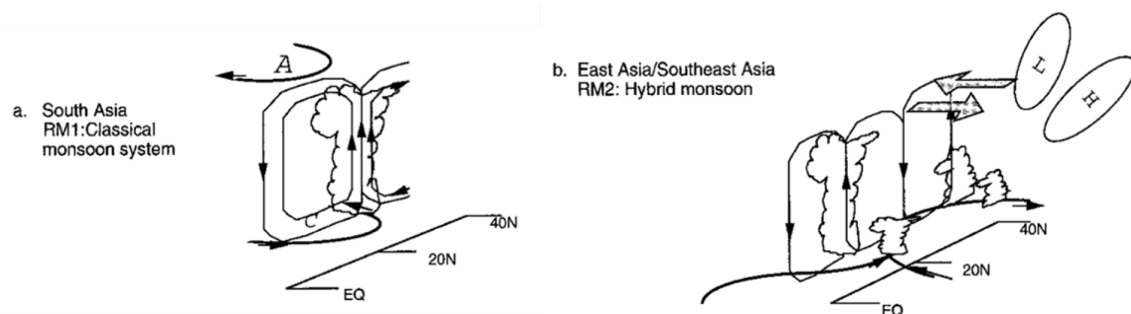


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As for precipitation, the EASM is characterized as a northeastward elongated rainband that which results in spatiotemporally continuous rainfall, affecting East China, the Korean Peninsula, and Japan during summer. This area constitutes the world's most populous region (e.g., Ding & Chan, 2005). Part of the East Asian regions receive EASM precipitation (defined as precipitation from mid-June to mid-July) which accounts for more than 50% of their summer precipitation (**Figure 1-3d**; Y. Ding et al., 2020), with a regional maximum reaching 74% near central-eastern China. (Day et al., 2018). This rainband is typically identified as a planetary-scale stationary frontal system situated in East Asia, due to large scale thermal contrast (e.g., large equivalent potential temperature gradient; **Figure 1-3c**). Namely, the large scale convergence of the air parcels or flows that from higher latitude and lower latitude, near the East Asia (**Figure 1-3b**; Volonté et al., 2022).

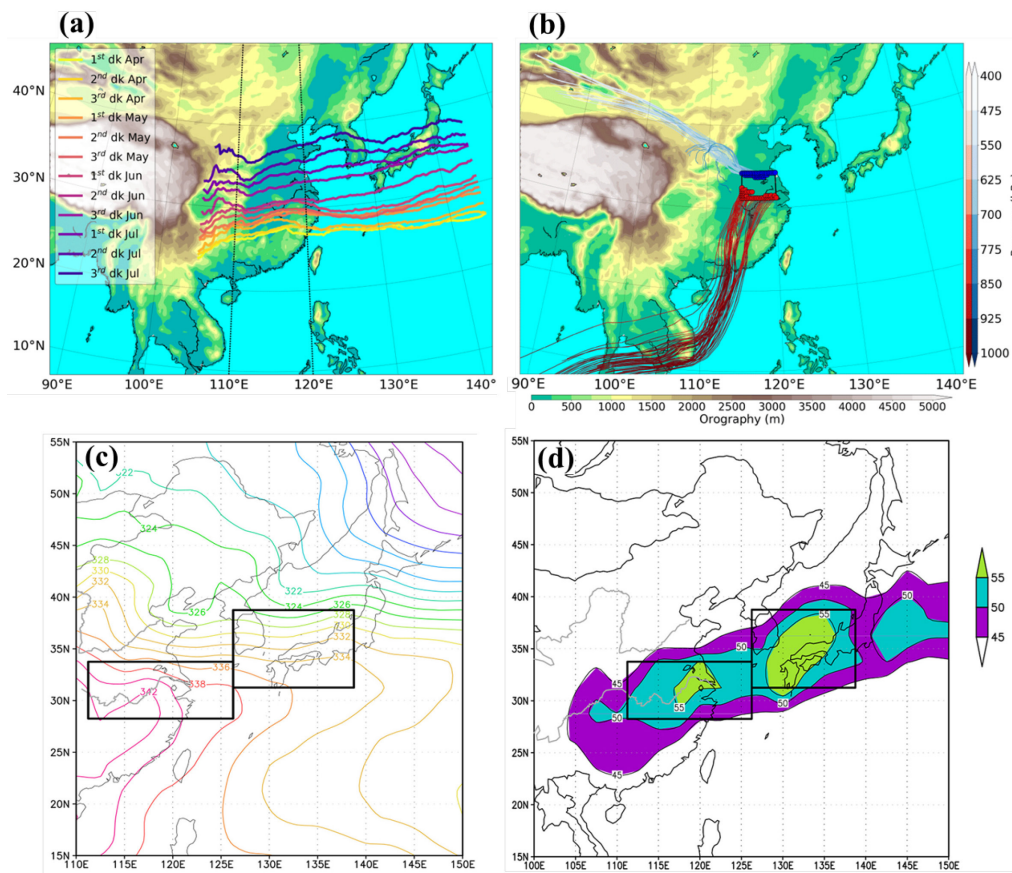


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This stationary front, say the EASM rainband, migrates northward from early summer to late summer (i.e., June–August), transiting according to its development and intensity (**Figure 1-3a**). Hence, the East Asian monsoonal season can be roughly categorized into three stages based on its dynamic features: the onset, development leading to its peak or maturity, and the subsequent decline as it retreats during summer (e.g., Ding & Chan, 2005). On climatology, the migration of EASM rainband is recognized as step-by-step from lower latitudes to mid-latitudes over East Asian sector gradually (**Figure 1-3a** and **Figure 1-4**), while in some years the “migration” might have an abrupt jump without any transition, i.e., skipped onset signature while suddenly outbreak in East Asia which shows extraordinary chaotic nature (Dai et al., 2020). Based on 6-hourly snapshots from the climatological mean of reanalysis data, Liu et al. (2022) identified EASM rainbands activity in boreal summer and discovered that the EASM rainband potentially reaches its peak in June (as shown in **Figure 1-4**). Namely, they found June features the rainband with the largest spatiotemporal coverage, influencing an area exceeding 896,000 km² in East Asian domain. Meanwhile, it also records the highest amount of areal mean precipitation rate in those regions, exceeding 28 mm/day.

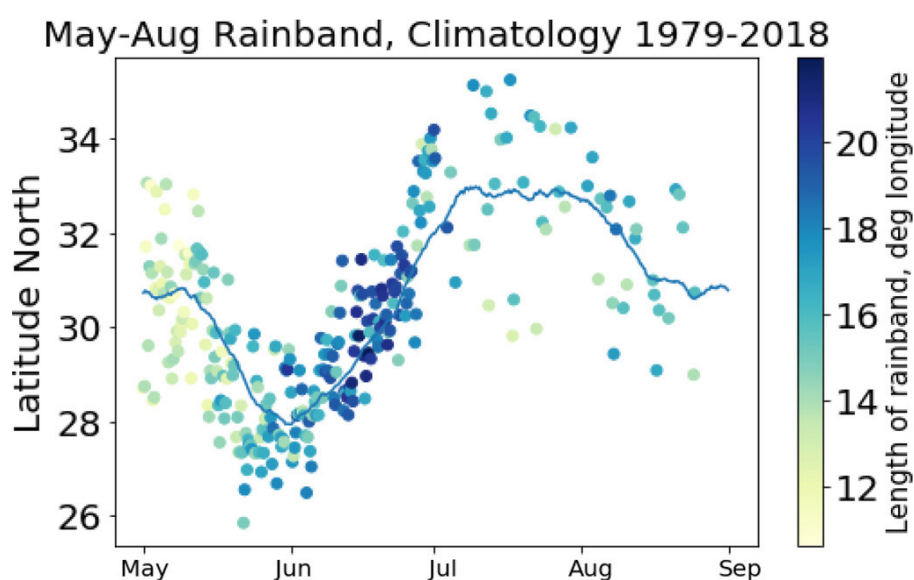


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1.1.2. Influencing factors

In the recent literatures, the term “hybrid” for EASM is used to infer that EASM might be influenced by multiple factors remotely (e.g., Jin & Stan, 2019), which are originating from the tropical region, mid-latitude region, and the Arctic region, owing to EASM’s multiple meridional cells as illustrated in [section 1.1.1](#) (**Figure 1-2b**). A brief review of these identified influencing factors is provided in this section.

The impact of tropical forcing such as El Niño Southern Oscillation (ENSO) and Indian ocean dipole (IOD) on EASM has been long and extensively investigated (e.g., B. Wang et al., 2000; R. Wu et al., 2003; Jianling Yang et al., 2007), which has been proved that they are the important predictors of EASM (Y. Takaya et al., 2020, 2021). In brief, this is through the “capacitor” effect due to the oceanic memory, which anchors a lower-level large-scale anomalous anticyclone that spans across South Asia and East Asia, further influencing the monsoon systems (S. P. Xie et al., 2016).

The mid-latitude forcing involves large internal atmospheric variability, such as the interaction between the Indian summer monsoon and the circumglobal teleconnection (described in detail in [section 1.2](#)), the Tibetan Plateau (TP) thermal forcing (described in detail in [section 1.3](#)), as well as the jet dynamics (Horinouchi et al., 2019). The jet dynamics is crucial for EASM according to the classic theory of ageostrophic circulation, a transverse circulation is established, featuring an ascending branch to the south of the jet stream (wet) and a descending branch to the north (dry). This configuration ultimately contributes to accelerating the upper-level zonal wind (Holton & Hakim, 2013), which can influence the EASM.

Recent research also suggests that the sea ice and the blocking in the pan-Arctic region are important to modulate the EASM rainfall (X. Chen et al., 2021; Nakamura & Sato, 2022). The reduction of sea ice promotes an increase of blocking events at higher latitude. This allows the cold air advection invasion to East Asian domain, consequently amplifying the thermal gradient near the monsoonal stationary front or the rainband.

The complex influencing factors of EASM across different distances (i.e., from local factors to teleconnections) and timescales (i.e., from subseasonal to interannual), making the prediction of EASM still be a challenging task.

1.1.3. The importance for better prediction of East Asian summer monsoon

Forecasting weather and climate events, particularly extreme occurrences like droughts and floods at the subseasonal to seasonal (S2S) time scale, stands as a significant topic in climate prediction, and addressing this challenge is the focal point of a collaborative research initiative between the World Weather/Climate Research Program under the World Meteorological Organization (WMO; Zhang et al., 2019). The large variability of both the precipitation intensity and the duration of EASM often results in droughts or floods in East Asia, seriously affecting the local socioeconomic development. Although some forcings, such as ENSO and IOD, and their associated predictability are extensively investigated (see [section 1.1.2](#)), the skill of predicting EASM activity has not been satisfactory to date, and may require the identification of other forcings that contribute to the continuous improvement of the prediction skill.

For example, from 11 June to 15 July of 2020, which is characterized as one of the extreme EASM years, the accumulated precipitation area-averaged over the Yangtze River basin (27°–34°N, 105°–121°E) reached 167.2 mm and broke the record since 1961 according to the China Meteorological Administration (B. Liu et al., 2020). The floods in Yangtze River basin caused 141 people lost their lives, 28,000 homes were flattened, 3.53 million hectares of crops were affected, with the direct economic loss approximately 11.76 billion US dollars (Zhou et al., 2021). The situation in Japan was also severe, it has reported 84 people lost their lives and 16,599 houses were damaged (based on Prime Minister’s Office of Japan, 2021; web access through the link below:

https://www.bousai.go.jp/updates/r2_07ooame/pdf/r20703_ooame_40.pdf). The S2S models, integral to the WMO operational infrastructure, supply forecast data output from the state-of-the-art GCM that are targeted on decision-making. Despite the involvement of over 11 operational centers/GCMs from various regions in the project (Vitart et al., 2017), these models encountered limitations in accurately predicting the intensity of the EASM in 2020 due to the extraordinary chaotic nature of the precipitation (B. Liu et al., 2020; Nakamura & Sato, 2022; Qiao et al., 2021). It is encouraged to delve deeper into our understanding of EASM, and associated prediction beyond current knowledge is necessary. This exploration may enable us to enhance model performance, leading to benefiting disaster management and preparedness, agriculture, food security, and water resource management within local societies.

1.2. The circumglobal teleconnections

1.2.1. General characteristics

The interannual variability of winter stream function near the mid-latitude waveguide is found circumscribes the Northern Hemisphere, which exhibit a zonal wavenumber-5 feature with a nearly fixed longitudinal location. This circulation pattern therefore is named as the circumglobal teleconnection (CGT; Branstator, 2002). This pattern is believed to be connected to the influence of the waveguide. Its meridional vorticity gradients are about 10 times larger than its zonal vorticity gradients, confining the Rossby wave path to a zonal propagation to downstream (Hoskins & Ambrizzi, 1993; Zhao et al., 2015), in response to a perturbation or forcing near the jet stream.

As for the summer season, the evidence shows that CGT exists as well, with action of centers of geopotential height anomalies over Atlantic–western Europe, west-central Asia, northeast Asia, North Pacific, and North America, still with a wavenumber-5 feature (Q. Ding & Wang, 2005). The summer CGT in the South and East Asian sector is also known as the “silk road pattern” (Enomoto et al., 2003), which describe the Rossby wave activity that trapped in the Asian jet stream. Therefore, the summer CGT can also be described by the 200-hPa meridional wind, and is characterized as the leading circulation mode during the summer season over the mid-latitude Eurasia continent (black box in **Figure 1-5**; Yasui & Watanabe, 2010), based on the empirical orthogonal function (EOF) analysis. Similarly, according to the 200-hPa meridional wind, signal of the summer CGT wave train starts from the Atlantic Ocean and move to and across Eurasian continent, the Pacific Ocean, finally terminated to North America and vanished there.

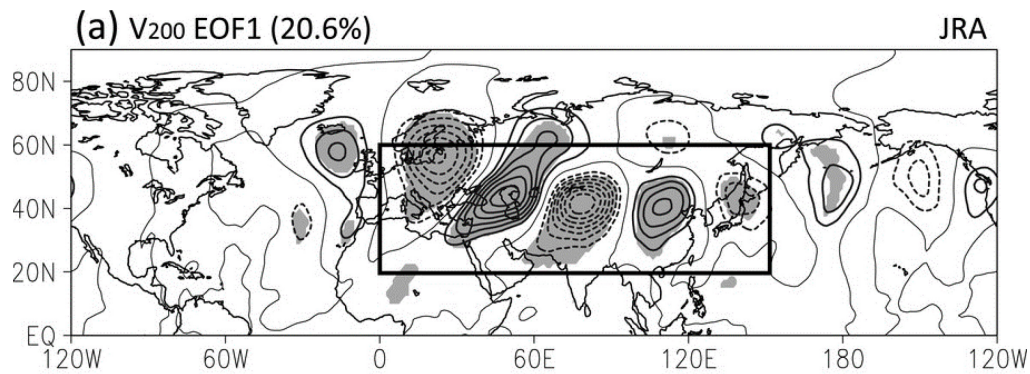


Figure 1-5 The CGT pattern characterized by 200-hPa meridional wind. Leading EOF pattern of 200-hPa meridional wind in June to August during 1980–2004 based on JRA-25, as presented by the linear regression on to the associated PC time series. The EOF is performed over the 20°–60°N, 0°–150°E of mid-latitude Eurasian continent, indicating by thick rectangle. The units are m/s, and the contour interval is 0.5 m/s. The grey shading indicates the anomalies significant at the 95% level. The figure is taken from Yasui & Watanabe (2010).

In this dissertation, hereafter, the phrase of “CGT” will be used to represent the CGT in the summer season, since targeting on EASM and therefore with a preference for the summer season phenomenon. In summary, this section introduced the general characteristics of the CGT, which is manifested as a leading mode of the summer circulation over mid-latitude Eurasia with a zonal wavenumber-5 structure. The 200-hPa geopotential height have a in phase variation over East Asia, the North Pacific, North America, and the Northeast Atlantic when these regions encounter the CGT in summer. In the following section, the factors that trigger or control the CGT is introduced.

1.2.2. Possible triggering sources

To highlight the possible interaction between Asian monsoon systems and CGT, in comparison with other detected teleconnections related to Asian monsoon systems such as “Tokyo-Chicago express” (i.e., which is triggered by the EASM; Lau et al., 2004), Ding and Wang (2005) explored the relationship between Indian summer monsoon (ISM) and CGT at interannual time scale. Based on an index (i.e., CGT-index) of area-averaged 200-hPa geopotential height over Indian subcontinent or west-central Asia (35°–40°N, 60°–70°E), the composite circulation pattern also shows a wavenumber-5 feature, and is found coupled to ISM (**Figure 1-6**; Q. Ding & Wang, 2005).

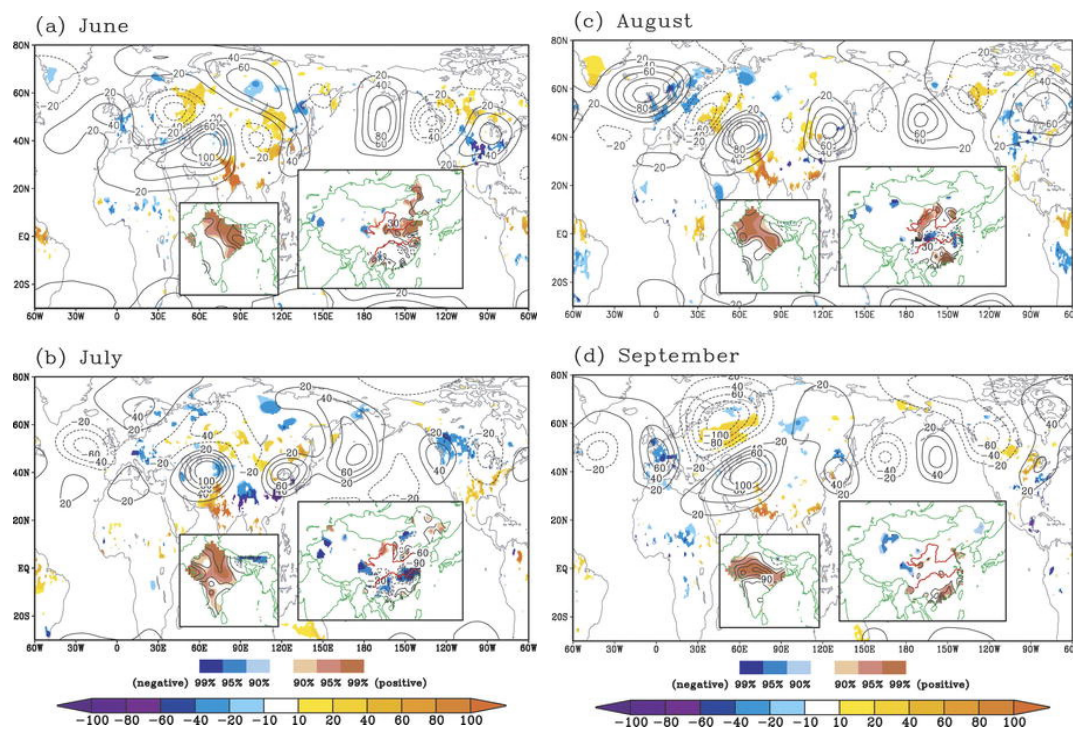


Figure 1-6 The circulation and precipitation pattern during CGT episode in each summer month. Composite difference of 200-hPa geopotential height (contour; units: m), global gridded precipitation (global map; units: mm/month) and precipitation from corresponding meteorological agencies (small map) between positive and negative CGTI years for (a) June, (b) July, (c) August, and (d) September. Differences with $p < 0.1$ are shown by shading in global map. For India and China station data are from corresponding meteorological agency, with color shadings are representing their confidence interval by composite. This figure is taken from Ding & Wang (2005).

The circulation patterns are similar for each summer months from June to September, namely a wave train originated from the west Europe and propagates to ISM domain and further downstream such as East Asia and North America. Meanwhile, it is coupled to persistent abnormal precipitation over the ISM domain. Notably, the East Asia is covered by an anticyclone at 200-hPa, accompanied with precipitation anomalies over East China. The locations of East China precipitation anomalies differ in each summer month with respective of the location of the 200-hPa anticyclone.

Based on those evidence from observation, Ding and Wang (2005) hypothesized that there might be two scenarios that can describe the interaction between ISM and CGT. In the first scenario, the ISM with more active role as a Rossby wave source to trigger the CGT. While in the second scenario the ISM is influenced by the CGT wave train that is originated from western Europe, which brings strong stationary wave energy enhanced convection near the ISM domain. The follow-up works targeting on ISM–CGT interaction show these two scenarios might also work collaboratively at

subseasonal time scale during June to September (e.g., Q. Ding & Wang, 2007, 2009), namely, the CGT wave train triggers the ISM to become active, and then its heating helps reinforce the CGT wave train alternatively, leading to the situation in which the CGT wave train and ISM induced local CGT wave train are in-phase and propagates to East Asia (**Figure 1-7**).

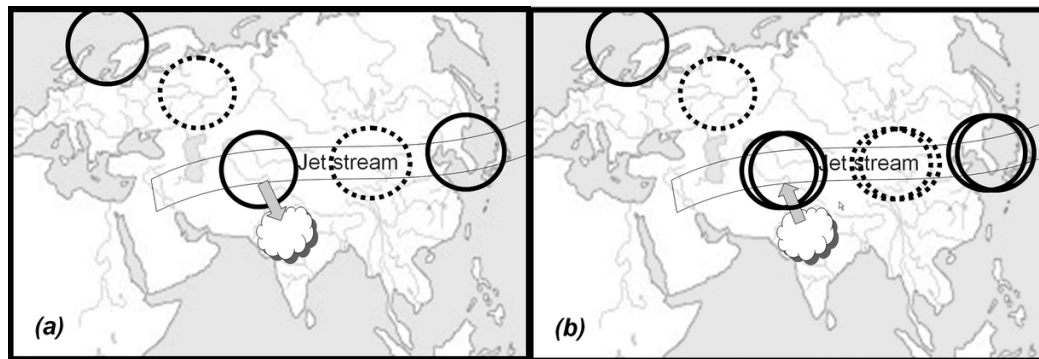


Figure 1-7 Possible interaction between the CGT and the ISM. (a) The strong ISM convection is triggered by the CGT that originated from western Europe, and (b) the ISM convection, alternatively, excites a Rossby wave that propagates in phase (or reinforcement) of the CGT to downstream. This figure is taken from Ding & Wang (2007).

However, there is also evidence to suggest that the ISM may not be the primary trigger for the CGT. Yasui & Watanabe (2010) used numerical simulation to investigate the forcing mechanism and subsequent predictability of the June to August CGT variation, since previous research mostly are only based on observational data. The simulation using the nonlinear baroclinic model in their result shows that imposing observed global diabatic heating pattern is able to reproduce the observed CGT variation at interannual time scale, for which using climatological global diabatic heating pattern cannot do. To further quantify the regional diabatic heating contribution to the CGT predictability, based on the singular values decomposition between CGT pattern and global diabatic heating pattern, Yasui & Watanabe (2010) selected 11 regions for further investigation using the linear baroclinic model (LBM). The selected sub-regions as a total successfully triggered the CGT pattern simulated by LBM (**Figure 1-8a, b**). At the same time, the individual forcing from each 11 sub-regions that prescribed in LBM shows the region K might be the most important forcing to trigger the CGT (explained near 60% of the total variance), which is located near the Mediterranean (**Figure 1-8c**, Yasui & Watanabe, 2010). Additionally, the regions C and I, corresponding to regions of equatorial Africa and North America, are also important candidates which both explain more than 15% of the total variance, respectively. The heating in South Asia, encompassing the ISM domain, only accounts for less than 5% of the total variance.

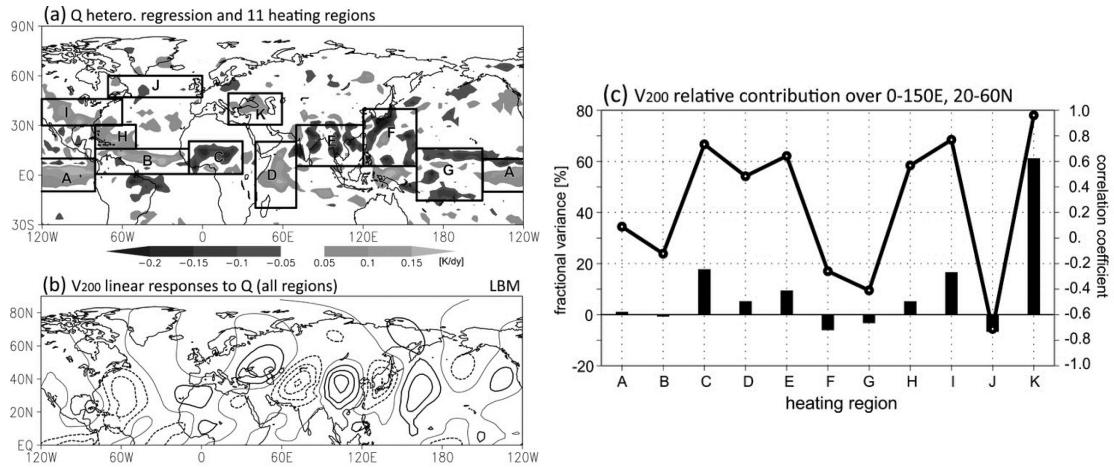


Figure 1-8 The decomposed regional diabatic forcing of CGT. (a) The global diabatic heating pattern coupled to CGT in June to August, with 11 decomposed regional diabatic forcing (boxes from A to K) that used in the LBM (b) the LBM response to the sum of the 11 regional forcings that defined in (a). (c) Fractional variance (bar) and spatial correlation (curve) of the 200-hPa meridional wind response to each regional forcing with that from (b), evaluated over mid-latitude Eurasia (20°–60°N, 0°–150°E). This figure is taken from Yasui & Watanabe (2010).

Likewise, using hindcast-type simulation with nudging of specific regions in simulation towards the reanalysis data, Beverley et al. (2021) show that Western Europe (45°–65°N, 25°W–5°E) might be the key region for improving the simulated interannual variation of August CGT. In comparison those with nudging of west Asian (35°–40°N, 60°–70°E) and ISM domain (0°–30°N, 60°–100°E), respectively, only with a slight improvement or even worse relative to the control run. This feature may be due to the fact that Western Europe is the region closest to the entrance of the jet stream, i.e. where the CGT started. In their experiment, it also shows the results obtained from August (i.e., which highlight the importance of western Europe) may not be suitable for other summer months. This implies that the formation mechanism of CGT is quite different in each summer month.

In recent years, the forcing mechanism of CGT and its impact on weather and climate has become popular topic. The forcing mechanisms of the CGT are also found in relation to land surface processes in West Asia on interannual time scales (Jingqiu Yang et al., 2021) or due to the tropical convection which enhance the midlatitude circulation waviness at interannual to interdecadal time scale (Sun et al., 2022). Therefore, it is still questionable whether the CGT is related to ISM forcing. The popularity of research focus on GCT in recent years because of its important impact on the weather and climate over mid-latitude region, which will be illustrated in the next section.

1.2.3. Impact on the weather and climate

Boers et al. (2019) found that extreme precipitation in south-central Asia (i.e., northern Indian subcontinent) has the ability to synchronize the daily extreme rainfall events around the globe (**Figure 1-9**), using gauge-calibrated and satellite-derived rainfall dataset of from Tropical Rainfall Measurement Mission (known as TRMM). The daily extreme precipitation over regions with significant synchronization, or in other words, significantly tele-connected to the ISM extreme precipitation, which includes eastern US, western Europe, Tibetan Plateau and East Asia, etc., could be attributed to the influence of the CGT wave train with the interaction of the ISM. These results are in accordance with another study, which refers to ISM as the “heart” for tropical climate prediction (B. Wang et al., 2015). Therefore, the ISM is often believed as the important forcing source to alter the weather and climate conditions in the East Asia. The anticyclone (or high pressure) as one of the centers of actions of CGT, as illustrated in [section 1.2.2](#), is found tightly coupled to the interannual variability of precipitation and temperature over the East Asia. This high pressure steers southeasterly flow along its southern edge, which transports abundant moisture into East Asian countries, meanwhile, it also blocks the perturbation and troughs from the upstream and therefore beneficial to the heavy rainfall to its west in East Asia (Q. Ding & Wang, 2005). At the same time, the southeasterly flow leads to increase in surface air temperature and specific humidity, accompany with the subsidence motion in the center of the high pressure, cause the heatwaves that occurs near the Korean peninsula (Cha et al., 2021; Kim et al., 2019).

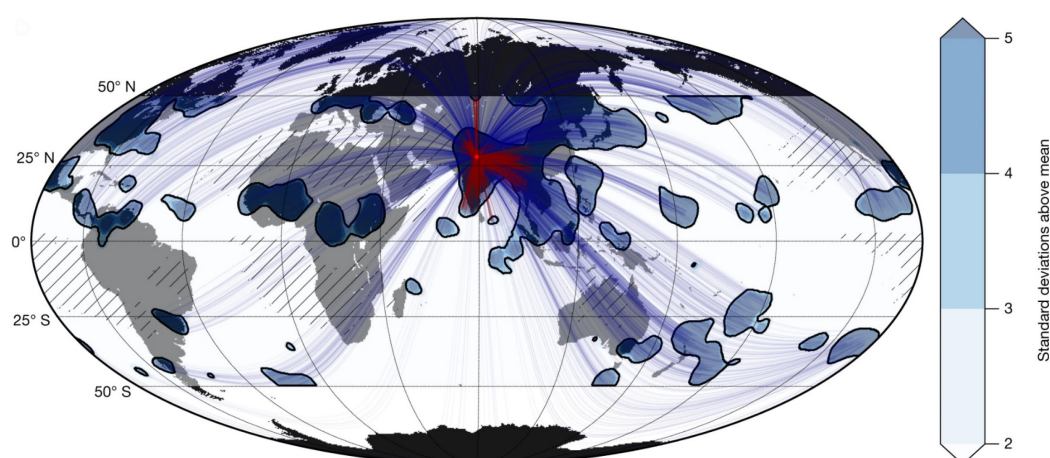


Figure 1-9 Teleconnection pattern for regional extreme rainfall synchronized with extreme ISM. Global extreme precipitation (50° S to 50° N) that significantly ($p < 0.001$) synchronized with ISM rainfall (shading). Links stand for regional connective density, with shorter (longer) than 2,500 km shown in red (blue). This figure is taken from Boers et al. (2019).

As depicted in [section 1.1.3](#), the EASM in 2020 was characterized as unusually extreme. The 2020 June to July EASM precipitation anomalies shows an outstanding wet band that extend from central-east China, across the Korean peninsula, and terminated to southern Japan (Pan et al., 2021). Within Asian summer monsoon region, the ISM domain also with noticeable abnormal precipitation anomalies (**Figure 1-10a**).

The precipitation anomalies are coupled to 200-hPa circulation pattern show a CGT-like pattern across the Eurasian continent (**Figure 1-10b**). The 20-year integrated AGCM experiment with prescribed climatological SST and an imposed heating anomaly near the ISM region (8°–25°N, 60°–85°E) indicates that the ISM heating might serve as one of the triggers for the CGT pattern and abnormal EASM in 2020 (**Figure 1-10c**).

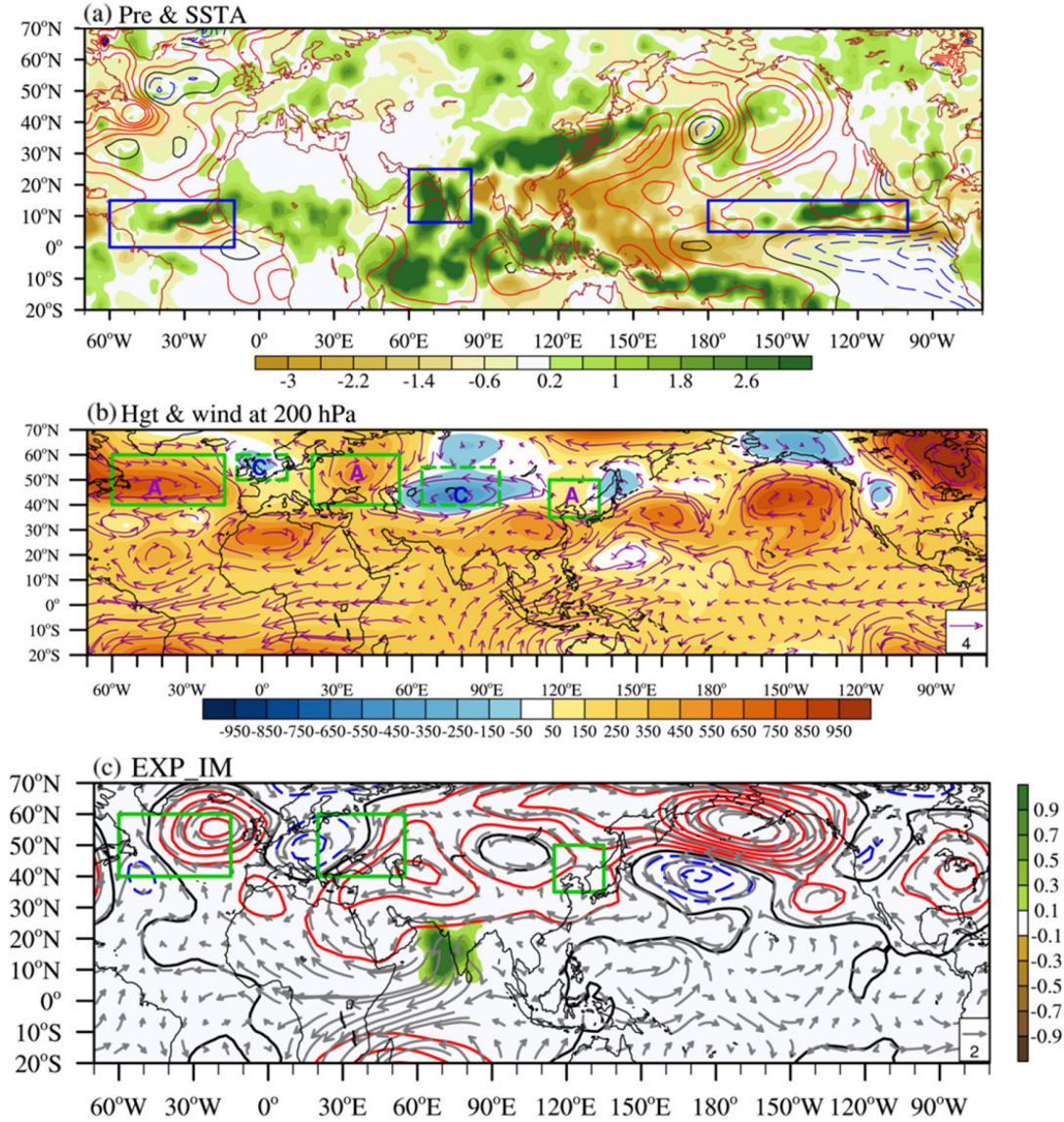


Figure 1-10 The 2020 abnormal EASM and its circulation pattern. 2020 June to July (a) precipitation anomalies (shading; unit: mm/day) and SST anomalies (contour; K), and (b) 200-hPa geopotential height (shading; unit: m) and horizontal wind (vectors; unit: m/s). (c) AGCM experiment with imposed ISM heating anomaly (shading; unit: K/day) with simulated 200-geopotential height difference with control run. This figure is taken from Pan et al. (2021).

On top of that, the glaciers over the southern Tibetan Plateau are also influenced by the ISM–CGT interaction. In short, the onset of the ISM also causes the airflow from

the Indian subcontinent to move towards the Tibetan Plateau, bringing more precipitation to the interior of the Tibetan Plateau, which contributes to the determination of the annual mass balance for the glaciers, forcing the annual mass balance to be positive (i.e., accumulation; Mölg et al., 2014).

In conclusion, the impacts of CGT and ISM with their interaction greatly influence the weather and climate in the mid-latitude Northern Hemisphere, and therefore is considered as an important predictor. For the interest of this dissertation, with particular focus on the EASM, more information on the interaction between the CGT, the ISM and the EASM is presented in detail in the next section.

1.2.4. Debates in connection between the Indian and East Asian summer monsoons

On climatology, the onset date of the Indian monsoon is near beginning of June (or near end of May), while that for the EASM is usually later, with an average lag time of about 2 weeks (Y. Ding et al., 2020). Interestingly, the year-to-year onset dates of these two monsoon systems indeed have a good correspondence during 1948–1999 (**Figure 1-11**; Y. Liu & Ding, 2008), with a correlation coefficient of 0.30 ($p < 0.05$). In other words, when the ISM onset earlier than climatology, the East Asia would enter the EASM season earlier than climatology as well, and vice versa. This statistical relationship denotes the ISM–CGT interaction, which could potentially influence the EASM as described in [sections 1.2.2](#) and [1.2.3](#). If this theory proves effective and works well, it has the potential to yield partial predictability of the EASM, consequently enhancing the accuracy of EASM forecasting to some degree.

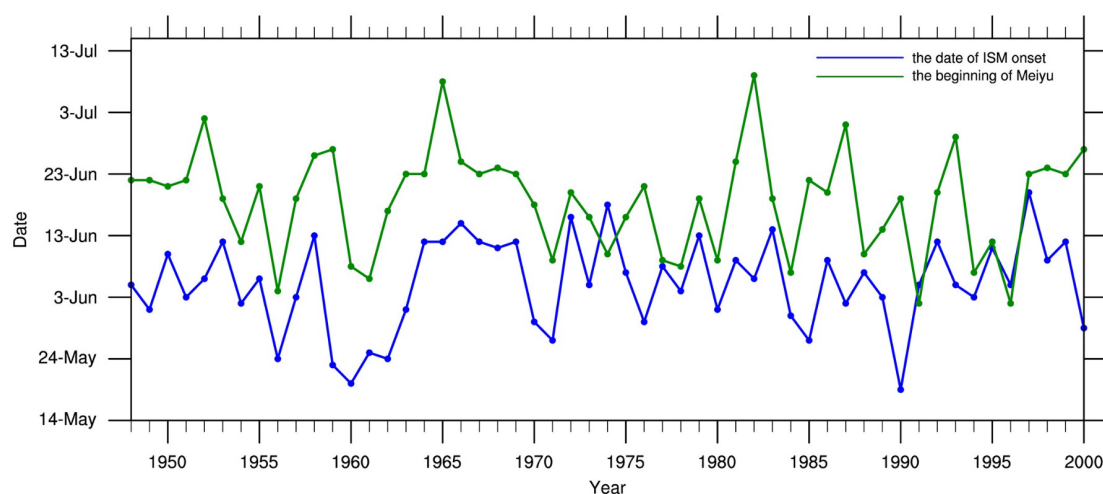


Figure 1-11 The coincidence of the onset dates between ISM and EASM. Time series of the onset dates for ISM (blue) and EASM (green) during 1948–1999, with correlation coefficient of 0.3 ($p < 0.05$). The EASM is labelled as Meiyu, which is the name of the EASM in Chinese. This figure is taken from Ding et al. (2020).

Nevertheless, the precipitation over ISM domain and EASM domain does not have a long-term correlation, despite their good correspondence in terms of the onset dates at interannual time scale. For instance, the June to September ISM precipitation is positively correlated to the precipitation over Northeast China, while negatively

correlated to the precipitation over southern Japan during 1959–1979 (R. Wu, 2017). This relationship obtained of 1959–1979, however, is disappeared or reversed in the recent decades (**Figure 1-12**; 1985–2005), respectively, for ISM precipitation versus northeast China and ISM precipitation versus southern Japan. This indicates that the ISM–EASM relationship is not robust. Therefore, ISM does not serve as a candidate for EASM prediction, which has sparked many discussions for the plausible reasons and mechanisms behind this non-stationarity connection. The discussions are summarized as three aspects below.

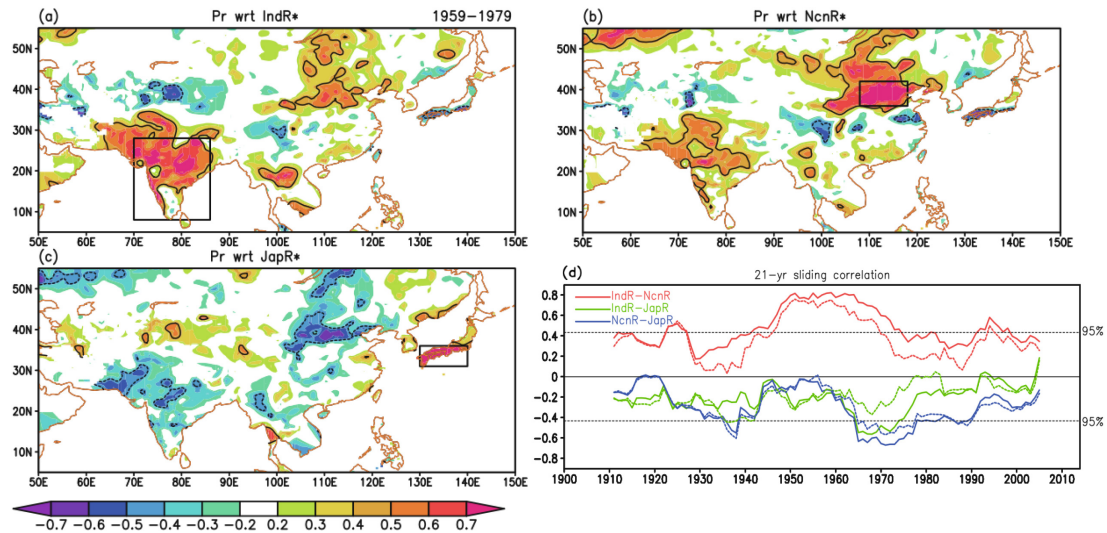


Figure 1-12 Precipitation correlations of ISM and EASM. Correlation coefficients of Eurasian precipitation with respect to area-mean (a) Indian, (b) northern China and (c) southern Japan rainfall (black boxes in each panel) during 1959–1979, the black contours denote regions with $p < 0.05$; and (d) their 21-year moving correlation between area-mean rainfall of Indian versus northern China (red), Indian versus southern Japan (green), northern China versus southern Japan (blue), with (dashed) and without (solid) eliminating the ENSO signal, the dashed line indicated the 95% confidence interval. This figure is taken from Wu (2017).

The first key factor influencing the ISM–EASM connection is related to the variation of the upper-tropospheric westerly jet stream, namely, the waveguide. In theory, the perturbations, such as those caused by ISM heating, can serve as a Rossby wave source to induce Rossby waves near the westerly jet stream. These waves are expected to be constrained and guided by the westerly jet stream, propagating to downstream along its path before dispersion. (Hoskins & Ambrizzi, 1993). Therefore, the position, structure, and speed of the upper-tropospheric westerly jet stream is proposed important for ISM–EASM connection. Lin et al. (2017) found that the wavelength near the westerly jet stream is reduced in 1980–1998 compared to that in 1962–1979, this difference instigates the location of the anticyclone over East Asia as part of CGT. The location of anticyclone thereby moves from East China to Korean Peninsula–Japan Sea related to the reduction of the wavelength, which result in the shifting of the EASM precipitation as one of the responses as well. To summarize, jet

stream is important for the ISM–EASM connection, which can lead to ISM–EASM connection complex and less predictable.

The second factor is attributed to the decoupling between ISM and ENSO. The severe floods in India have often been aligned with La Niña events, and vice versa for droughts (e.g., Kumar et al., 2006). The ENSO is accordingly considered the most crucial factor for predicting the ISM, because its interaction with Walker circulation controls the large-scale divergence and convergence, as well as other impacts (e.g., land-sea tropospheric temperature gradient; Chowdary et al., 2021). In spite of that, in recent decades, the relationship between the ISM and ENSO has become less consistent, sometimes even showing an inversed correlation due to the southward shift of the Walker circulation. This has led to the possibility of strong ISM occurring during El Niño event (Kumar et al., 1999). H. Wang et al. (2012) found the variability of the ISM precipitation is reduced in recent decades, due to its decoupling with ENSO, which also suppress the CGT variability and further weakened the major CGT action of centers (e.g., the East Asian anticyclone). This may also be one of the possible mechanisms to explain the non-stationarity of the ISM–EASM connection.

As for the last factor, it has already mentioned and explained in [section 1.2.2](#). Namely, it is still questionable whether the ISM is the primary trigger of the CGT, or if ISM can enhance the predictability of the CGT (e.g., Yasui & Watanabe, 2010, Beverley et al., 2019).

To bring all subsections in [section 1.2](#) together, it is still yet unclear that if ISM is able to modulate the EASM, which is plausibly bridged by CGT. This question primarily lies on three issues: the first concern is whether the ISM can trigger the CGT wave train towards East Asia (i.e., regional CGT), the second issue is how regional CGT alter the EASM, and the last one for whether the coupling between regional CGT and the EASM is contingent on particular jet stream structures. Meanwhile, it is also speculated that the previous studies mainly using monthly mean or seasonal mean precipitation may hinder the signal of ISM–EASM connection. This is because both monsoon systems are influenced by multiple factors across various timescales. Therefore, we expect subseasonal analysis for ISM–EASM connection might be more appropriate.

1.3. The Tibetan plateau thermal forcings

1.3.1. General characteristics

The Tibetan Plateau (TP), often referred to as the “roof of the world” or “the third pole”, is a vast and elevated region located near central Asia, with an average elevation of over 4000 m. Due to this high elevation with existing cryosphere, which makes TP’s unique and complex environment with interactions among the atmosphere, biosphere, and other components of the climate system and provides a substantial large effect on the global climate and water cycles.

The tropospheric heating over TP on climatology, is characterized by a heat sink in the winter season (about negative 50 W/m² during November to February) while is known as a strong heat source in the summer season (up to 150 W/m²). Thus, land and atmosphere is coupled through the sensible and latent heat fluxes (Yanai et al., 1992). The sensible heat flux (dry processes) and latent heat flux (wet processes) is equally

important to maintain vertical circulation near the TP (Ueda et al., 2003). Given that these processes occur mainly in the mid-troposphere near the surface of the TP, it follows that the TP could contain and maintain a large-scale vertical circulation associated with this thermal forcing, shaping the weather and climate of its surrounding (e.g., Sato, 2005; Sato & Kimura, 2007). Meanwhile, the idealized AGCM experiment shows that large scale vertical circulation near TP is largely dominated by the TP thermal forcing rather than the TP orographic forcing (Yimin Liu et al., 2007).

The heat source/sink over the TP exhibits substantial variability on an interannual time scale (Jingchen Liu et al., 2022; Huan Wang & Li, 2019), consequently, this variation influences the weather and climate in the surrounding areas. The TP source/sink is primarily influenced by well-known oceanic drivers like ENSO and IOD (Guan & Yamagata, 2003; Wang & Xu, 2018). Recent literature has underscored the significance of local land surface processes, making them a noteworthy and upcoming topic, as will be elaborated in the subsequent section.

1.3.2. Coupling with the land surface processes

Snowpack is an important component of the climate system over the TP, which with physical properties of high reflectivity and emissivity while low thermal conductivity, forming the regional lower boundary of the atmosphere (S. Li et al., 2021). Hence, it exerts a direct impact on local surface radiation budgets and turbulent heat fluxes, thereby influencing the atmospheric circulation. In addition to the direct impact closely associated with the "albedo feedback," another important effect is known as "the snow hydrological feedback." This effect indicates the prolonged influence of melted snow on soil moisture and its interaction with the atmosphere (Yasunari et al., 1991) make the impact of snow on climate system which persists through winter to spring. For instance, the April to May mean TP surface air temperature is found strongly correlated to February to April averaged TP snow cover days (Y. Zhang et al., 2019). It is important to note that the snow cover is generally shallow over vast region of TP, with the exception of the western TP. As a result, many studies indicate that the variation in snow cover, which influences the heat exchange between the TP and the atmosphere during spring and summer, is predominantly observed in the western TP, whose impact is relative local (e.g., Xiao et al., 2019). Meanwhile, results consistently show that the snow effect may vanish in late spring to early summer (e.g., H. Liu et al., 2004; Xiao & Duan, 2016).

On the other hand, the soil enthalpy in part of soil memory over TP might be crucial for the seasonal prediction from spring to summer. The definition of the soil enthalpy (Q_{se}), which is the depth-integrated product of soil temperature and specific heat capacity, could be expressed as follows:

$$Q_{se} = \int_0^Z C_h \cdot T(z) dz$$

where T is temperature at each depth of the soil (z ; positive downward), and the C_h stands for specific heat capacity of the frozen and unfrozen soil (i.e., bounded by 273.15 K), respectively, with values of 1.4×10^6 J/m³/K and 3.0×10^6 J/m³/K (DeGaetano et al., 1996). It is not difficult to find that the soil enthalpy is largely determined by the soil temperature, and various atmospheric and land processes can contribute to the soil

temperature fluctuation. At the same time, the soil water phase related to the soil temperature is recognized as significant. This is due to the considerable difference in latent heat during the water phase change, which is approximately 78 times greater than the specific heat of liquid water per unit of temperature (H. Xu et al., 2022). Hu & Feng (2004) indicated that anomalies in soil enthalpy within the top 1 meter of soil can endure for 2–3 months, a notably longer duration compared to soil moisture anomalies, which last for approximately 1.8 months. Therefore, the soil temperature exerts a stronger soil memory than that of soil moisture. This is notably the case for the TP, as in situ measurements from meteorological stations revealed that the soil temperature below the top 0.4 meters retains a memory for close to 5 months, while the shallower soil temperature at 0.2 meters maintains a memory for around 3 months (Ye Liu et al., 2020).

In recent literatures, the discussion on the soil enthalpy associated to soil temperature in the climate system has grown due to advancements in modeling capabilities, improved observations, and an increased understanding of the role of land in climate variability. More importantly, it is shown that better understanding and accurately representing land surface processes are crucial for improving climate predictions, and is beneficial to the subseasonal to seasonal forecast (Dirmeyer & Halder, 2017; Merryfield et al., 2020). Numerous studies also suggest that land surface processes play a crucial role in predicting the EASM. This significance will be further explored in the following section ([section 1.3.3](#)), with a specific focus on the TP.

1.3.3. Interaction with East Asian summer monsoon

The formation of the EASM has long been attributed to the influence of the TP, particularly in terms of its thermal forcing. It has been shown that in the idealized AGCM experiment, in which the global terrain is set to 0 meters to represent only the land-sea thermal contrast, this experiment alone does not substantially explain the precipitation pattern of the EASM. Instead, there is almost no precipitation occurring in East Asia (**Figure 1-13**; G. Wu et al., 2012). Meanwhile, the experiment incorporating TP terrain but setting the vertical diffusive heating term to be 0 in the atmospheric thermodynamic equation also fails to explain the precipitation associated with the EASM, which highlights the impact of TP heating is crucial for EASM (G. Wu et al., 2012). The comparison between the relative effects of TP orographic forcing and TP thermal forcing on the EASM has been the subject of extensive investigation. However, the conclusions drawn from these studies are remarkably diverse, namely, some studies show that TP thermal forcing on the EASM is critical, others do not (e.g., Son et al., 2020; G. Wu et al., 2012). This is likely influenced by factors such as model configurations, experimental designs, and other complicated processes since the nowadays AGCM are enough complicated.

B. Wang et al. (2008) discovered a correlation between the 2-meter air temperature (T2m) of the TP and the EASM. Their study revealed that warmer temperatures in the TP are associated with reduced precipitation in northeastern China. Conversely, increased precipitation occurs, forming a rainband that extends from central South China to the Korean Peninsula. Further numerical simulation shows that heat source over TP can trigger two Rossby wave trains. One Rossby wave train at 200-hPa exhibits an anticyclone near the Tibetan Plateau, extending toward East Asia and contributing to strengthening of the South Asian high. The intensified upper-level divergence

associated with this high-pressure system induces upward movement to the east of the Tibetan Plateau, consequently resulting in increased precipitation over East Asia. Another distinct wave train at the 850-hPa level travels from the northern Arabian Sea to the South China Sea, generating an anticyclone there which facilitates the convergence of water vapor towards the EASM regions. In this perspective, the interaction between the TP heat source or sink and the EASM is recognized by its influence on the large-scale atmospheric circulation, in particular by exciting the Rossby waves propagating to the EASM, inducing the large-scale convergence and divergence to control the EASM.

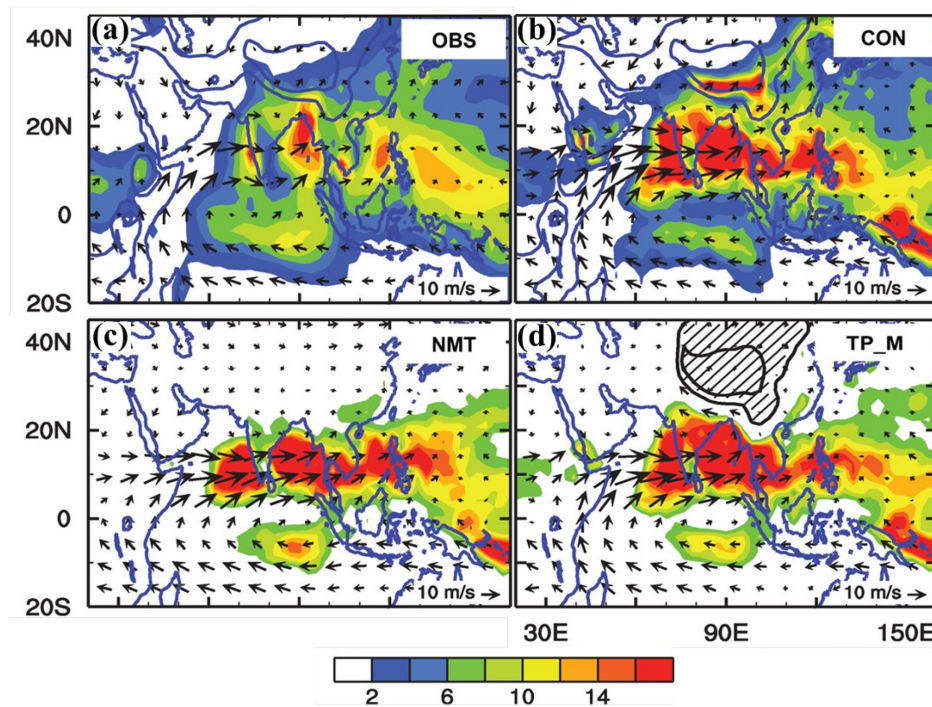


Figure 1-13 Impacts of TP on EASM. Summer June to August precipitation rate (color shading, unit mm/day) and 850-hPa winds (vectors) for (a) observation averaged from 1979–2009, (b) the control experiment (CON), (c) the experiment without terrain across global (i.e., land area with elevation of 0 m) (NMT), and (d) the experiment with presence of terrain for TP only, but eliminate the thermal forcing of it (TP_M). The experiments are averaged over 7 years period. Thick contours and hatched regions indicate terrain that present in the model. This figure is taken from Wu et al. (2012).

Further evidence of how land surface processes is involved in this interaction is presented in the work by Xue et al. (2018), primarily focusing on a case study from the year 2003. In 2003, the observed area-mean May TP T2m is -0.86°C colder than usual, while most reanalysis data show a biased value compared to this in situ observational values (Xue et al., 2021). Simultaneously, in June, central East China experienced severe floods, while southern China suffered from droughts. To prove the possible linkage between May cold TP and June abnormal EASM activity, Xue et al. (2018) conducted a hindcast-like experiment for 2 months, from end of April to beginning of July with eight ensemble members. The simulations initialized from the reanalysis data with updating the observed SST and sea ice. Apart from the control run with no tuning,

additional experiments include one involved imposing cold soil enthalpy, while the other utilized climatological SST and sea ice, respectively, to examine the impact of cold TP and SST on 2003 abnormal EASM. The experiment turns out that the observed 2003 abnormal EASM is mainly regulated by the cold TP, instead of the SST and sea ice forcing which could only partially explained (**Figure 1-14**; Xue et al., 2018). These results provide an insightful understanding of the initialization of the land component to improve the EASM prediction, which motivates more researchers start from 2003 case study to explore a general relationship of TP soil enthalpy and EASM. This will also be one of the goals in this dissertation.

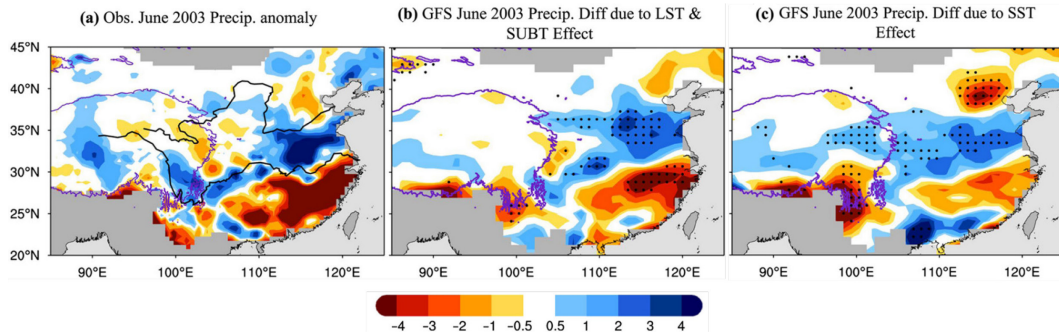


Figure 1-14 Impacts of cold TP and SST/sea ice forcing on 2003 abnormal EASM. 2003 June precipitation rate (color shading, unit mm/day) for (a) observational anomalies (b) difference between cold TP soil enthalpy experiment minus control run, and (c) difference between control run minus climatological SST/sea ice experiment. The dotted regions with $p < 0.1$. This figure is taken from Xue et al. (2018).

1.4. Objectives

As discussed in [section 1.2](#), prior studies have indicated that the ISM serves as the driver of the CGT, which confined within the westerly jet stream. Consequently, the propagation of the CGT along the westerlies jet stream potentially links the ISM and EASM. Despite of this, long-term correlation between precipitation in the ISM and EASM is not found, which sparked extensive discussions. The proposed mechanisms from previous literatures, including the potential decoupling between the ENSO and the ISM, variations of the wave guide, or the likelihood that the ISM not necessarily be the primary trigger of the CGT, are investigated mainly at interannual timescale. As illustrated in [section 1.2.4](#), using monthly mean or seasonal mean precipitation is speculated to hinder the signal of ISM–EASM connection. Although two scenarios of ISM–CGT interaction (i.e., **Figure 1-7**) are proposed by Q. Ding & Wang (2007, 2009) through the subseasonal analysis, but the subseasonal ISM–EASM connection remains unknown. It is prompted to investigate the plausible bridging role of CGT between ISM and EASM. This dissertation delves into the ISM–EASM connection at the subseasonal time scale, examining the interplay roles of ENSO and the jet stream with it.

As introduced in [section 1.3](#), through a case study by 2003, which shows the improved land initialization strategy (i.e., accurately prescribe the soil enthalpy in

AGCM) can greatly improve the EASM prediction. However, the general relationship between TP soil enthalpy and EASM is yet unclear. Namely, it is yet unknown if the strategy of simulation for 2003 (i.e., **Figure 1-14**) can be directly applied to other years or not, which motivated this dissertation to add more evidence and mechanism of it. Namely, the endeavor is to delineate a general coupling pattern between TP soil enthalpy and EASM, as well as present the hints of the potential underlying mechanisms.

These investigations and discussions are allocated to [chapter 2](#) and [chapter 3](#), respectively, and they are referred to as mid-latitude forcing I and II, since they are in the scope of mid-latitude forcings.

1.5. Data and Methods

(i) Data

In this section, a part of the data and methods that will be used throughout the dissertation is introduced. To investigate the precipitation variability associated to EASM, the monthly precipitation data with a horizontal mesh of $2.5^\circ \times 2.5^\circ$ from the Global Precipitation Climatology Project (GPCP, Adler et al., 2003) version 2.3 was used for 1979–2020, which partially derived from the satellite data.

The atmospheric circulation and hydrological parameters were derived from the Japanese 55-year Reanalysis datasets (JRA-55; Kobayshi et al., 2015) with spatial resolution of $1.25^\circ \times 1.25^\circ$, spanning from 1958 to 2020. Both monthly-mean and daily mean values from JRA-55 are analyzed. The gridded soil parameters are used from ERA5-Land (Muñoz-Sabater et al., 2021), with a horizontal resolution of 9 km.

To achieve higher confidence and higher accuracy of the statistics for the results, a large-ensemble AGCM experiment was used to obtain a large number of cases to satisfy the requirements for analysis on interannual atmospheric variability. The output from the 100-member ensemble simulations was obtained from the “Database for Policy Decision-Making for Future Climate Change” (d4PDF; Mizuta et al., 2017). We used the historical climate experiment (HIST) between 1951 and 2011, with horizontal mesh size of 60 km, and the circulation-related output, with $1.25^\circ \times 1.25^\circ$ resolution. This experiment was forced using observed monthly mean SST and sea ice from the COBE-SST2 dataset (Hirahara et al., 2014) and other observed external forcing agencies such as greenhouse gases, aerosols, and ozone. The ensemble members were created using perturbed lower boundary conditions (i.e., small perturbations are added to SST based on SST error analysis) and initial conditions. The monthly mean and daily mean output from HIST were extensively analyzed in this study.

The land surface schemes in ERA5-Land and HIST (d4PDF) are the third generation models (Sellers et al., 1997), i.e. including the biophysical and carbon cycle (**Figure 1-15**). As a matter of interest, it should be noted here that either of the land surface scheme shown in **Figure 1-15**, with characteristics of the four layers to represent the soil, taking into account soil heat transfer, soil water cycle and their interaction with the atmosphere and ecosystems. More importantly, soil water phase change with temperature is considered, with associated processes such as: thermal effects associated with freezing/thawing, changes in soil water transfer and soil thermal conductivity, etc.

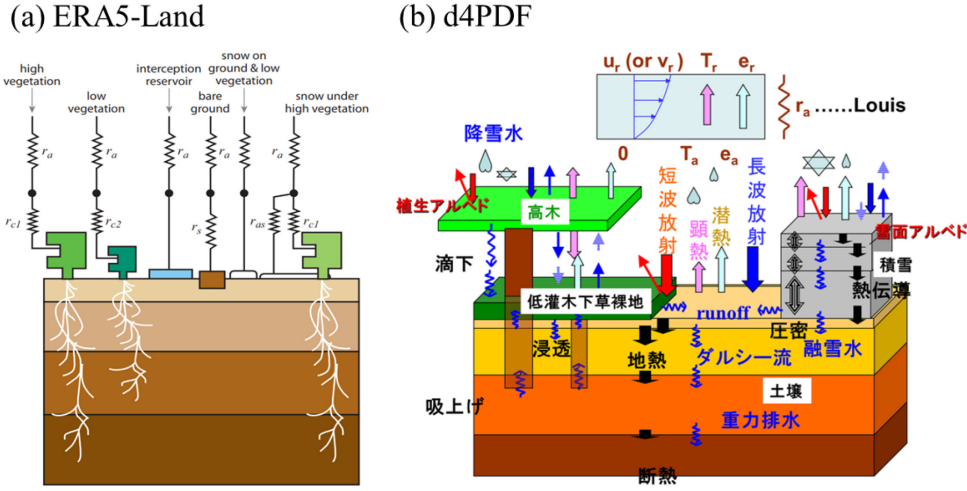


Figure 1-15 Land surface schemes used in data. Schematic charts of the land surface scheme in (a) ERA5-Land and (b) MRI-AGCM3.2. Figure (a) is taken from ECMWF documentation CY41R2 (<https://www.ecmwf.int/en/publications/ifs-documentation>) and figure (b) is taken from technical reports of the Meteorological Research Institute No.73 (https://www.mri-jma.go.jp/Publish/Technical/DATA/VOL_73/03.pdf).

(ii) Methodology

The apparent heat source (Q_1) based on thermodynamic energy equation was computed for both JRA-55 and HIST, based on their daily fields to diagnose the thermodynamic processes, following Yanai et al. (1973) and is defined as below:

$$Q_1 = c_p \left(\frac{p}{p_0} \right)^{\frac{R}{c_p}} \left(\frac{\partial \theta}{\partial t} + \vec{V} \cdot \nabla \theta + \omega \frac{\partial \theta}{\partial p} \right) \quad (1)$$

where θ is potential temperature, $\vec{V}$ is the velocity of the horizontal wind, ω is the vertical pressure velocity, R stands for the gas constant, c_p is the specific heat at constant pressure for dry air, and p is pressure ($p_0 = 1000$ hPa). The three terms within the parentheses on the right side of the equation represent temperature variation from local change ($\frac{\partial \theta}{\partial t}$), horizontal advection ($\vec{V} \cdot \nabla \theta$), and vertical diffusion ($\omega \frac{\partial \theta}{\partial p}$), respectively. Therefore, $\frac{Q_1}{c_p}$ represents the total diabatic heating, which includes heating caused by radiation, latent heating, surface heat flux, and sub-grid heat flux convergences (Hsu & Li, 2011). In areas dominated by monsoons, diabatic heating primarily arises from precipitation (condensation heating). Conversely, outside the regions experiencing active precipitation, diabatic heating is predominantly regulated by net radiative heating (Holton & Hakim, 2013).

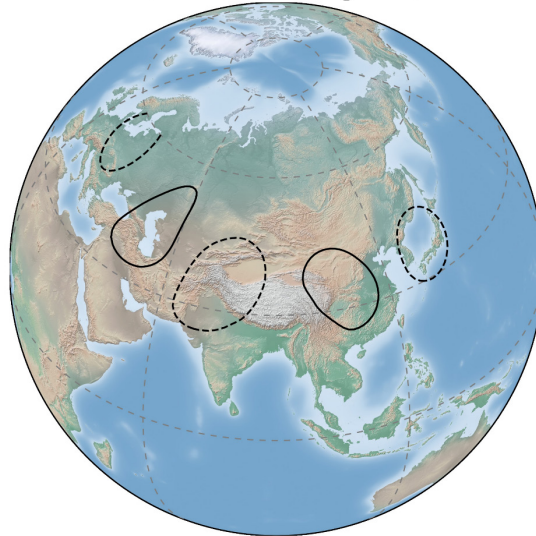
The wave activity flux described from K. Takaya & Nakamura (2001) is computed (hereafter; TN-WAF) to diagnose the stationary Rossby wave propagation:

$$\vec{W} = \frac{p \cos \varphi}{2|\vec{V}|} \begin{pmatrix} \frac{\bar{u}}{a^2 \cos^2 \Phi} \left[\left(\frac{\partial \Psi'}{\partial \lambda} \right)^2 - \Psi' \frac{\partial^2 \Psi'}{\partial \lambda^2} \right] + \frac{\bar{v}}{a^2 \cos \Phi} \left[\frac{\partial \Psi'}{\partial \lambda} \frac{\partial \Psi'}{\partial \Phi} - \Psi' \frac{\partial^2 \Psi'}{\partial \lambda \partial \Phi} \right] \\ \frac{\bar{u}}{a^2 \cos \Phi} \left[\frac{\partial \Psi'}{\partial \lambda} \frac{\partial \Psi'}{\partial \Phi} - \Psi' \frac{\partial^2 \Psi'}{\partial \lambda \partial \Phi} \right] + \frac{\bar{v}}{a^2} \left[\left(\frac{\partial \Psi'}{\partial \Phi} \right)^2 - \Psi' \frac{\partial^2 \Psi'}{\partial \Phi^2} \right] \\ \frac{f_0^2}{N^2} \left\{ \frac{\bar{u}}{a \cos \Phi} \left[\frac{\partial \Psi'}{\partial \lambda} \frac{\partial \Psi'}{\partial z} - \Psi' \frac{\partial^2 \Psi'}{\partial \lambda \partial z} \right] + \frac{\bar{v}}{a} \left[\frac{\partial \Psi'}{\partial \Phi} \frac{\partial \Psi'}{\partial z} - \Psi' \frac{\partial^2 \Psi'}{\partial \Phi \partial z} \right] \right\} \end{pmatrix}$$

where p is pressure, $\bar{u}$ and $\bar{v}$ are the basic state of the horizontal winds, set as climatological values in this study, $|\vec{V}|$ denotes the wind velocity computed from $\bar{u}$ and $\bar{v}$; Φ , λ , and z are latitude, longitude, and height, respectively, and Ψ' represents anomalies of the stream function. The f_0 and N are Coriolis parameter and Brunt-Väisälä frequency, respectively. The TN-WAF might be known as a “modified version” of the Plumb wave activity flux, which bolstering the meridional component, allowing for a better description of large-amplitude westerly jet Rossby long-wave disturbances in meridionally non-uniform airflow (Shi et al., 2017).

2

Mid-latitude forcing I: CGT



Connections between the Indian and East Asian summer monsoons bridged by the circumglobal teleconnection

Brief summary

The responses of the East Asian summer monsoon (EASM) to the Indian summer monsoon (ISM) have been the subject of extensive investigation. Nevertheless, it remains uncertain whether the ISM can serve as a predictor for the EASM. Here in this section, on the basis of both observations and a large-ensemble climate model experiment, we show that the subseasonal variability of abnormal diabatic heating over India subcontinent enhances precipitation over central East China, the Korean Peninsula, and southern Japan in early summer. ISM diabatic heating triggers Rossby wave propagation along the subtropical jet, promoting southerly winds over East Asia. The southerly winds help steer anomalous mid-tropospheric warm advection and lower-tropospheric moisture advection toward East Asia, providing conditions preferential for rainband formation. Cluster analysis shows that, depending on jet stream structures, ISM heating can serve as a trigger as well as a reinforcer of the rainband.

2. Connections between the Indian and East Asian summer monsoons bridged by the circumglobal teleconnection

2.1. Introduction

The East Asian summer monsoon (EASM) is one of the most prominent components of the Asian summer monsoon system. The EASM is characterized as a northeastward extension of a rainband that results in spatiotemporally continuous rainfall, affecting East China, the Korean Peninsula, and Japan in summer (Y. Ding & Chan, 2005; B. Wang & Ho, 2002). The rainband migrates northward from early summer to late summer (i.e., June–August) with the largest areal coverage and greatest amount of banded precipitation occurring in June (Jiabin Liu et al., 2022). As a subtropical monsoon, the EASM is influenced by multiple remote factors that originate from tropical (Nitta, 1987; B. Wang et al., 2000; R. Wu et al., 2003; Jianling Yang et al., 2007), mid-latitude (Horinouchi et al., 2019; Z. Wang et al., 2018; Xue et al., 2021), and even polar regions (X. Chen et al., 2021; Sung et al., 2006; Z. Wu et al., 2009). Thus, making predictions regarding the EASM rainband remains challenging (B. Liu et al., 2020; Nakamura & Sato, 2022; B. Wang et al., 2005) despite the increasing levels of attention focused on related extreme weather events that include droughts and floods (Easterling et al., 2000; Merryfield et al., 2020). Consequently, unanticipated EASM events could have devastating influence on agriculture, water resources, and human health in East Asian countries (Zhou et al., 2021) if appropriate mitigation strategies are not undertaken based on operational forecasts.

Another principal component of the Asian summer monsoon system is the Indian summer monsoon (ISM), which is independent of the EASM despite their recognized interaction (Q. Ding & Wang, 2005; Kripalani & Kulkarni, 2001; R. Zhang, 2001). The rainfall and resultant diabatic heating of the ISM trigger a Rossby wave train (Rodwell & Hoskins, 1996), known as circumglobal teleconnection (CGT) that is manifested as a leading mode of the summer circulation over mid-latitude Eurasia with zonal wavenumber-5 structure (Branstator, 2002; Q. Ding & Wang, 2005; Yasui & Watanabe, 2010). In CGT episodes, the 200-hPa geopotential height over East Asia, the North Pacific, North America, and the Northeast Atlantic are all nearly in phase and synchronized with the geopotential height variations over the ISM area, specifically over the northern Indian subcontinent and West-to-Central Asia. Through such covariation, the CGT affects the weather (Boers et al., 2019) and climate (Q. Ding & Wang, 2005; Kim et al., 2019; Mölg et al., 2014) across the Northern Hemisphere. Hence, it is reasonable to hypothesize that the CGT acts as a bridge in terms of precipitation variability between the ISM and the EASM. A better understanding of the ISM–EASM bridge could add value to EASM predictions.

Summer precipitation in these two monsoon regions does not have long-term correlation on the interannual time scale; their correlation appears only in certain periods or in different subregions other than in relation to the rainband in East Asia (Krishnan & Sugi, 2001; Lin et al., 2017; R. Wu, 2017). This unstable relationship is perhaps attributable to variations of the subtropical westerly jet, which acts as waveguide of the CGT wave train and which determined the location of CGT action of centers (Lin et al., 2017). It might also be related to the weakening of the ISM–El Niño relationship (Kumar et al., 1999), which has diminished the ISM and its associated CGT variability since the mid-1970s (Hong Wang et al., 2012). Evidence also shows that it is because the ISM might not be the primary trigger of the CGT, and which is controlled by the forces that are near the Mediterranean (Beverley et al., 2021; Yasui & Watanabe,

2010). Hence, there is currently no agreement on whether the ISM can be regarded as a robust predictor for the EASM (as discussed in [section 1.2.4](#), see which for more details). Considering the complex influencing factors of the Asian summer monsoon systems that operate across multiple time scales, treatments of monthly and seasonal averaging that used in previous study may have the effect of obscuring the dynamical relationship between the ISM and the EASM. Although two scenarios of ISM–CGT interaction (i.e., **Figure 1-7**) are proposed by Q. Ding & Wang (2007, 2009) through the subseasonal analysis, the subseasonal ISM–EASM or CGT–EASM connection remains unknown. Here, the analyses for ISM–EASM relationship are conducted at the subseasonal time scale, which is more appropriate for determining physical linkages between the ISM and the EASM.

2.2. Objectives

The question of whether CGT could bridge the ISM and EASM may lie primarily in the interacting mechanism between the ISM and CGT, potentially perturbed by the jet stream variation, and the SST forcing (i.e., ENSO), based on the discussion of [section 1.2.4](#) and [section 2.1](#).

Here, we depict the interaction of CGT, ISM, and EASM at subseasonal time scale, which is less noticed and understood, with further classify the subtropical jet patterns and ENSO conditions to discuss how the ISM induced signal differs in the EASM domain. This chapter ([chapter 2](#)) focused on the early summer season, specifically June, which is the month when the EASM rainband generally peaks and dominants in the East Asian sector (**Figure 1-4**).

2.3. Data and Methods

(i) Observation and simulation data

The observed leading precipitation interannual variability mode (1979 to 2020) was investigated using monthly data from GPCP version 2.3. The monthly mean atmospheric circulation and hydrological parameters were derived from the JRA-55. The same parameters but for simulation are from HIST, to confirm the robustness of the result. For the description of HIST see [section 1.5](#) for detail. The interannual analysis is not comprehensively presented here since it is neither in the main targets nor objectives, but it may provide vital information, hints, and evidence for ISM–EASM connection which inspired me to conduct the subseasonal analysis. Therefore, the interannual analysis is included briefly in the dissertation.

The subseasonal analysis was conducted over an extended period (1958–2020) to enlarge the sample size of episodes. The daily circulation and hydrological parameters are again from the JRA-55 reanalysis. The analysis also used Asian Precipitation—Highly Resolved Observational Data Integration Towards Evaluation of Water Resources (APHRODITE) version 2 (Yatagai et al., 2012) daily data for the period 1958–2015, which merged great amount of the rain-gauge data over the Asian sector. And again, to ensure the robustness of the result, the HIST experiment was used to

obtain a suitable number of cases to satisfy the requirements for analysis on atmospheric variability for subseasonal time scale as well.

The anomalies in this chapter ([chapter 2](#)) represent the deviation of the variables from their 11-year moving average of the same calendar month or Julian day based on data time scales to remove the decadal variability. Thereafter, an additional 5-day moving average was applied to the daily anomaly fields for subseasonal analysis.

(ii) Indices for tracking the tropical intraseasonal oscillation

The subseasonal cycle of ISM precipitation and associated heating is known as composed of the active phase and break phase which are controlled by the propagation of tropical intraseasonal oscillation (ISO; e.g., Chang et al., 2021; Q. Ding & Wang, 2007; X. Guo et al., 2021; Kikuchi, 2021). To investigate the role of tropical ISO in the subseasonal ISM–EASM connection, i.e., Madden–Julian Oscillation (MJO) and Boreal Summer Intraseasonal Oscillation (BSISO) indices are employed. The MJO and BSISO indices are employed from Kikuchi (2021), Kikuchi (2020), and Kikuchi et al. (2012), obtained via the International Pacific Research Center, University of Hawaii (https://iprc.soest.hawaii.edu/users/kazuoyosh/Bimodal_ISO.html). These indices are created by applying the extended EOF to the 25–90-day Lanczos bandpass filtered outgoing longwave radiation of boreal summer (June–October) and winter (from December to next April). The MJO mode and BSISO mode are defined using the first two extended EOFs for each season. Subsequently, the MJO index and BSISO index are derived by projecting the filtered OLR anomaly fields, consisting of the same three-time lags (-10 , -5 , and 0 days), onto the corresponding extended EOFs of each mode (Kikuchi, 2020).

(iii) Methodology

The Q_1 and TN-WAF are used here to investigate the ISM-released condensation heating and its associated consequent wave propagation (see [section 1.5](#) for details).

To investigate the subseasonal ISM–EASM relationship, we selected strong Indian monsoonal diabatic heating events. The abnormal Indian monsoonal diabatic heating events of June were collated based on the following criteria for both JRA-55 and HIST. The daily anomalies of the diabatic heating rate (Q_1/C_p ; unit: K/day) averaged over 10° – 30° N, 70° – 85° E (i.e., Indian subcontinent) were computed for each year from May 1 to July 31. The definition of the anomaly follows that in the previous sub-section. Subsequently, the vertical mass-weighted mean from 850 to 200 hPa gives the averaged ISM heating released to the troposphere as an index (ISMH). Because 1 standard deviation of daily ISMH in June for JRA-55 and HIST is 1.06 and 1.10 K/day, respectively, an abnormal ISM heating event was defined as the case when ISMH is greater than 1.00 K/day with a duration of 5 or more consecutive days. The first day of each event is referred to as the “event0” day, and the day observing the highest ISMH value during each event is designated as the “lag0” day (**Figure 2-1**). To highlight the event in June, the lag0 day must occur in June; otherwise, it is rejected. If multiple events are detected in a year, the event with the highest ISMH (i.e., the strongest heating anomaly) is adopted. Overall, 29 lag0 days and 2757 lag0 days were selected based on JRA-55 and HIST, respectively. Finally, composite analysis was performed to investigate the lead–lag relationship relative to the lag0 day.

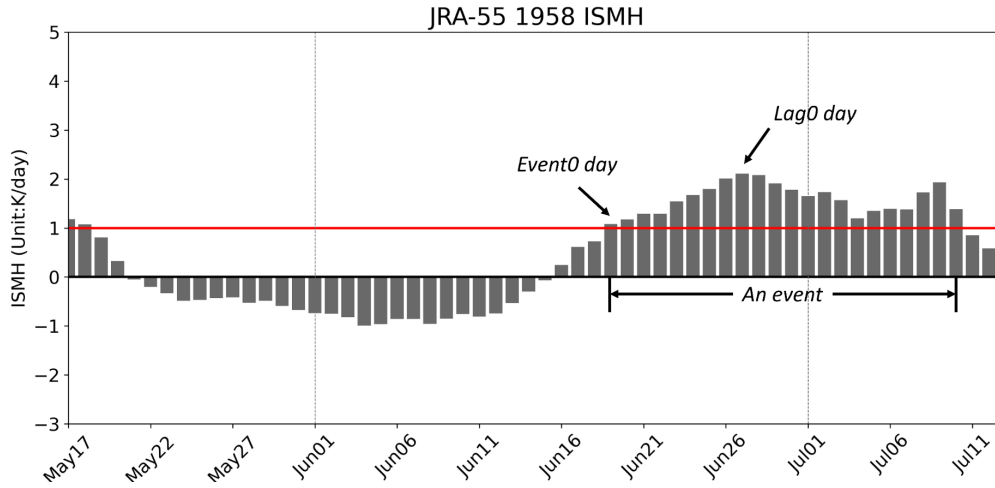


Figure 2-1 Schematic illustrating the method adopted for identification of strong diabatic heating over the Indian summer monsoon region. Bars are diabatic heating from 17 May 1958 to 13 July 1958 based on JRA-55, averaged over 10° – 30° N, 70° – 85° E and vertically mass-weighted from 850 to 200 hPa. Red line is the threshold (i.e., 1 K/day).

In addition, the CGT indices are calculated to measure the subseasonal variations of the CGT. The CGT reference pattern is created by applying EOF analysis to seasonal mean (June to September) meridional wind at 200-hPa over the domain of 20° – 60° N, 0° – 150° E, following previous study (Yasui & Watanabe, 2010; see [section 1.2.1](#) for detail). The leading patterns of meridional wind at 200-hPa from JRA-55 (1958–2020) and HIST (6100 years, i.e., 100 ensemble members with 61 years concatenated on time dimension) resemble each other (**Figure 2-2**). For the convenience, the HIST leading pattern is used as reference pattern. Then, composite daily 200-hPa meridional wind anomalies are projected onto this reference pattern to obtain the daily CGT index (hereafter, CGTI-PC1) for both JRA-55 and HIST, which represents the subseasonal variation of CGT wave pattern. Another CGT index (Q. Ding & Wang, 2005) was also computed for comparison, by area-averaging the 200-hPa geopotential height anomalies over 35° – 40° N, 60° – 70° E (hereafter, CGTI-DW).

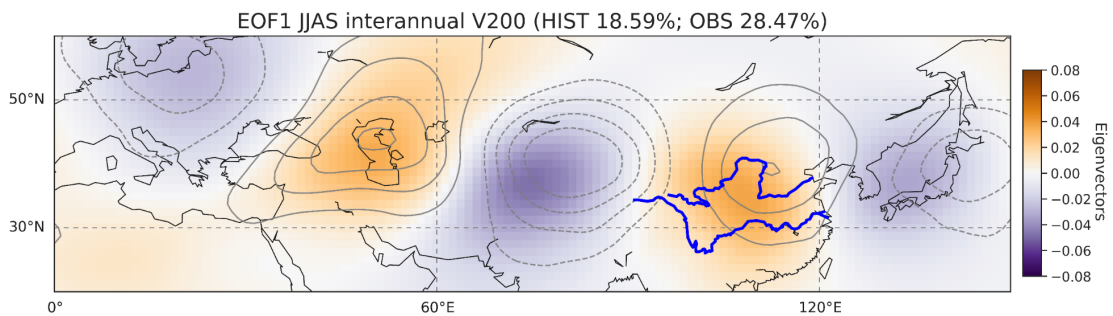


Figure 2-2 The leading EOF pattern of June to September meridional wind at 200-hPa. The first leading EOF eigenvectors of interannual June to September seasonal mean meridional wind anomalies at 200-hPa. The contour is the result for JRA-55 reanalysis (interval: 0.01) and the color shading is for the result of HIST. The blue curves are for two major river systems in China, the Yellow River (north) and the Yangtze River (south).

(iii) Linear baroclinic model experiments

To investigate the role of ISM heating on the CGT wave train over the Asian jet stream, the experiments using the linear baroclinic model is investigated (LBM; Watanabe & Kimoto, 2000), which is feasible to simulate a linear response to a prescribed forcing. The setting of LBM generally follows Watanabe & Jin, (2003), which described briefly as following. The LBM was set to the resolution of T42L20 (128×64 grids horizontally and 20 sigma levels vertically). Horizontal diffusion with an e-folding time scale of 2 hours was applied. We considered five scenarios, representing each month from May to September as basic states, using JRA-55 monthly mean climatological values from 1979 to 2020. For each scenario, we imposed the same composite June lag0 day ISM diabatic heating from observation (see **Figure 2-10**) over the 10° – 30° N, 70° – 85° E region, which was regirded from JRA-55 to T42L20 resolution (**Figure 2-3**).

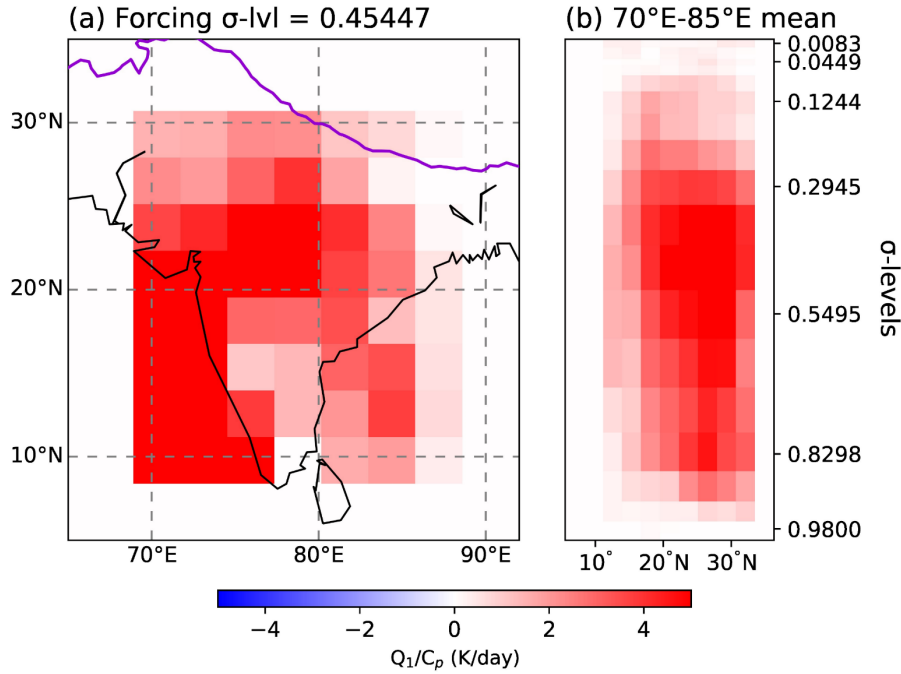


Figure 2-3 The forcing used in LBM. The diabatic heating pattern that was used as forcing in linear baroclinic model, (a) spatial pattern of diabatic heating at σ -level near 0.45 (mid-troposphere), and (b) cross section (70° – 85° E mean) of the diabatic heating pattern of the latitude-vertical profile.

2.4. Results

2.4.1. The leading precipitation mode of East Asian summer monsoon

The first leading mode of June interannual East Asian precipitation shows a clear dipole pattern with a wet (dry) band in the north (south) in both the observations (GPCP) and the large-ensemble historical climate model experiment (HIST; **Figure 2-4a and b**). Their spatial correlation (0.82; GPCP EOF1 versus HIST EOF1) suggests

that a realistic pattern of precipitation variability is captured by HIST. The boundary between the wet and dry bands extends from the lower reaches of the Yangtze River to southern Japan. The wet band has two centers: one near the Yangtze River Basin and the other extending from South Korea to South Japan, indicating regions with potential risk of flood events when the score of the first principal component (PC1) is high. The values of the explained variance and eigenvectors of first leading mode in HIST are lower than those observed, possibly owing to the finer resolution and greater number of samples of HIST in comparison with the observations (i.e., $\sim 0.56^\circ$ versus 2.5° , and 6100 years versus 42 years for HIST and the observations, respectively). The ensemble means of PC1 in HIST reflect the contributions of external forcing such as sea surface temperature (SST) to the dipole precipitation pattern (**Figure 2-4c**). Oceanic forcing has been shown to be important to constrain the EASM (e.g., B. Wang et al., 2000; R. Wu et al., 2003; Jianling Yang et al., 2007), indeed, positive shift (e.g., 2001) and negative shift (e.g., 1993) of the PC1 score for a 100-member ensemble is likely attributable to oceanic effects on the dipole precipitation mode. However, the HIST ensemble mean PC1 is not correlated with the PC1 of the observations (correlation coefficient: -0.05 during 1979–2011). The near-zero correlation coefficient indicates that the dipole precipitation pattern is strongly regulated by internal atmospheric variability.

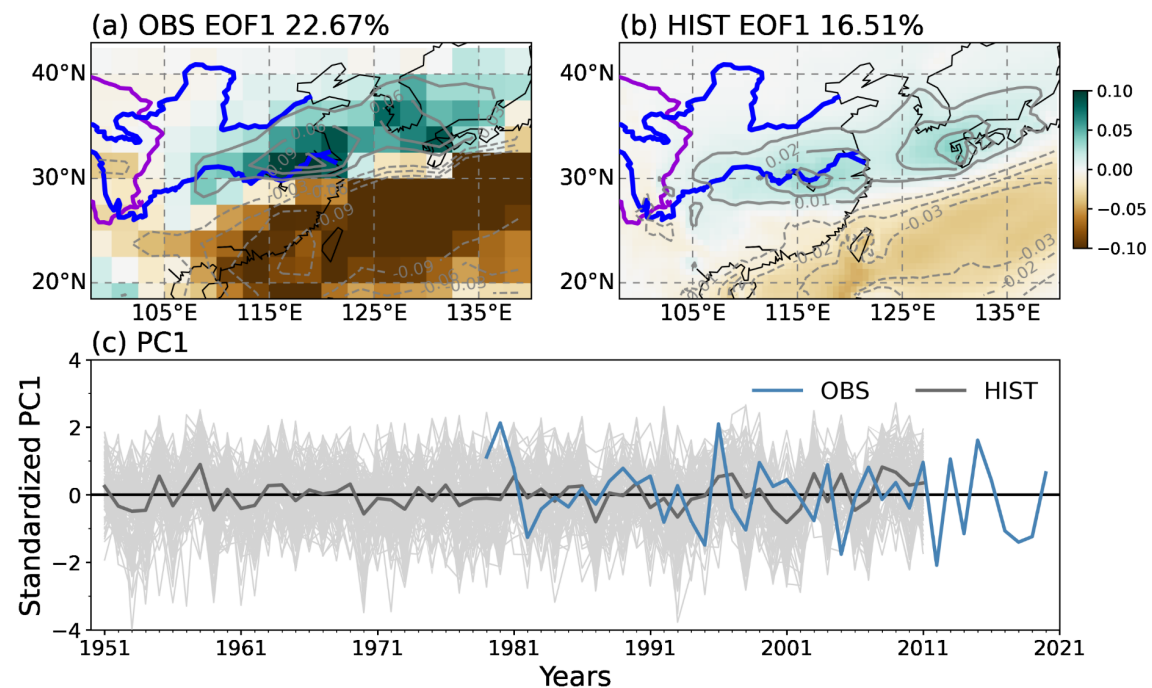


Figure 2-4 The leading pattern of June EASM. First leading mode of EOF analysis for June precipitation from (a) observations (GPCP), (b) HIST 6100 years (i.e., 61 years with 100 members), and (c) their standardized principal component (PC) for GPCP (blue line) and each ensemble member in HIST (gray lines). Gray contours denote the values of eigenvectors. Purple contour represents elevation of 2500 m. The dark blue curves represent the Yellow River (north) and the Yangtze River (south).

In observation, the June climatological rainband over East Asia that extends from South China to southern Japan is supported by strong southwesterly water vapor flux

(Figure 2-5a). The regression map of PC1 shows negative precipitation anomalies in this area, accompanied by a northeasterly component of water vapor flux anomalies that reduce the regional moisture supply (Figure 2-5b and c), whereas greater precipitation anomalies are evident over the Yangtze River Basin, Korean Peninsula, and southern Japan, fueled by strengthened moisture transport. Notably, Indian precipitation and moisture flux anomalies are significantly correlated with PC1 (Figure 2-5d), despite the fact that the EOF domain does not encompass the Indian region (Figure 2-4). Intensification of low-level southwesterly winds over India is an indicator of ISM onset (B. Wang et al., 2009). Therefore, Figure 2-5 depicts a potential connection between ISM onset and the EASM rainband in June. The simulated results from HIST ensemble simulation is similar to those of observation (Figure 2-6).

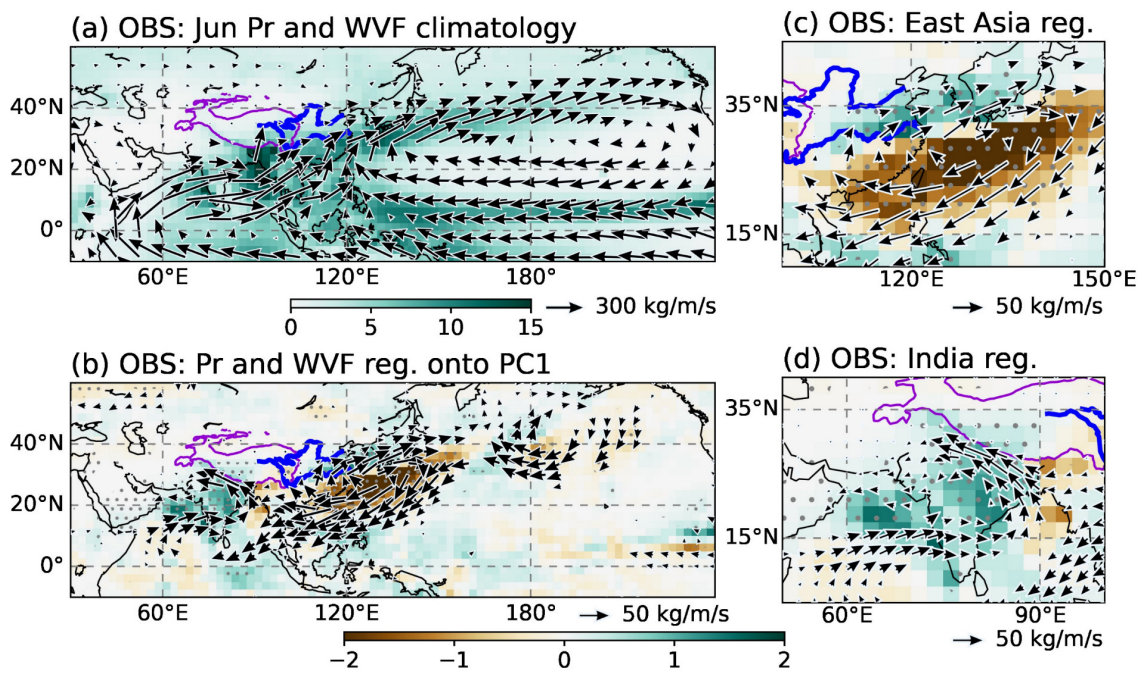


Figure 2-5 The observed rainfall pattern of June climatology and PC1 regression. Simulated June (a) climatological precipitation (color shading, unit: mm/day) and vertical integrated water vapor flux (arrows, unit: kg/m/s), and (b) anomalies of precipitation and water vapor flux regressed onto PC1. (c) and (d) The zoomed PC1 regression for East Asia and the Indian region, respectively. Dotted regions and black arrows in (b), (c), and (d) denote anomalies significant at the 90% confidence level based on the two-sided Student's t test (insignificant arrows are omitted).

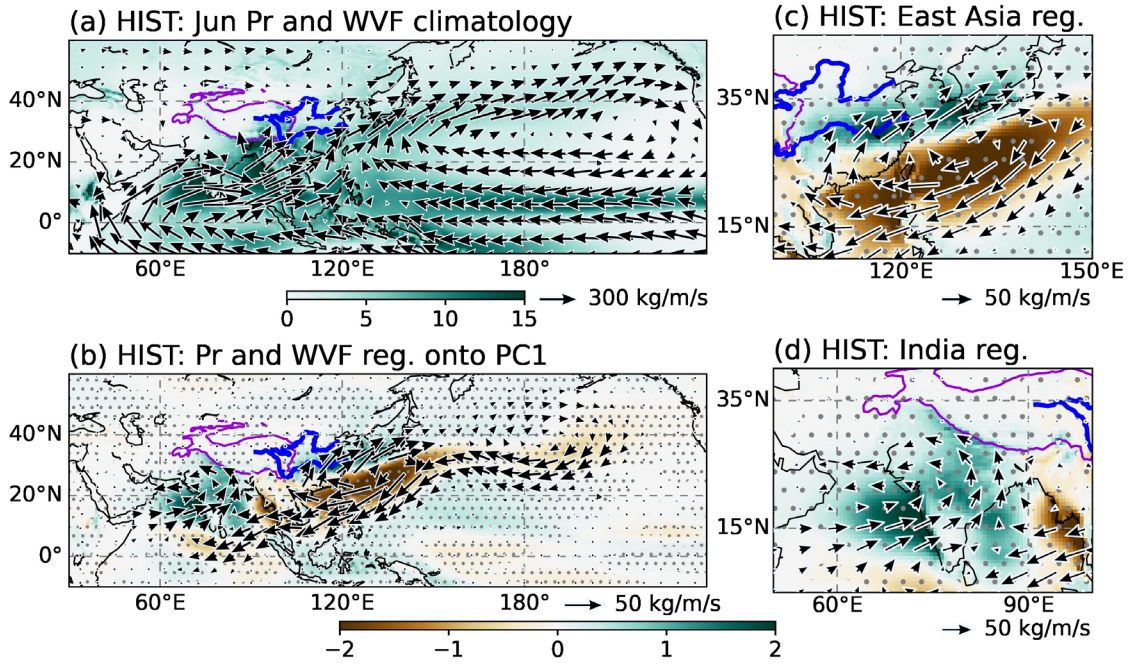


Figure 2-6 The simulated rainfall pattern of June climatology and PC1 regression. As in **Figure 2-5**, but for HIST with significant test of 99% confidence level.

The observed diabatic heating rate (Q_1/C_p) reveals the relationship between the ISM and PC1. In **Figure 2-7**, regions with strong diabatic heating at 500 hPa are generally consistent with the precipitation anomalies shown in **Figure 2-7b**. Strong diabatic heating is seen throughout the troposphere over the Indian region, and the heating is particularly strong near the mid-troposphere (**Figure 2-7c**). Furthermore, the diabatic heating rate is negative within the atmospheric boundary layer below 850 hPa, which reflects the effect of evaporative cooling in part of the convective processes (K. Yang et al., 2004). This heating structure over the Indian subcontinent reflects a typical onset and post-onset feature of the monsoonal processes (Johnson & Ciesielski, 2002). Results from HIST are consistent with those of observation (**Figure 2-8**). Meanwhile, no clear signal is found over the Tibetan Plateau for either observation or HIST. Hence, it is hypothesized that the monsoonal process over the Indian region (especially the heating) is the key factor that controls the dipole precipitation over East Asia. The following analysis elucidates the lead-lag relationship between them at the subseasonal time scale relative to the ISM diabatic heating.

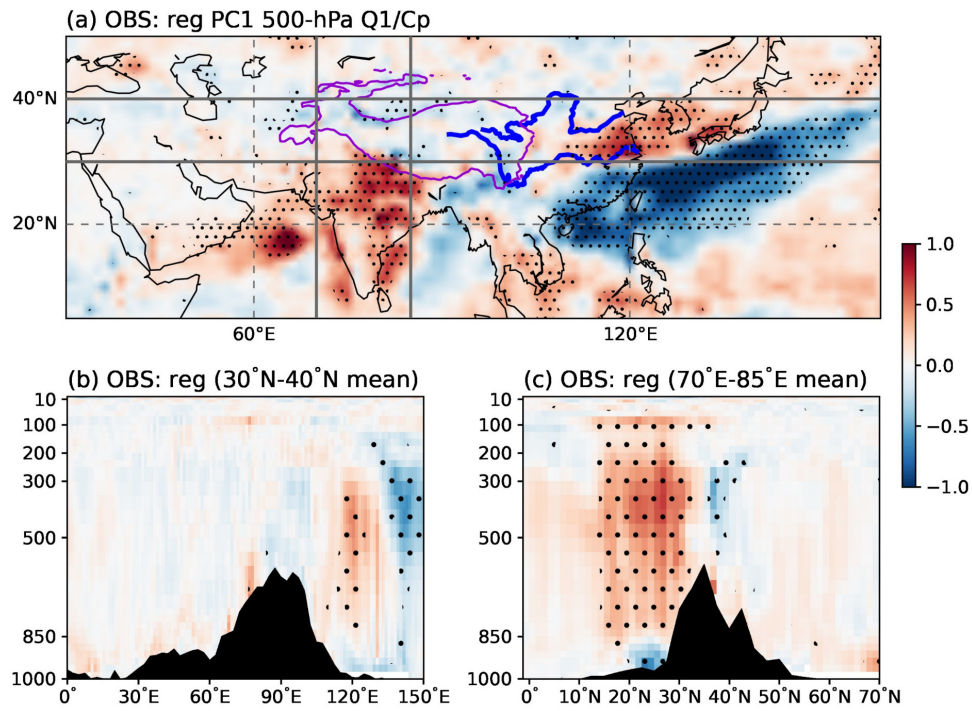


Figure 2-7 The observed diabatic heating pattern of PC1 regression. Diabatic heating rate (unit: K/day) regressed on the standardized PC1 from observation, for (a) spatial distributions at 500-hPa and (b) cross sections of the zonal mean over 70°–85°E and (c) cross sections of the meridional mean over 30°–40°N. Gray lines indicate the domain of the cross sections. Dotted regions denote anomalies significant at the 90% confidence level based on the two-sided Student's t test.

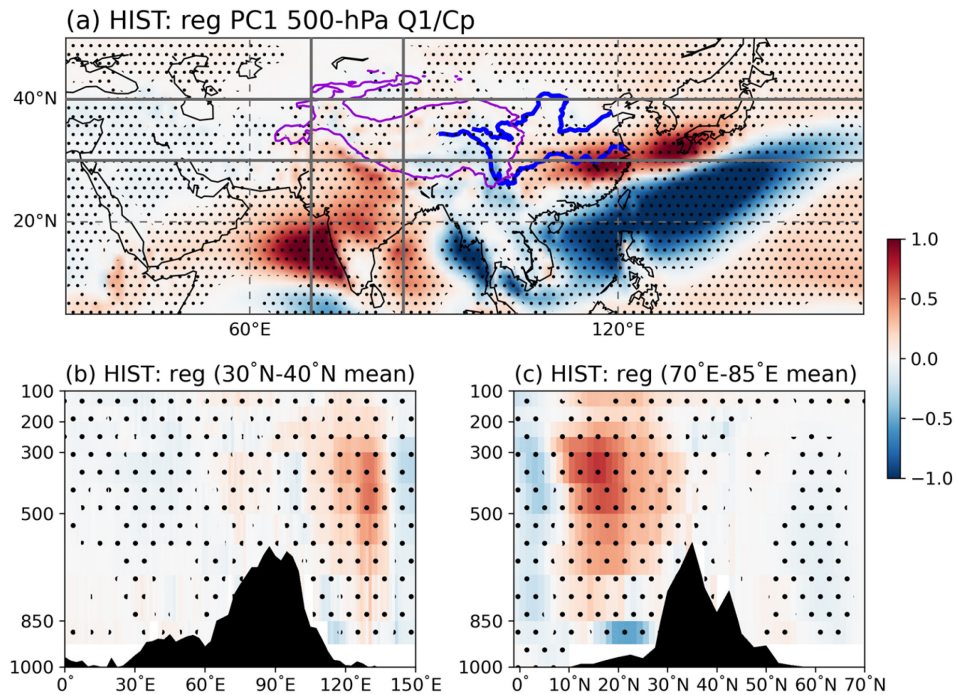


Figure 2-8 The simulated diabatic heating pattern of PC1 regression. As in Figure 2-7, but for HIST with significant test of 99% confidence level.

2.4.2. Subseasonal synchronization

In this section, the ISM–EASM causal connection is examined through lead–lag composite analysis with respect to abnormal ISM heating on the subseasonal time scale.

In HIST, the ISM diabatic heating rate increases from lead of 1–2 days to lag of 0 days, thereby reaching a peak, and finally decays to retain weak positive heating from lag of 1–2 days to lag of 7–8 days corresponding to weakening of the Indian precipitation anomalies (**Figure 2-9**). The life cycle of the abnormal ISM activities is approximately 3 pentads (**Figure 2-9** and **Figure 2-11**). From lead of 1–2 days to lag of 1–2 days, the strong ISM diabatic heating triggers Rossby waves at 200-hPa that originate from the Indian subcontinent and terminate over East Asia, forming the CGT-like wavy pattern over the mid-latitudes of Eurasia (arrows in **Figure 2-9**, TN-WAF). Coincident with the center of action of the CGT, East Asia is covered by an anticyclone at 200-hPa during these days, although the wave activity flux later vanishes (**Figure 2-9**; lag of 7–8 days). Meanwhile, a dipole-like precipitation pattern is found in East Asia, with a wet (dry) band in the north (south). The wet band over East Asia is intensified gradually after the lag0 day and it persists over nearly 2 pentads. The dry band, conversely, exists before the lag0 day and has small fluctuations over time. We therefore conjecture that the dry band may not be influenced by ISMH (**Figure 2-9** and **Figure 2-11**).

As for the observations, the result supports that of HIST, except that the wet band begins to form on the lag0 day and becomes more pronounced over the following days (**Figure 2-10** and **Figure 2-11**).

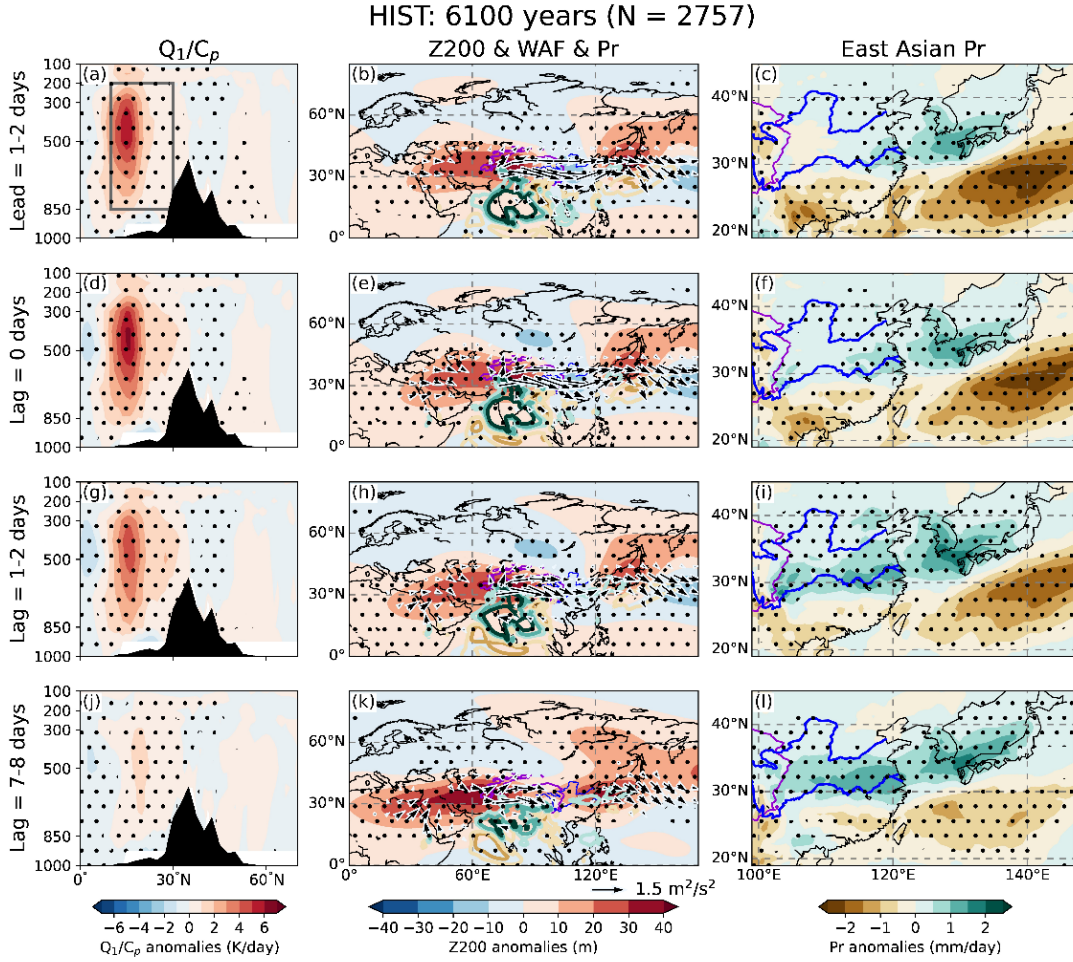


Figure 2-9 Simulated lead-lag characteristics of circulation and precipitation relative to abnormal Indian summer monsoon heating. Lead-lag composite at (a–c) lead of 1–2 days, (d–f) lag of 0 days, (g–i) lag of 1–2 days, and (j–l) lag of 7–8 days based on large ensemble historical climate experiment (HIST). (a, d, g, j) Cross-section of diabatic heating rate anomalies averaged over 70°–85°E, (b, e, h, k) 200-hPa geopotential height anomalies (shading), TN-WAF (arrows), and precipitation anomalies (contours, unit: mm/day; with intervals of ± 1 , ± 2 , and ± 4 , positive values in green contours while negative values in brown contours), and (c, f, i, l) precipitation anomalies over East Asia. The number on the top (N = 2757) represents the number of samples used for the composite. Dotted areas denote anomalies significant at the 99% confidence level based on the two-sided Student's t test.

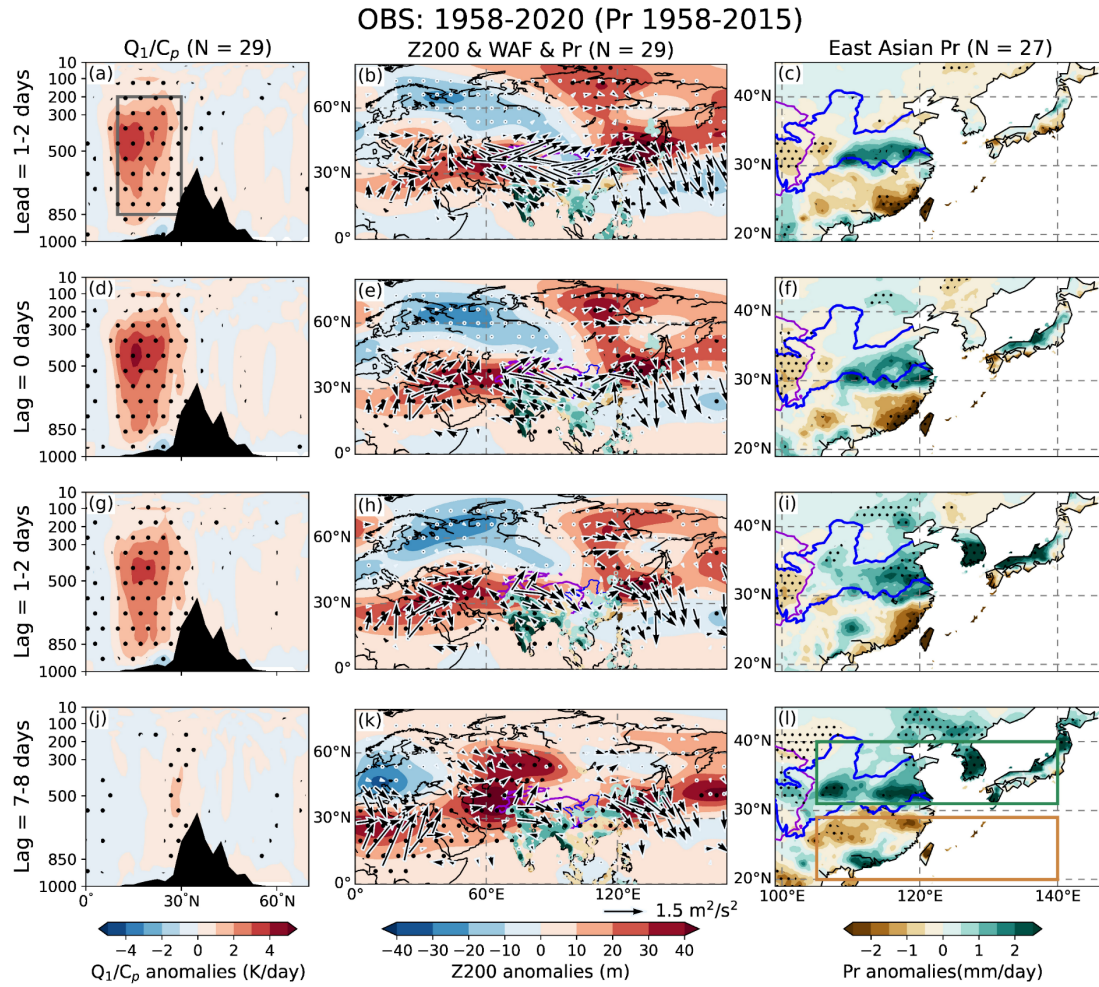


Figure 2-10 Observed lead-lag characteristics of circulation and precipitation relative to abnormal Indian summer monsoon heating. As in Figure 2-9, but for the observations with significant test at the 90% confidence level. The green and brown rectangles indicate the wet and dry band areas, respectively, used in Figure 2-11.

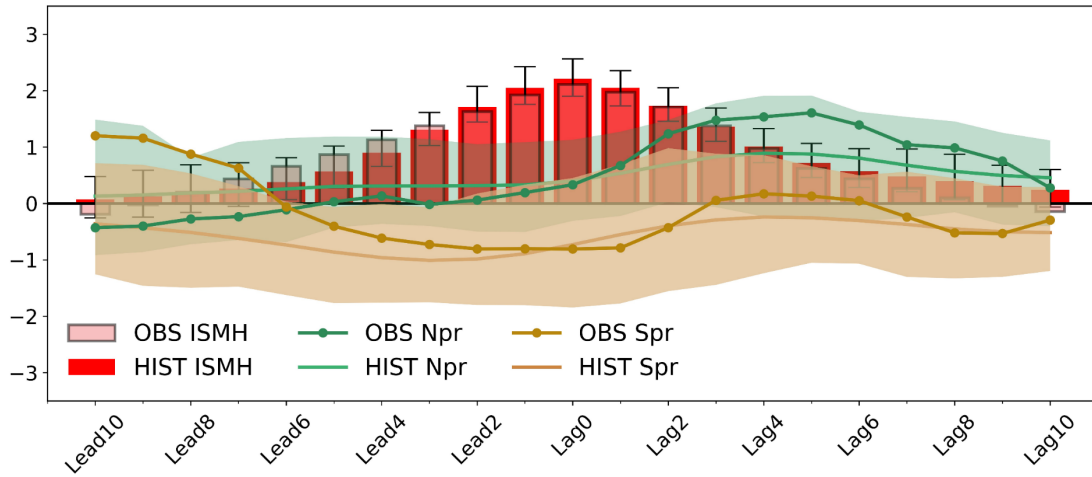


Figure 2-11 The lead–lag variations of ISMH and associated EASM precipitation. Lead–lag composite time series of diabatic heating anomalies (unit: K/day) averaged over 10° – 30° N, 70° – 85° E and vertical mass-weighted from 850 to 200 hPa (i.e., ISMH; [section 2.3](#)) and area-averaged precipitation anomalies (unit: mm/day) for wet band area (Npr) and the dry band area (Spr). The areas for Npr and Spr are indicated in Supplementary Figure 1. The error bars and shading represent the 100–member ensemble range obtained from large ensemble historical climate experiment (HIST).

2.4.3. The role of CGT as a bridge and its periodicity

In this section, how CGT is able to bridge the ISM and the EASM in June is depicted. The mean variations of two CGTIs (CGTI–PC1 and CGTI–DW; see [section 2.3](#) for detail) from lead of 20 days to lag of 20 days are shown in **Figure 2-12a and b**. They both show a bimodal M-shaped (i.e., peak-valley-peak) time variation, meaning CGT activity is strengthened in pre- and post-lag0 day. Power spectrum analysis for CGTI–PC1 shows a peak around 0.06 – 0.08 day^{-1} which corresponds to 10–20-day, suggesting that CGT’s main periodicity is close to the timescale of biweekly oscillation (**Figure 2-12c and d**). The significant power peak (i.e., exceeding the 95% red noise significance limit) for each selected event falls in the 5–25-day range (**Figure 2-12e**), with a mode at 0.07 day^{-1} (approximately 14-day, **Figure 2-12f**). These results suggest that the CGT behaves as a quasi-biweekly oscillation.

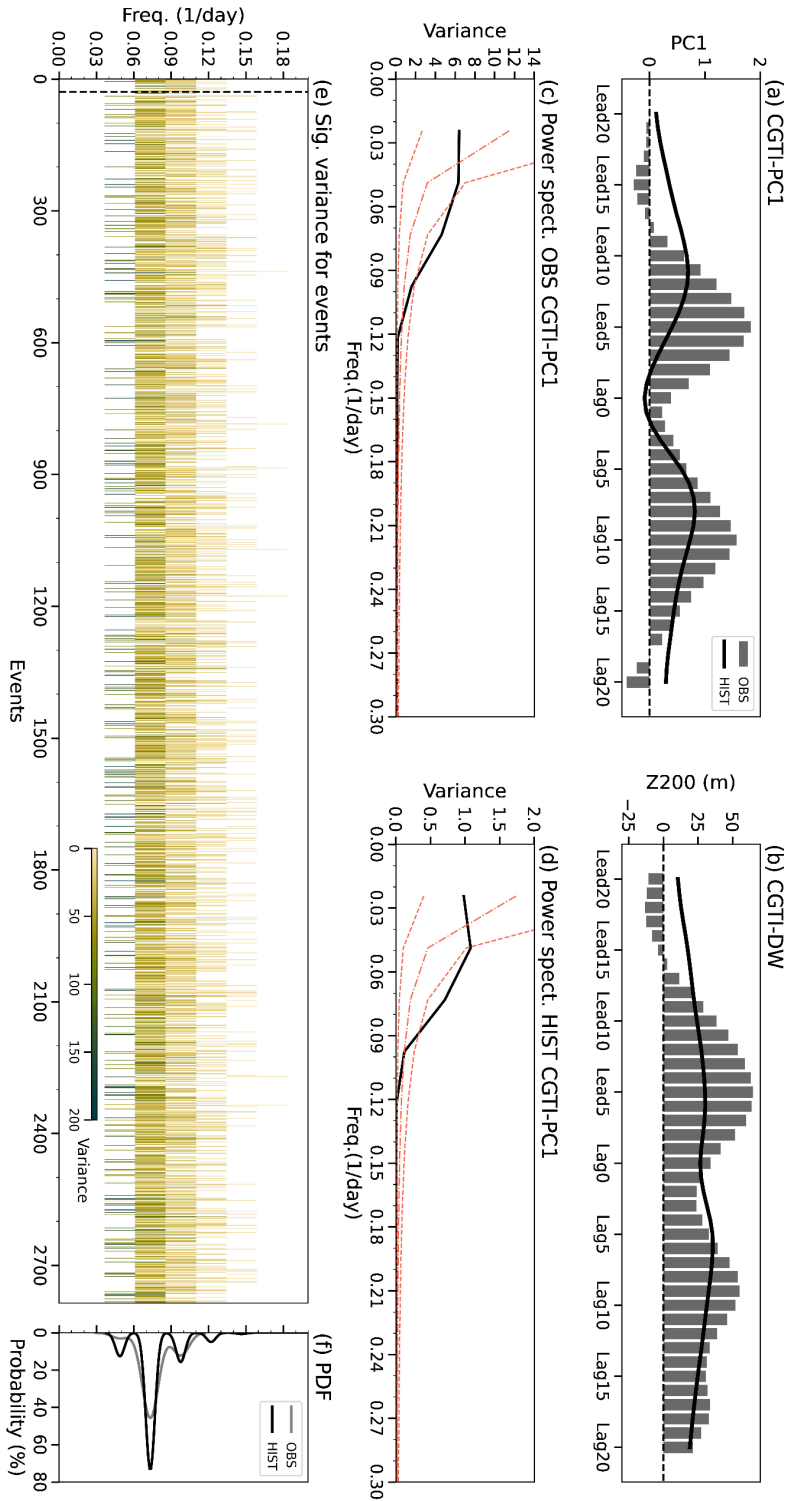


Figure 2-12 The periodicity of the subseasonal variation of the CGT. Time variation of the circumglobal teleconnection (CGT) from lead of 20 days to lag of 20 days represented by (a) CGT index based on PC1 of June to September interannual 200-hPa meridional wind (CGT-PC1) and (b) CGT index based on 200-hPa geopotential height averaged over 35°–40°N, 60°–70°E (CGT-DW). Power spectrum analysis of (c) CGT-PC1 of observation and (d) CGT-PC1 of large ensemble historical climate experiment (HIST) from (a), respectively, where the dash-dot and dashed red curve denote the red noise and its 95% significance boundary. (e) A list of the spectral variance values for each event of CGT-PC1. The range of significant frequency (day⁻¹) exceeding the 95% red noise confidence interval from the power spectrum are plotted, and the color shading represents the spectral variance. The vertical dash line is for the demarcation of the events collected from observation (left) and HIST (right). (f) probability distribution function (PDF) of the frequency based on (e).

Figure 2-13 illustrates the time variations of the spatial pattern of CGT. During the first peak of the M-shaped CGT subseasonal variation, there is a wave train running from western Europe toward the ISM domain. At this stage, the intensity of the wave train is rather faint over East Asia (**Figure 2-13a** and **b**). Near lag0 day, the vigorous ISM diabatic heating forces a local wave train to East Asia, but which is orthogonal to the CGT (**Figure 2-13c** and **d**), as denoted by the valley of the M-shaped CGT subseasonal variation in **Figure 2-12a** and **b**. Few days later, the CGT emerges again but much more energetically in the East Asian region (**Figure 2-13e** and **f**). According to the LBM experiment using June basic state, the ISM-induced wave train well overlaps the CGT wave train over East Asia on lag of 10 days, suggesting their local resonance along the westerlies (**Figure 2-13f**). The transition of the wave pattern between lag of 0 days and lag of 10 days over East Asia is proposed to be linked to ISMH activities as a similar transition pattern was also seen in the LBM simulation (**Figure 2-14**).

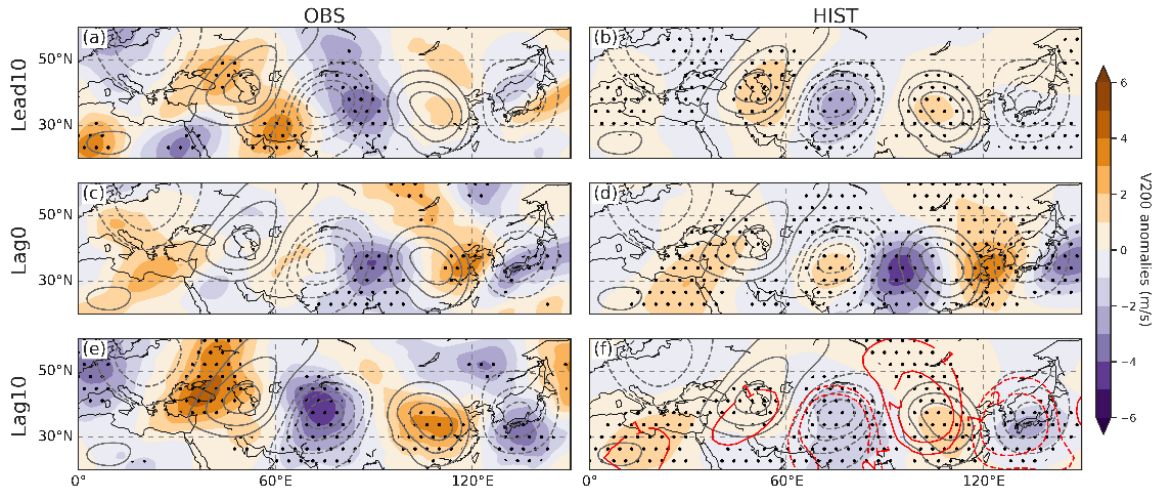


Figure 2-13 Precedent and subsequent circumglobal teleconnection (CGT) wave train relative to abnormal Indian summer monsoon heating. The shading indicates the composite 200-hPa meridional wind anomalies at (a) lead of 10 days, (b) lag of 0 days, and (c) lag of 10 days for observation and for large ensemble historical climate experiment (HIST; b, d, f). Red contours in (f) are from linear baroclinic model (LBM) experiment using June basic state, averaged from 30 to 50 days when LBM integration reached equilibrium. Grey contours show the eigenvectors of HIST first leading mode of interannual 200-hPa meridional wind during June to September (CGT pattern). Dotted regions denote anomalies significant at the 90% and 99% confidence level for the observation and HIST, respectively.

Overall, the results suggest that CGT wave train interacting with ISM diabatic heating exhibits a character of quasi-biweekly oscillations. The first phase is the propagation of wave train from western Europe, far upstream of the ISM domain. The second phase is the onset of the ISM convection and diabatic heating which excites the local wave train and maintains propagation to East Asia, which reinforces the CGT wave train due to the local resonance along the westerlies (**Figure 2-14**). In this instance, the ISMH activities might be related to European-CGT wave train propagation to ISM domain, as illustrated in the previous study (Q. Ding & Wang, 2007, 2009).

Finally, the wave train takes on a CGT-like pattern as it extends horizontally along the westerly jet, with the decaying the forcing from the ISM heating. One may notice this condition leads to the intensification of southerly wind over East Asia, in particular near the wet band area (**Figure 2-14**). The role of ISM-induced wave train and the southerly wind on East Asian precipitation will be elaborated in the next section.

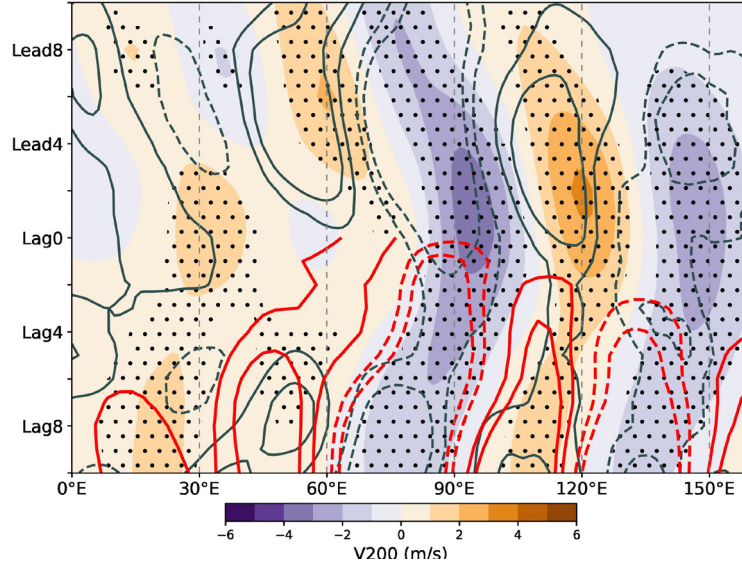


Figure 2-14 Hovmöller diagram of the CGT wave train relative to abnormal Indian summer monsoon heating. The Hovmöller diagram of the composite 200-hPa meridional wind anomalies from lead of 10 days to lag of 10 days (y-axis) averaged from 25°–45°N based on HIST (shading), JRA-55 (dark blue contours, with interval of ± 1 and ± 2), and LBM experiment (red contours, with interval of ± 0.5 and ± 1), with longitude range of 0°–180°E. Dotted regions denote anomalies significant at 99% confidence level for HIST. The green vertical lines are for the zonal extent of the wet band area (105°–140°E; **Figure 2-10**).

2.4.4. The influence of the wave guide

In this section, the interaction of CGT and East Asian wet band is analyzed based on their temporal evolution following the emergence of the Indian heating (i.e., event0 day; see [section 2.3](#)). To examine how ISM diabatic heating induced wave train is influenced by the preconditioning waveguide and its potential influence on the ISM–EASM connection, i.e., the jet patterns were classified. The jet axes of event0 day obtained from both JRA-55 and HIST were input together in the K-means clustering (Lloyd, 1982). The methodology to obtain the jet axis was as follows. First, zonal wind at 200-hPa (U200) was reconstructed by adding U200 anomalies at the event0 day and their corresponding pentad (± 2 -day) climatological U200 (**Figure 2-15a and b**). Second, the jet axis was identified as the latitudinal location of the maximum reconstructed U200 at each longitude within the range of 20°–50°N. This domain setting was intended to avoid potential influence by the polar front jet that appears in higher latitudes (Kong & Chiang, 2020). The input vectors to the K-means clustering included the values of latitude and longitude of the jet and the reconstructed U200 along the jet axis over the sector 40°–140°E (Fig. 2.4.4–1c; black box). Finally, we created

four clusters of the jet pattern at the event0 day as the optimal configuration based on the “elbow method.” We also input the same data into principal component analysis, and result indicates the population mean characteristics of four clusters are likely to be orthogonal to each other (**Figure 2-16**).

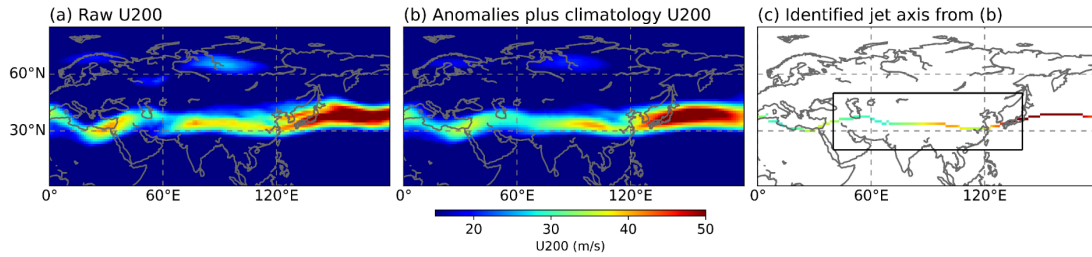


Figure 2-15 Procedures to obtain a jet axis. Schematic showing an example of 200-hPa zonal wind based on (a) original data, (b) reconstructed data, and (c) identification of the jet axis from (b).

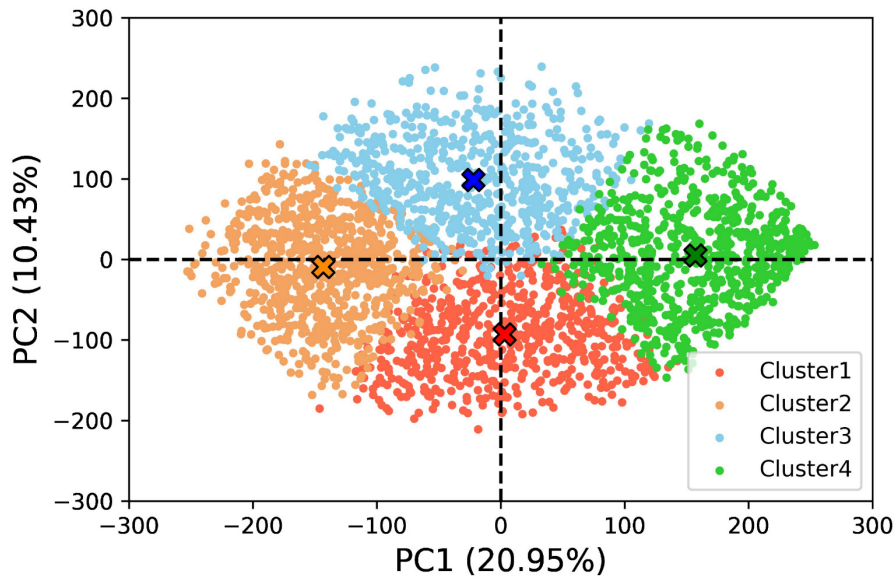


Figure 2-16 Orthogonality of the K-means clusters. Comparison of four clusters about the jet pattern created by K-means clustering and two leading modes from principal component analysis. The percentage values mean the proportion of explanation for PC1 and PC2. The crosses represent the group mean of each cluster.

Subsequently, 4 clusters by K-means clustering are applied to elucidate how ISM forcing propagates along the waveguide and stimulate the East Asian rainband. Clusters 1 and 4 exhibit a strong jet core over East Asia, while clusters 2 and 3 show a wavy jet stream with deacceleration near East Asia in comparison with clusters 1 and 4 (**Figure 2-17**). The clustered East Asian jet patterns are also found related to the jet meridional displacement, i.e., in clusters 1 and 4 the East Asian jets are displaced to north, with corresponding north-positive and south-negative zonal wind anomalies extending from the upper troposphere to near surface (**Figure 2-18**). In contrast in clusters 2 and 3, the

jets are displaced to south with associated reversed zonal wind anomalies in the troposphere. The following discussion is mainly based on the HIST owing to sufficient samples.

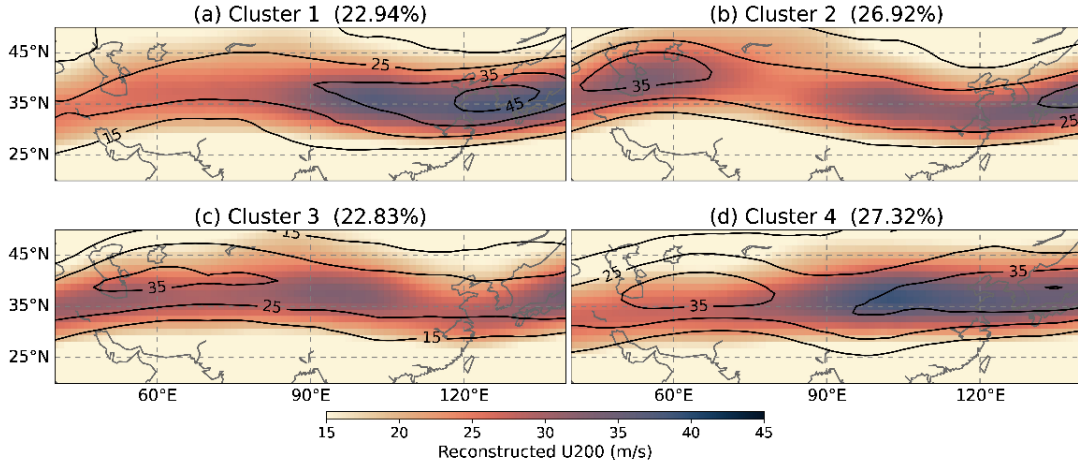


Figure 2-17 Clustered jet stream patterns. The reconstructed 200-hPa zonal wind (jet stream) patterns for (a) cluster 1, (b) cluster 2, (c) cluster 3, and (d) cluster 4 created by K-means clustering. Color shading represents values from HIST, and black contours represent values from JRA-55 reanalysis. The value in the parenthesis denotes the occurrence frequency obtained from all 2786 samples (i.e., 2757 samples from HIST and 29 samples from JRA-55 reanalysis).

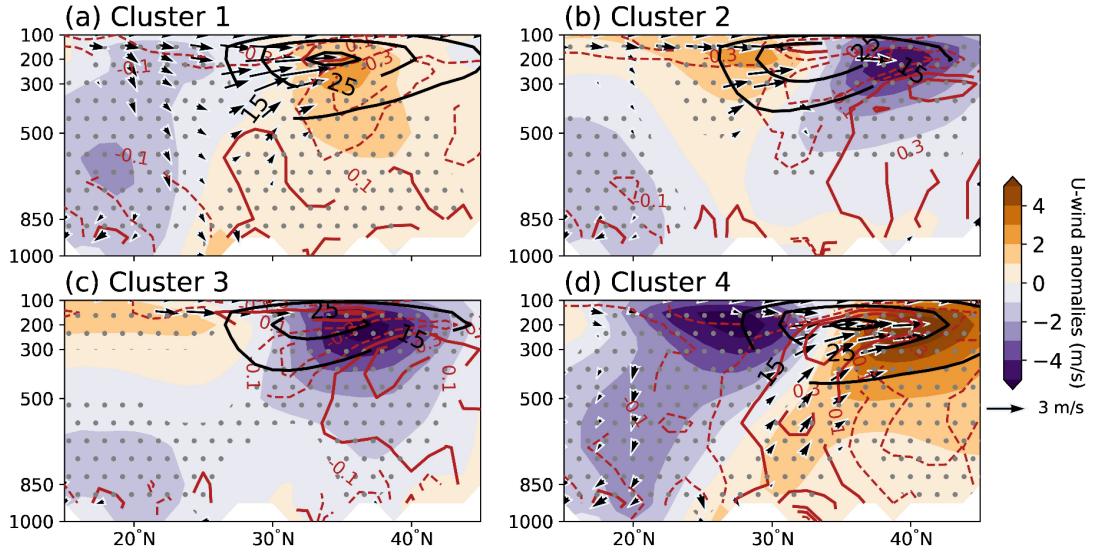


Figure 2-18 Vertical structure of clustered jet stream patterns. Cross section along 120°E showing the anomalous wind field based on HIST for (a) cluster 1, (b) cluster 2, (c) cluster 3, and (d) cluster 4 at event0 day. The color shading represents zonal wind anomalies, the arrows indicate anomalous meridional wind with vertical velocity, and the black and red contours depict reconstructed zonal wind velocity corresponding jet patterns (unit: m/s) and temperature advection anomalies (unit: K/day), respectively. Stippled regions and arrows indicate anomalies that are statistically significant at the 99% confidence level based on the two-sided Student's t test.

Similar to **Figure 2-9**, lead–lag composite analyses are conducted here but for each of the cluster. At lag of 3–4 days, all clusters exhibit a dipole precipitation pattern over East Asia (**Figure 2-19e to h**). The wet band is shown to pre-exist in clusters 1 and 4 and is intensified after the lag0 day (**Figure 2-19i to l**). Conversely, for clusters 2 and 3, the wet band can be activated by ISM heating. The wet band is closely related to ISMH activities, as illustrated in **Figure 2-9**, **Figure 2-10**, and **Figure 2-11**. The strength of ISMH in each cluster is nearly equivalent at lag0 day (approximately 2.20 K/day; **Figure 2-19a to d**). Therefore, the difference in the atmospheric fields among the clusters is likely attributable to the temporal evolution of the dynamical processes that include interaction between the subtropical jet and ISM heating, as well as the effects from other remote forcings that can be captured by the jet. Strong abnormal ISM heating events is connected to anomalous 200-hPa CGT-like circulation patterns in all clusters at lag0 day but with slight differences (**Figure 2-19e to h**). The CGT pattern in clusters 1 and 4, shows a strong cold trough located near Mongolia and extending to the north of the Yangtze River Basin, whereas the other clusters show troughs located near the Yangtze River Basin. These differences can be attributed to differential westerly jet stream pattern among clusters. The vigorous ISM diabatic heating near lag0 day perturbs the air in proximity to the westerly jet, serving as a Rossby wave source. The westerly jet stream acts as a waveguide (Hoskins & Ambrizzi, 1993; see discussion in [section 1.2](#) for detail), which confine the wave and direct it toward East Asia. Our findings indicate that clusters 2 and 3 exhibit a southward bending of the jet stream over East Asia (i.e., East Asian jet displaced to south), resulting in a southward motion of the associated wave train and the formation of an anomalous cold center (negative geopotential height anomaly) over East Asian regions on lag0 day (**Figure 2-19** and **Figure 2-20**). The cold center is consequently positioned southward by the jet (south of 35°N). Conversely, clusters 1 and 4 demonstrate relatively flat jet streams in comparison to clusters 2 and 3 (i.e., East Asian jet displaced to north), resulting in a predominantly zonal propagation of the wave toward East Asia. This dynamical constraint leads to the cold anomaly center being positioned further north in the East Asian sector (**Figure 2-20a and d**). The waveguide effect is also confirmed by stationary Rossby wavenumber (K_s ; Hoskins & Ambrizzi, 1993), which is calculated based on the jet pattern of **Figure 2-17**. The dominant wavenumbers are longer in clusters 2 and 3 while shorter in clusters 1 and 4 (**Figure 2-21**). Hence, the CGT pattern and the location of the downstream trough differ between clusters (**Figure 2-19**).

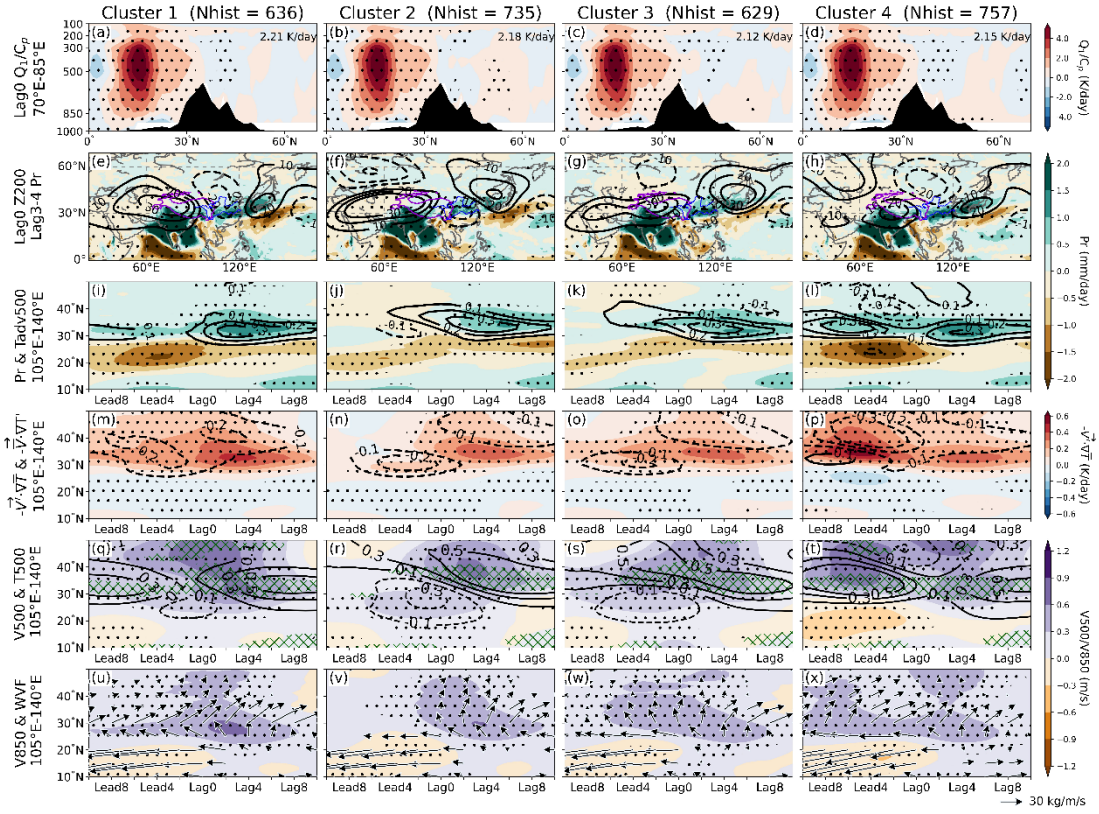


Figure 2-19 Simulated lead-lag dynamical characteristics relative to abnormal ISM heating in each cluster. (a–d) Composite of diabatic heating rate anomalies averaged over 70°–85°E at lag of 0 days, where 850–200 hPa mass-weighted and 10°–30°N averaged values (i.e., ISMH) are displayed on upper right, and (e–h) composite of precipitation anomalies (shading) and 200-hPa geopotential height anomalies (contours unit: m) at lag of 3–4 days. Temporal evolutions of the zonal mean fields averaged over 105°–140°E computed from lead of 10 days to lag of 10 days of (i–l) zonal mean precipitation (shading) and 500-hPa temperature advection (contours, unit: K/day) anomalies, (m–p) 500-hPa temperature advection anomalies computed by anomalous wind field with climatological temperature gradient (shading) and climatological wind field with anomalous temperature gradient (contours, unit: K/day), (q–t) 500-hPa meridional wind (shading) and temperature (contours, unit: K) anomalies, and (u–x) 850-hPa meridional wind (shading) and vertical integrated water vapor flux (arrows) anomalies. The green hatching in (q–t) indicates vertical velocity (unit: Pa/s) lower than 0, significant at the 99% confidence level based on the two-sided Student's t test. (a, e, i, m, q, u) cluster 1, (b, f, j, n, r, v) cluster 2, (c, g, k, o, s, w) cluster 3, (d, h, l, p, t, x) cluster 4. The stippled regions denote anomalies significant at the 99% confidence level based on the two-sided Student's t test. The number of samples used for the composite for each cluster is shown on the top (Nhst).

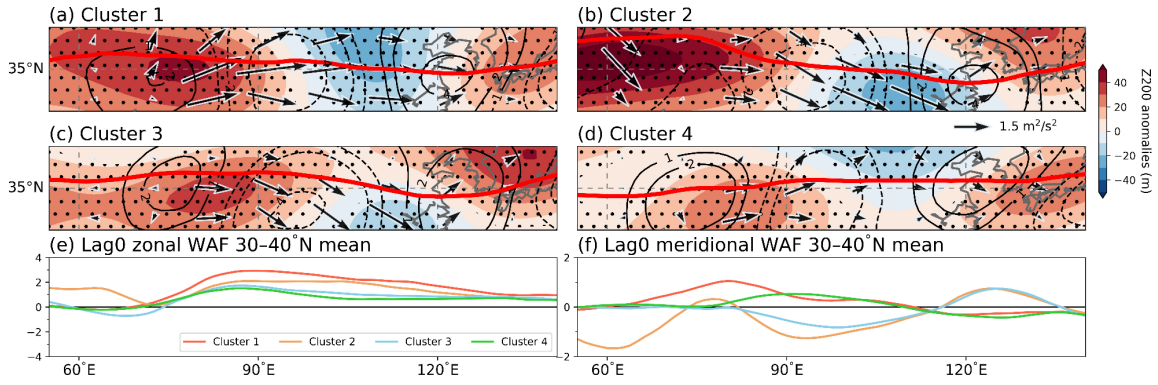


Figure 2-20 Rossby wave propagation in each cluster. Composite based on HIST of jet axis at event0 day (thick red curve), 200-hPa geopotential height anomalies at lag of 0 days (shading), and 200-hPa meridional wind anomalies (contours; unit: m/s) together with TN-WAF at lag of 0 days (arrows). (a) cluster 1, (b) cluster 2, (c) cluster 3, and (d) cluster 4. (e) and (f) depict zonal and meridional component of the TN-WAF (unit: m^2/s^2) averaged over $30^\circ\text{--}40^\circ\text{N}$ at lag of 0 days, respectively, for each cluster. Stippled regions indicate anomalies that are statistically significant at the 99% confidence level based on the two-sided Student's *t* test.

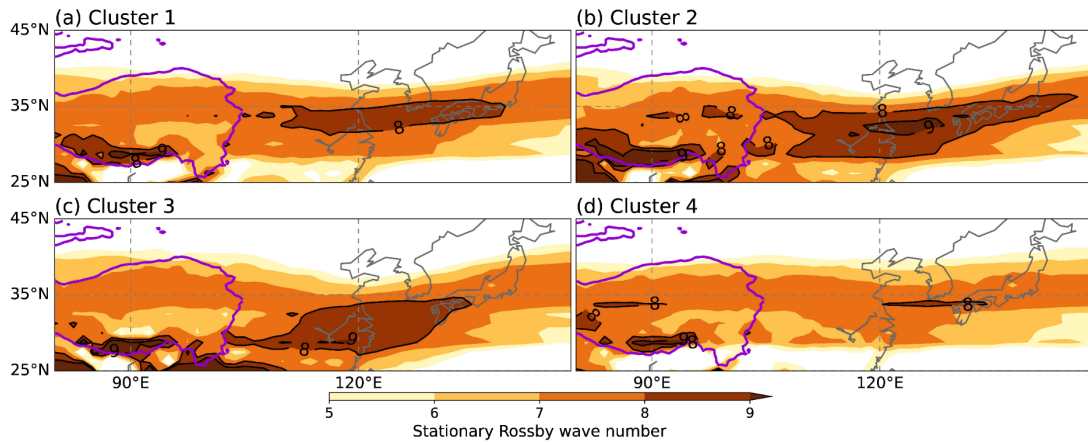


Figure 2-21 Stationary Rossby wavenumber for each cluster. Stationary Rossby wavenumber for (a) cluster 1, (b) cluster 2, (c) cluster 3, and (d) cluster 4. The wavenumber is determined based on Figure 2.4.4–3.

East Asia is covered by anomalous upper-level southerly wind and associated 500-hPa anomalous warm advection moving toward East Asia during lag days, regardless of the differences in the location and strength of the cold trough between clusters at lag0 day (Figure 2-19, i–l and q–t). These upper-level southerly winds cause warm advection over East Asia (Figure 2-19, m–p). Namely, strong anomalous 500-hPa warm advection near the wet band region is observed for all clusters, whose main body is overlapped by the anomalous southerly wind-driven warm advection during lag days (Figure 2-22), are related to ISM diabatic heating induced CGT wave train. Specifically, for clusters 2 and 3, anomalous warm advection at 500-hPa appear over

East Asia a few days earlier than the occurrence of the wet band. Meanwhile, for clusters 1 and 4, an intensification of the wet band is confirmed after the excitation of the ISMH peak, accompanied by a strengthened anomalous 500-hPa warm advection, although the anomalies were already present at lead of 10 days. The pre-existing warm advection in clusters 1 and 4 is presumably due to the prevailing southerly wind. We hypothesize that the jet meridional displacement in East Asia sector is a sign of the pre-existing southerly wind and wet band. This speculation is supported by the subseasonal lead-lag correlations, which suggest the occurrence of southerly wind prior than wet band precipitation, while jet meridional displacement tends to occur after both southerly wind and wet band precipitation (**Figure 2-23**).

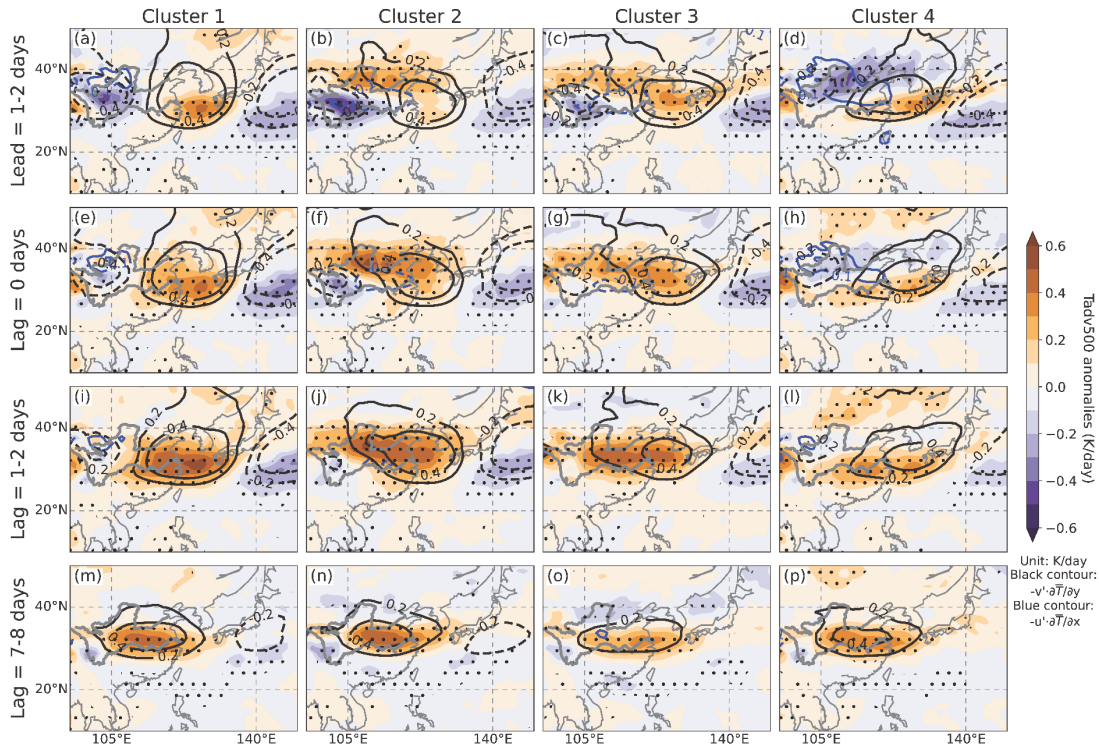


Figure 2-22 The spatial pattern of decomposition temperature advection drive by meridional wind and horizontal wind. Evolution of 500-hPa temperature advection anomalies on (a–d) lead of 1–2 days, (e–h) lag of 0 days, (i–l) lag of 1–2 days, and (m–p) lag of 7–8 days based on HIST. (a, e, i, m) cluster 1, (b, f, j, n) cluster 2, (c, g, k, o) cluster 3, and (d, h, l, p) cluster 4. The color shading represents the total 500-hPa temperature advection anomalies, and the black and blue contours indicate the anomalous 500-hPa temperature advections computed using anomalous meridional and zonal winds with climatological temperature gradient, respectively. Stippled regions indicate anomalies that are statistically significant at the 99% confidence level based on the two-sided Student's *t* test.

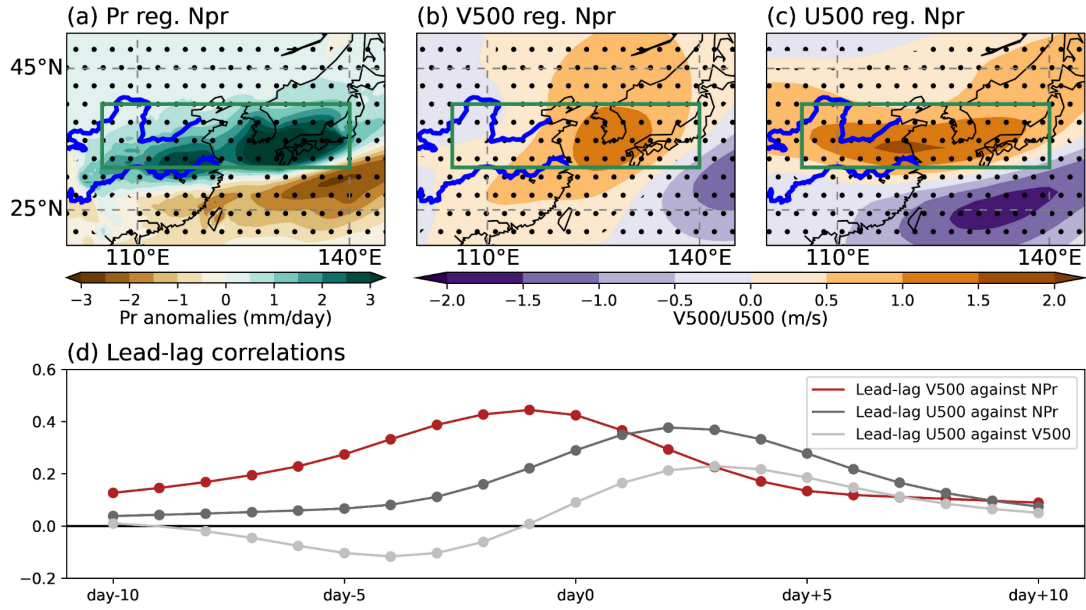


Figure 2-23 Relationship between EASM wet band precipitation, 500-hPa meridional wind and 500-hPa horizontal wind. The spatial pattern of (a) precipitation, and 500-hPa (b) meridional wind (V500), and (c) zonal wind (U500) regressed upon standardized daily wet band precipitation (Npr) in June created using large ensemble historical climate experiment (HIST). The green box denotes the wet band area. Stippled regions indicate anomalies that are statistically significant at the 99% confidence level based on the two-sided Student's t test. (d) lead-lag correlations among Npr, V500, and U500 averaged over the wet band area in June. The V500 lead-lag Npr means correlation of lead-lagged 500-hPa meridional wind against Npr, and similar manner for others. Lines with solid circle denotes correlations satisfy two-sided $p < 0.01$.

These results reveal that CGT wave train steers anomalous warm advection and ascending air over East Asia in lag days (Figure 2-19, i-l contours, and q-t green hatching). The warm advection and associated upward motion is known to lead to greater precipitation, which can anchor the EASM rain band (Sampe & Xie, 2010). Therefore, the ISM-induced CGT wave train and the associated anomalous warm advection by the anomalous southerly wind over the East Asian sector are important for the EASM rain band. Meanwhile, the location of the cold center guided by the westerlies jet is also important since it has an effect to strengthen anomalous cold advection to East Asia and determines the sign of the temperature advection anomalies (Figure 2-19, m-p). The low-level water vapor, in addition, fuels the wet band in response to the upper-level anomalous southerly wind. The maximum 500-hPa meridional wind anomalies are located to the north of 30°N, while those at 850-hPa are generally to the south of 30°N (Figure 2-19, q-t, u-x). This indicates that the southerly wind over East Asia tilts northward with height in the troposphere. The interaction between 500-hPa warm advection and 850-hPa water vapor flux driven by the southerly wind is essential for the growth of the wet band. In summary, the results suggest that

the CGT wave train tends to steer mid-tropospheric warm advection and lower-tropospheric water vapor toward East Asia that contribute to the formation of the wet band. Moreover, the results infer that ISMH acts as a trigger (for clusters 2 and 3) of the wet band, but also as a reinforcer (for clusters 1 and 4) if abnormal southerly winds and associated warm advection are pre-existing. Meanwhile, the location and intensity of the cold center in part of the CGT guided by the jet is also shown to be important for the wet band prediction.

The WAF for each cluster is presented in **Figure 2-24** to explore the pathway of wave propagation. In the previous section (i.e., [section 2.4.3](#)), the ISM is considered diabatic heating as essential forcing to excite the Rossby wave that interact with the CGT. Earlier studies suggested that the CGT is triggered by forcings other than ISMH, such as diabatic heating near the Mediterranean (Yasui & Watanabe, 2010), land surface processes over the Middle East (Jingqiu Yang et al., 2021), jet stream instability, and heating related to tropical convection (Sun et al., 2022).

In cluster 1, the CGT is pre-existing before ISM onset, which is consistent with the pre-existent wet band. Centers with positive stream function anomalies can be identified over the North Atlantic Ocean, western Mediterranean, northern Indian subcontinent, and East Asia in lead days. The role of the ISM in this case is to maintain the CGT and associated wet band over East Asia. In clusters 2 and 3, the WAF originates from the Ural Mountains and moves toward the Indian subcontinent in lead days. A strong meridional Rossby wave is a possible cause of the meandering of the jet stream (**Figure 2-17**) and the occurrence of ISMH activities (Q. Ding & Wang, 2007, 2009), indicating potential interaction between the polar jet and the subtropical jet. In this case, ISMH is the major trigger of the abnormal warm advection over East Asia and the associated wet band. In cluster 4, wave activity is seen at high latitudes prior to ISM diabatic heating (i.e., in lead days), whose pathway originates from western Europe and runs across the Ural Mountains and East Asia, causing an anomalous anticyclone over East Asia. On this occasion, the role of ISMH is to reinforce warm advection over East Asia, similar to cluster 1, and to help maintain the wet band over East Asia. The results demonstrate that the dipole precipitation pattern over East Asia is influenced by various remote factors with different pathways. ISM heating promotes development of the wet band over East Asia in collaboration with these remote factors.

In subseasonal lead-lag composite, we notice discrepancy in the precipitation response to ISMH between the observations and HIST (**Figure 2-9** and **Figure 2-10**). The pre-existing wet band precipitation anomalies in HIST is dissimilar to the result of the observations (**Figure 2-11**). It is interpreted that the greater number of samples in HIST could detect cases in which ISMH reinforces the pre-existing rainband.

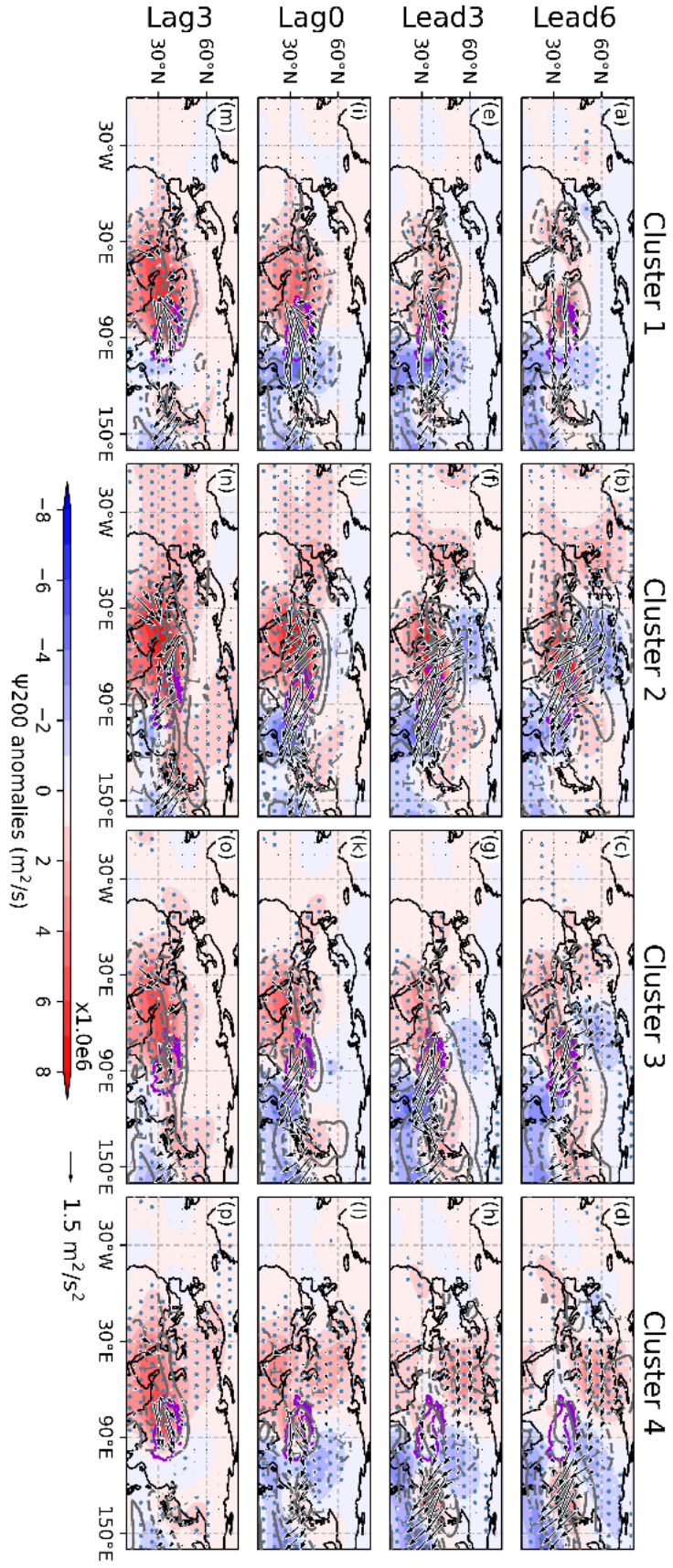


Figure 2-24 Lead-lag composite of wave activities relative to abnormal ISM heating in each cluster. Stream function (color shading), 200-hPa zonal wind anomalies (contours, unit: m/s), and wave activity flux (arrows) for (a–d) lead of 6 days, (e–h) lead of 3 days, (i–l) lag of 0 days, and (m–p) lag of 3 days. (a, e, i, m) cluster 1, (b, f, j, n) cluster 2, (c, g, k, o) cluster 3, and (d, h, l, p) cluster 4. The stippled regions denote anomalies significant at the 99% confidence level based on the two-sided Student's t test. Wave activity flux with values lower than 0.01 are omitted.

2.4.5. Quantification of East Asian rainfall by Indian heating

In this study, we focus on the ISM–EASM connection on the subseasonal time scale, building upon the observations and large-ensemble experiment. Here the quantitative insights is provided regarding the contribution of ISM to the EASM wet band (defined as the precipitation anomalies area-averaged over 31° – 40° N, 105° – 140° E; **Figure 2-10**) on the subseasonal timescale. **Figure 2-25** compares probability distribution of daily wet band precipitation anomalies created using the entire study period (i.e., all days from June and first 2 pentad of July) and using only lag of 0–10 days as the period that contains the ISM influence. In this figure, positive precipitation anomalies greater than 0.5 mm/day is regarded as wet band events. It is clear that the wet band precipitation anomalies during lag of 0–10 days tend to shift to positive values, with mean wet band precipitation anomalies of 1.04 mm/day for observations and 0.66 mm/day for HIST. Among all wet band events, the probability of wet band events from June to early July with presence of lag of 0–10 days is 21.02% for observations and 16.72% for HIST. Noteworthy, these probabilities rise to 23.26% and 19.77%, and further to 31.58% and 24.67% for the events exceeding +1 and +2 standard deviations of wet band precipitation anomalies in the observation and HIST, respectively.

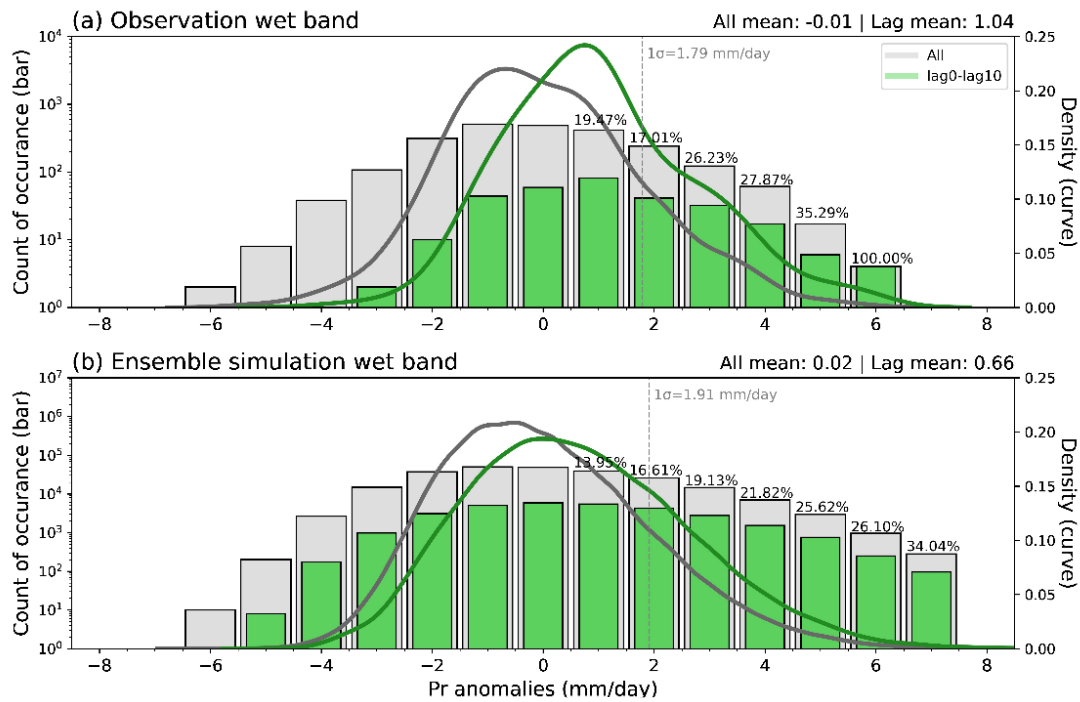


Figure 2-25 Potential contribution of ISM to EASM precipitation at subseasonal timescale. The probability distribution of wet band precipitation anomalies based on (a) observation and (b) HIST. Grey and green colors are for all days of the entire study period (i.e., days from June and first 2 pentads in July) and for days during lag of 0–10 days, respectively. The bars denote the count of occurrences (left Y-axis) for every 1 mm/day bin, and percentage value at the top of each bar refers to wet band occurrences coinciding with lag of 0–10 days. The occurrence of the wet band is defined as positive precipitation anomaly greater than 0.5 mm/day. The domain for the wet band is depicted as the green rectangle in Supplementary Fig. 1. The value of +1 standard deviation of June to early July wet band precipitation anomalies is illustrated by dashed vertical line. The curves represent the probability density (right Y-axis), with the mean values at the top right of each panel.

The probability of EASM wet band events with presence of lag days (i.e., influenced by the abnormal ISM heating) increases with increasing precipitation anomalies (**Figure 2-25**). These findings suggest that the ISM plays a prominent role in stronger wet band precipitation anomalies. In particular, the exceedance probabilities of +1 standard deviation of wet band precipitation anomalies during lag of 0–10 days are 1.83 times higher in observations and 1.59 times higher in HIST, respectively, in comparison to the probability estimated for daily wet band precipitation anomalies in June and early July. Therefore, this indicates the likelihood of wet band event is higher when abnormal ISM heating exists.

We are hereby also interested in if the subseasonal processes could be beneficial for the prediction of June mean precipitation, because the leading mode of June monthly mean precipitation in East Asia also exhibits a similar dipole pattern in both GPCP and HIST (**Figure 2-26a**). We select cases when these criteria on different time scales are either partially satisfied or both satisfied based on the detection of lag0, meaning the active ISMH, and the principal component for interannual June precipitation (hereafter, EAPR–PC1) which shows dipole precipitation pattern. The lag0 days was targeted for the case that occurred only within the first 5 pentads in June, considering the lead–lag relationship between the ISM and the EASM (**Figure 2-9**, **Figure 2-10**, and **Figure 2-11**). Comparison of monthly precipitation in June with and without detection of lag0 days (i.e., strong and persistent ISMH) allows us to establish the impact of subseasonal ISMH on monthly East Asian precipitation.

The EAPR–PC1 high years (defined as $\text{EAPR-PC1} > 0.5$) with detection of lag0 days show strong connection between Indian precipitation and the East Asian dipole precipitation pattern (**Figure 2-26b** and **f**) in June. Meanwhile, the EAPR–PC1 high years without strong ISMH show the East Asian dipole pattern of precipitation, but Indian precipitation is close to neutral (**Figure 2-26c** and **g**). Given the complex influencing factors of the EASM, the dipole precipitation pattern can also be induced by factors other than the ISM, including the Pacific–Japan-like pattern evident in the observational data shown in **Figure 2-26c**. Mean precipitation in June for the year with lag0 days shows a relatively weak dipole pattern in the result of the observations and HIST (**Figure 2-26d** and **h**), whose projected EAPR–PC1 scores are 0.65 and 0.37, respectively. These scores represent the extent to which subseasonal ISM affects interannual EASM. Meanwhile, for years where lag0 was successfully detected, the EAPR–PC1 score shifts accordingly to the positive values. We also confirm that higher ISMH values (indicating stronger ISM forcing) correspond to higher interannual EAPR–PC1 scores, as expected (**Figure 2-26i** and **f**). These findings underscore the role of subseasonal ISMH on the formation of the East Asian rainband. The signal of contribution of subseasonal ISM heating in the interannual EASM is magnified or obscured if multiple processes coexist. The unstable relationship of monthly or seasonal precipitation between the ISM and EASM, as mentioned in introduction, is speculated to indicate that the monthly average precipitation does not sufficiently represent the physical ISM–EASM connection signal.

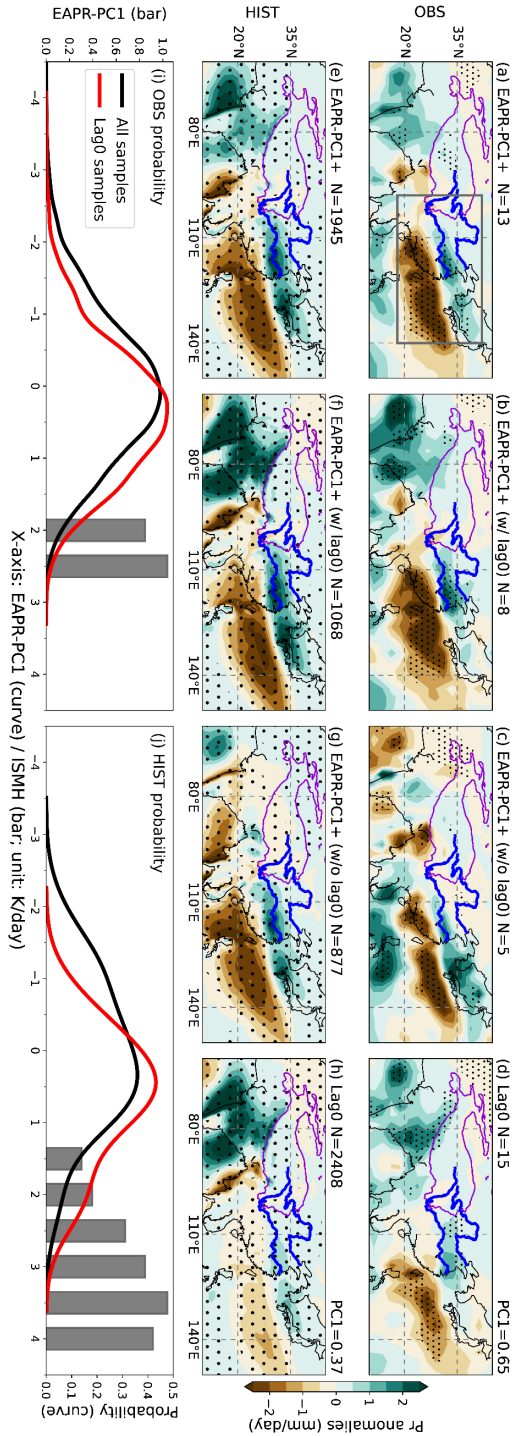


Figure 2-26 Potential contribution of ISM to EASM precipitation intrannual timescale. Composite June monthly mean precipitation pattern based on (a) years with first leading mode of June monthly mean East Asian precipitation (EAPR-PC1) greater than 0.5 (EAPR-PC1+), (b) years with both EAPR-PC1 greater than 0.5 and with detection of lag0, (c) years with EAPR-PC1 greater than 0.5 but without lag0, and (d) years with lag0 only. (i) probability (right y-axis) of the EAPR-PC1 (x-axis) distribution for all samples (black curve) and the samples with lag0 (red curve), where the grey bars represent the EAPR-PC1 score (left y-axis) estimated from the conditioned lag0 Indian monsoon heating (ISMH) magnitude for each 1 K/day interval (x-axis). (e-h) and (j) are similar as (a-d) and (i), respectively, but for large ensemble historical climate experiment (HIST). Dotted regions denote anomalies significant at the 90% and 99% confidence level for the observations and HIST, respectively. Number of samples used for the composite analysis are indicated as N. The projected EAPR-PC1 values are shown on the top right of (d) and (h). The domain to obtain EAPR-PC1 is shown in the grey box in (a). The ISMH is averaged over 10°–30°N, 70°–85°E and applied vertical mass-weighted from 850 to 200 hPa.

2.4.6. The influence of the sea surface temperature

In [section 1.2.4](#), the SST' forcing such as ENSO can lead to the ISM–EASM non-stationary is illustrated, due to the decoupling between ISM and ENSO. In this section, the impact of SST on the lead–lag composite results (i.e., results presented in [section 2.4.2](#)) are discussed. Since HIST is driven by the monthly mean SST/sea ice condition, i.e., the lower boundary conditions are near identical in the same calendar month of a year among ensemble members. Therefore, the number of ensemble members which successfully detected lag0 day is counted in each year (**Figure 2-27**), to identify the possible influence of SST (lower boundary conditions in HIST) on the occurrence probability of the lag0.

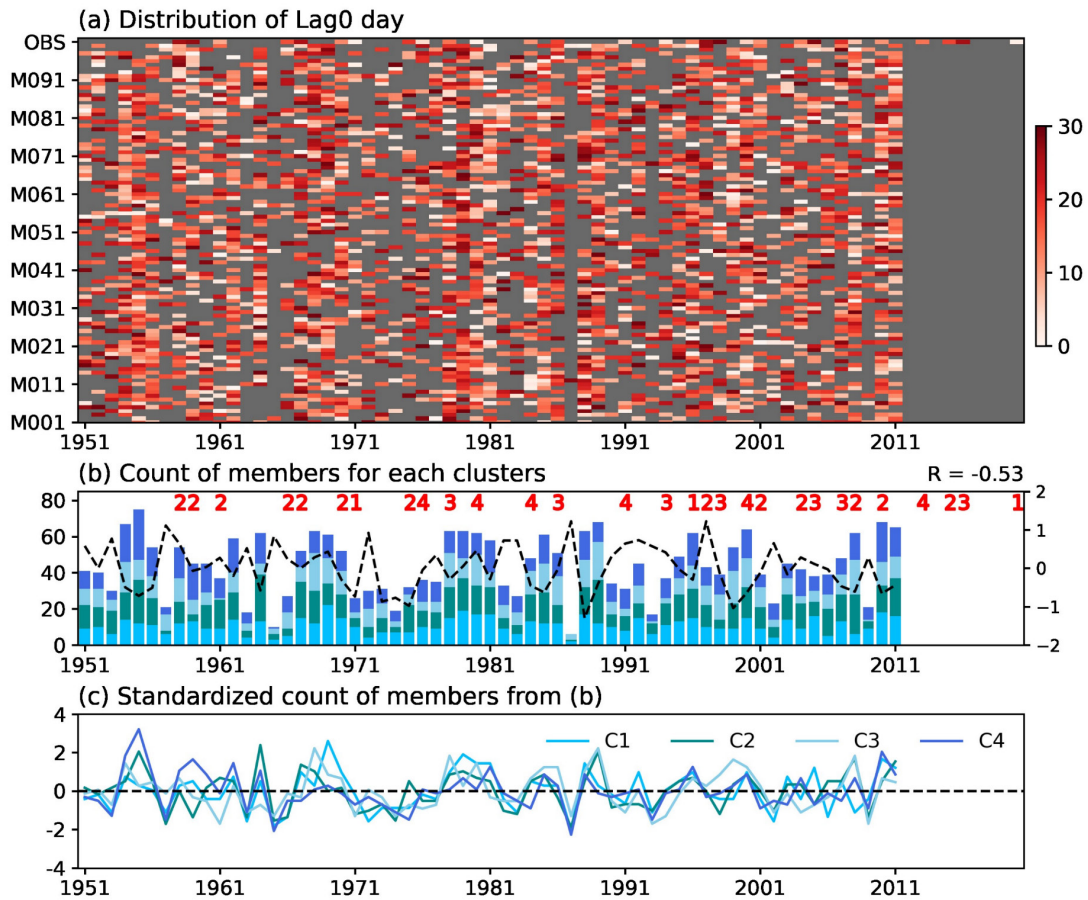


Figure 2-27 Interannual occurrence probability of lag0 day in relation to SST. (a) The heatmap showing the occurrence of lag0 day for observation and each HIST ensemble member (y-axis) during 1951 to 2020 (x-axis), with the color shading stands for the June calendar date of the lag0 day, and the grey shading for without detection of lag0 day or no data, and (b) the count of ensemble members with successful detection of lag0 day for each year (bars) and each cluster (colors; see legend of (c)), with June Niño 3.4 index (black dash) and labelled cluster number of observation (number in red), (c) the standardized count of ensemble members with successful detection of lag0 day for each cluster (color lines, see legend), which is based on (b).

The interannual count number of the HIST ensemble member with successful detection of the lag0 day is highly correlated with the June Nino 3.4 index negatively, with a correlation coefficient of -0.53 during 1951–2011 ($p < 0.01$), which reflect the constrain of ENSO forcing in the occurrence of lag0 day (i.e., ISM activity; **Figure 2-27b**). These results are in corroboration with the understanding in regular ENSO–ISM coupling, reflecting higher likelihood of strong ISM (i.e., strong and persistent ISM heating; lag0) in the La Niña summer, and vice versa for weak ISM in El Niño summer. As for the spatial pattern (**Figure 2-28**), a distinct La Niña pattern is found for each cluster, obtained through regressing June SST against the standardized count of ensemble members with successful detection of lag0 day (**Figure 2-27c**).

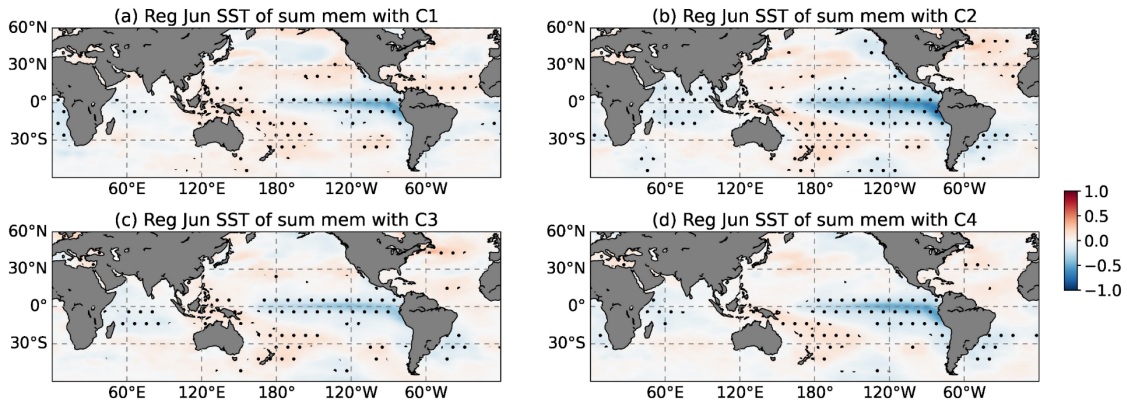


Figure 2-28 HIST June SST pattern with higher probability of lag0 day for each cluster. The regressed June SST (shading; unit: K) upon the standardized count of ensemble members with successful detection of lag0 day, based on Fig. 2.4.6–1c, for (a) cluster 1, (b) cluster 2, (c) cluster 3, and (d) cluster 4. Stippled regions indicate anomalies that are statistically significant at the 95% confidence level based on the two-sided Student’s t test.

Though it is well known that the ENSO forcing has a great impact on the ISM, the strong ISM (i.e., with successful detection of lag0 day) could still be found in El Niño summer (e.g., years 1963, 1965, and 1987). Indeed, every year within the HIST marks successful detection of lag 0 days with different SST/sea ice patterns spanning from 1951 to 2011. This feature suggests that internal atmospheric forcing plays a significant role in driving the ISM, potentially leading to its sustained activity even during strong El Niño summers. Among the 29 selected lag0 day samples obtained from observations, the composite June Nino 3.4 index registers at -0.09 ($p = 0.37$). Those observed samples comprise 5 occurrences of La Niña summers and 3 instances of El Niño summers, while the other samples exhibit Nino 3.4 index values within the range of ± 0.5 (ENSO-neutral). Thus, the significance of atmospheric internal variability in influencing intense ISM heating events is supported by both observational data and HIST ensemble simulation. This is also supported by the fact that the observed EAPR–PC1 is not correlated with the ensemble mean HIST EAPR–PC1, as illustrated in **Figure 2-4**. Based on these features, the observed outcomes (29 samples) are speculated to be the nearly random replicate from the HIST (2757 samples).

To confirm whether the ISM–EASM connection differs depending on the ENSO phase, the lead–lag composite is conducted again for HIST, but for each El Niño or La

Niña summer. Based on Nino 3.4 index with a threshold of ± 0.5 , the 2757 samples from HIST are further derived to 631 samples for El Niño (**Figure 2-29**), 880 samples for La Niña (**Figure 2-30**), and rest 1246 samples for ENSO-neutral summer. Again, the internal processes might be also important for the strong ISM heating activity, as we conjectured. Meanwhile, the ISM–EASM connection seems not influenced by ENSO phase, since the CGT pattern, the CGT wave propagation (as indicated by the TN-WAF), and the wet band precipitation responses in the lead–lag days are nearly identical (**Figure 2-29** and **Figure 2-30**). The only difference between them is the dry band and geopotential height near the tropical region, which is largely influenced by ENSO. This result suggests that the ISM–EASM connection that we presented is independent from ENSO.

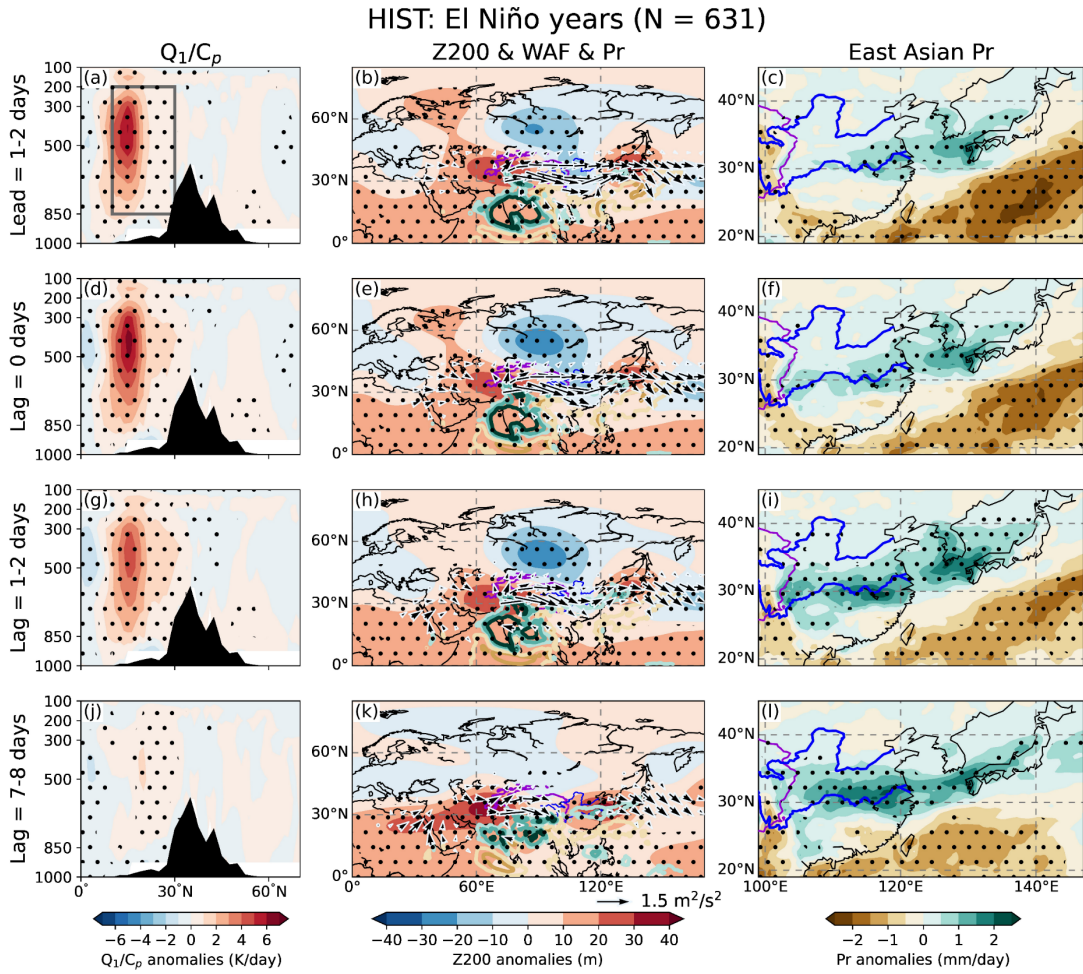


Figure 2-29 Simulated lead–lag characteristics of circulation and precipitation relative to abnormal ISM heating during El Niño summer. As in **Figure 2-9**, but for El Niño summer.

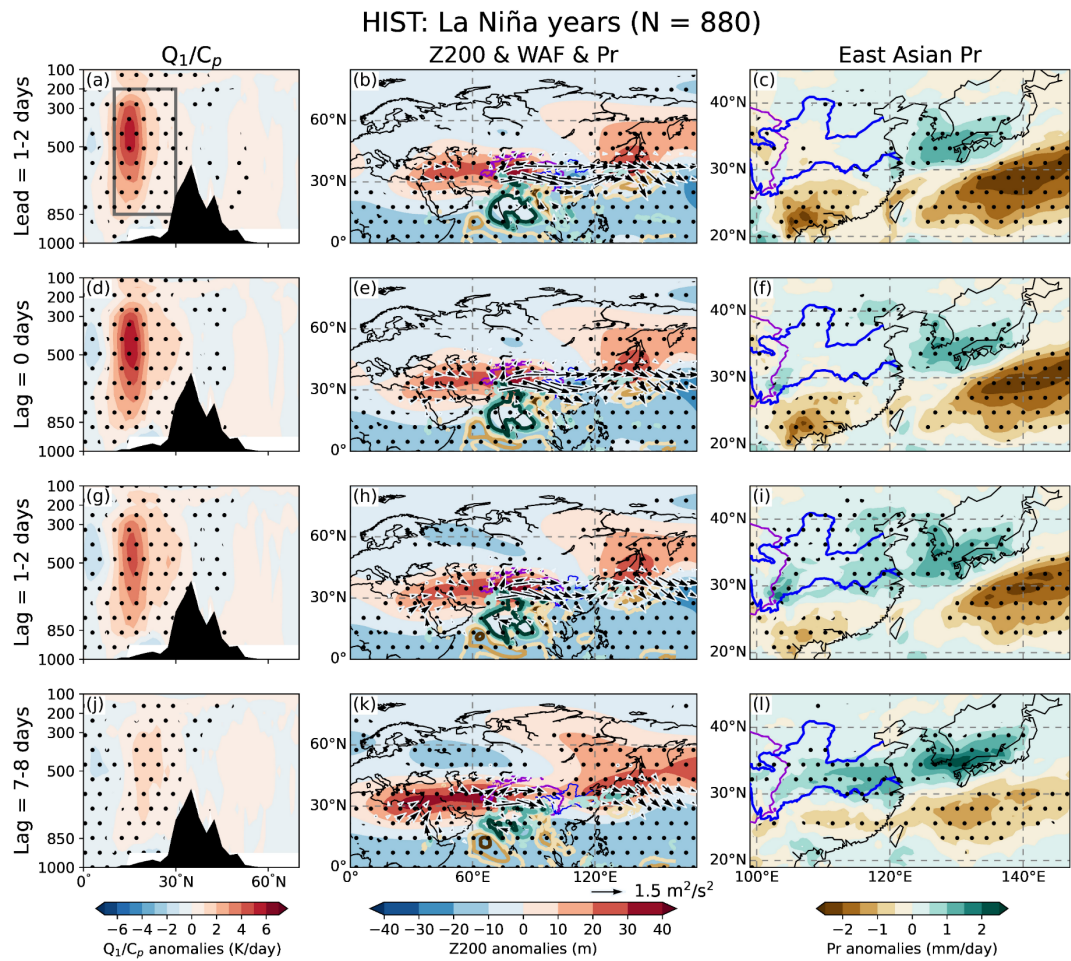


Figure 2-30 Simulated lead-lag characteristics of circulation and precipitation relative to abnormal ISM heating during La Niña summer. As in Figure 2-9, but for La Niña summer.

2.4.7. The influence of the tropical intraseasonal oscillation

The tropical intraseasonal oscillation (ISO) is the most predominant subseasonal variability in the tropics, characterized by a spectral peak of 30–60 days or 30–90 days (Kikuchi, 2021). The widely recognized tropical ISO is the Madden–Julian Oscillation (MJO), which is more vigorous and dominant in winter season propagates eastward in the tropics (Madden & Julian, 1971). As for the tropical ISO in summer season, which is referred to as the Boreal Summer Intraseasonal Oscillation (BSISO) with an additional northward propagation interacting with the Asian monsoon system, in comparison to the MJO (Chang et al., 2021). The northward propagation of the BSISO results in its significant influences on the active/break cycles of monsoon systems in South and Southeast Asia, contingent upon its specific propagation phase (e.g., Q. Ding & Wang, 2009; Lee et al., 2013; Yasunari, 1979). In this section, the potential role of ISO in the identified subseasonal connection between the ISM and EASM is carefully clarified, based on the observation data.

The observed BSISO and MJO indices during 1979–2020 are first examined (see [section 2.3](#) for details), from lag of 0 day to lag of 10 days for each abnormal ISM heating events (**Figure 2-31**). The MJO phase space diagram shows the MJO is absent during the lag days of the ISM–EASM interaction, as its principle component scores are generally lower than 1 for nearly all detected abnormal ISM heating events (**Figure 2-31b**). On the contrary, the BSISO propagation exists in the lag days of most abnormal ISM heating events with their principle component scores greater than one. The propagation of BSISO in the phase space diagram shows that most of the selected anomalous ISM heating events at lag0 are close to phase 3-5 and move counterclockwise to phase 6 or 7 at a lag of 10 days, with the average propagation from phase 4/5 to phase 6 from lag0 day to lag 10 days. The composite map of 30–60-day bandpass filtered outgoing longwave radiation (OLR) for each BSISO phase is presented in **Figure 2-32**, which indicates the Indian subcontinent are covered by significant negative OLR anomalies during the BSISO phase 4/5. This proves that the lag0 (i.e., the abnormal ISM heating) is influenced by the cycle of BSISO.

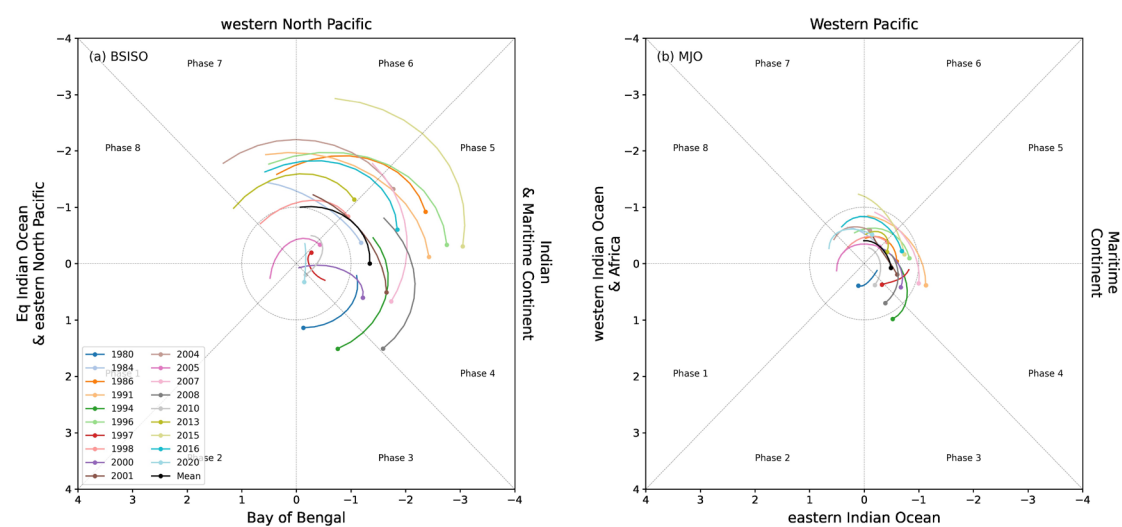


Figure 2-31 Phase space of tropical ISOs for each abnormal ISM heating events. Phase space based on (a) BSISO and (b) MJO indices from lag0 day to lag of 10 days for each selected subseasonal abnormal ISM heating events. The dot for each trajectory stands for lag0 day (starting point).

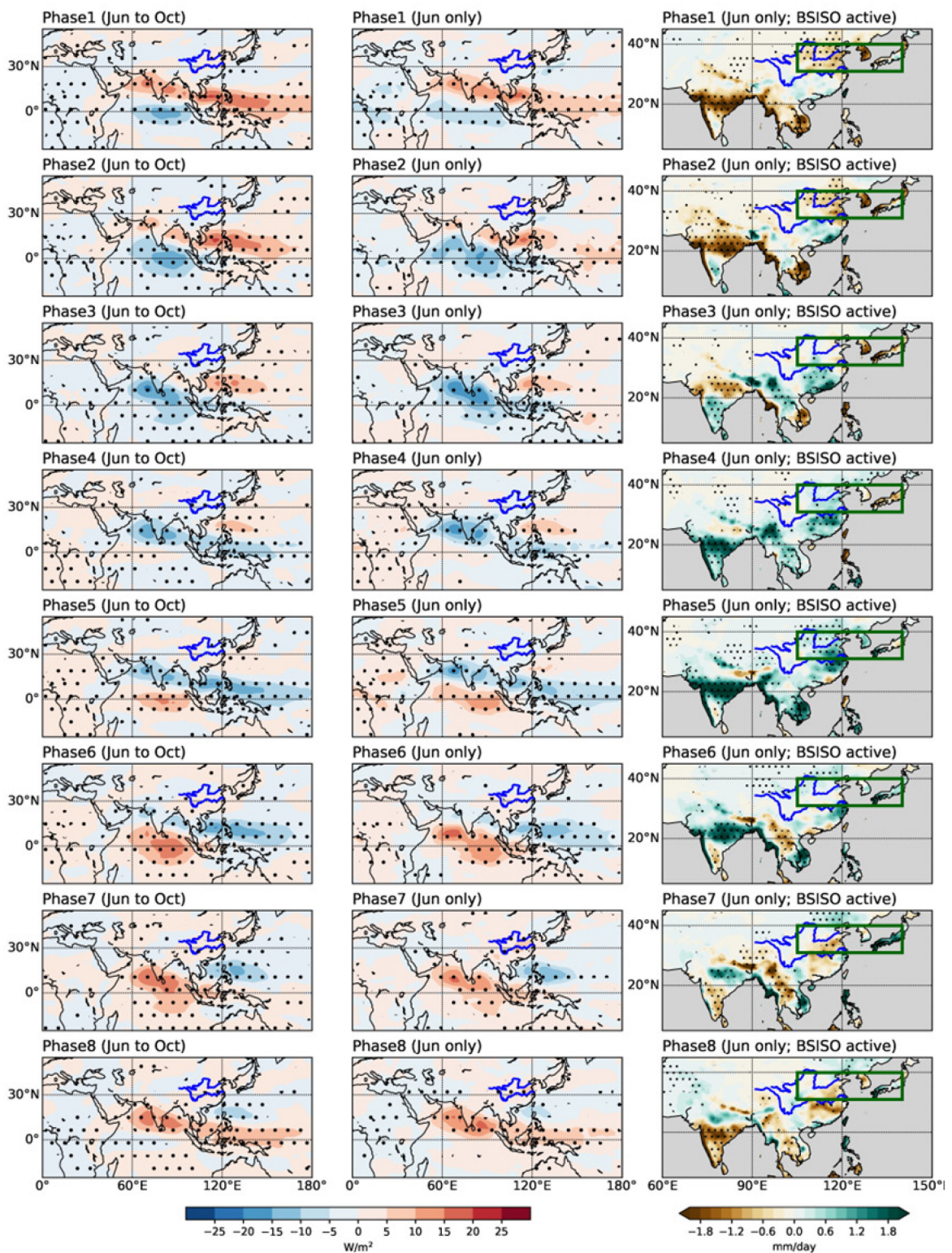


Figure 2-32 Composite BSISO life cycle. The composite map of 30–60-day bandpass filtered OLR anomalies for each BSISO phase that from boreal summer (left column) and June only (middle column), and 30–60-day bandpass filtered precipitation anomalies when BSISO is active (right column). Stippled regions indicate anomalies that are statistically significant at the 99% confidence level based on the two-sided Student's t test.

Then, the lead–lag composite analysis is applied to a 30–60-day bandpass filtered precipitation data. This approach offers insights into the precipitation signal associated with the BSISO. The lead–lag composite result again denotes that the Indian precipitation is related to BSISO, as positive precipitation anomalies are found over the Indian subcontinent (**Figure 2-33**). This aligns with **Figure 2-32**. In the previous study, it is proposed that the interaction between CGT from mid-latitudes and ISO from the equator served as triggers for abnormal convection near the ISM domain coincidentally (Q. Ding & Wang, 2007). On the other hand, East Asian precipitation is less influenced by the BSISO, as the East Asian dipole precipitation anomalies pattern vanishes when applying the 30–60-day bandpass filter (i.e., in comparison with leftmost and middle columns of the **Figure 2-33**, which without the 30–60-day bandpass filter). Specifically, the East Asian wet band precipitation occurring from lag of 0 days to lag of 8 days does not exceed 0.40 mm/day, whereas it reaches up to 1.93 mm/day when without applying the 30–60-day bandpass filter (**Figure 2-33**).

The result of using 10–20-day bandpass filter is also presented here, since this is another spectral peak of the ISO periodicity of ISM (Annamalai & Slingo, 2001). With this 10–20-day bandpass filter applied, the ISM–EASM connection is also not observed (**Figure 2-34**).

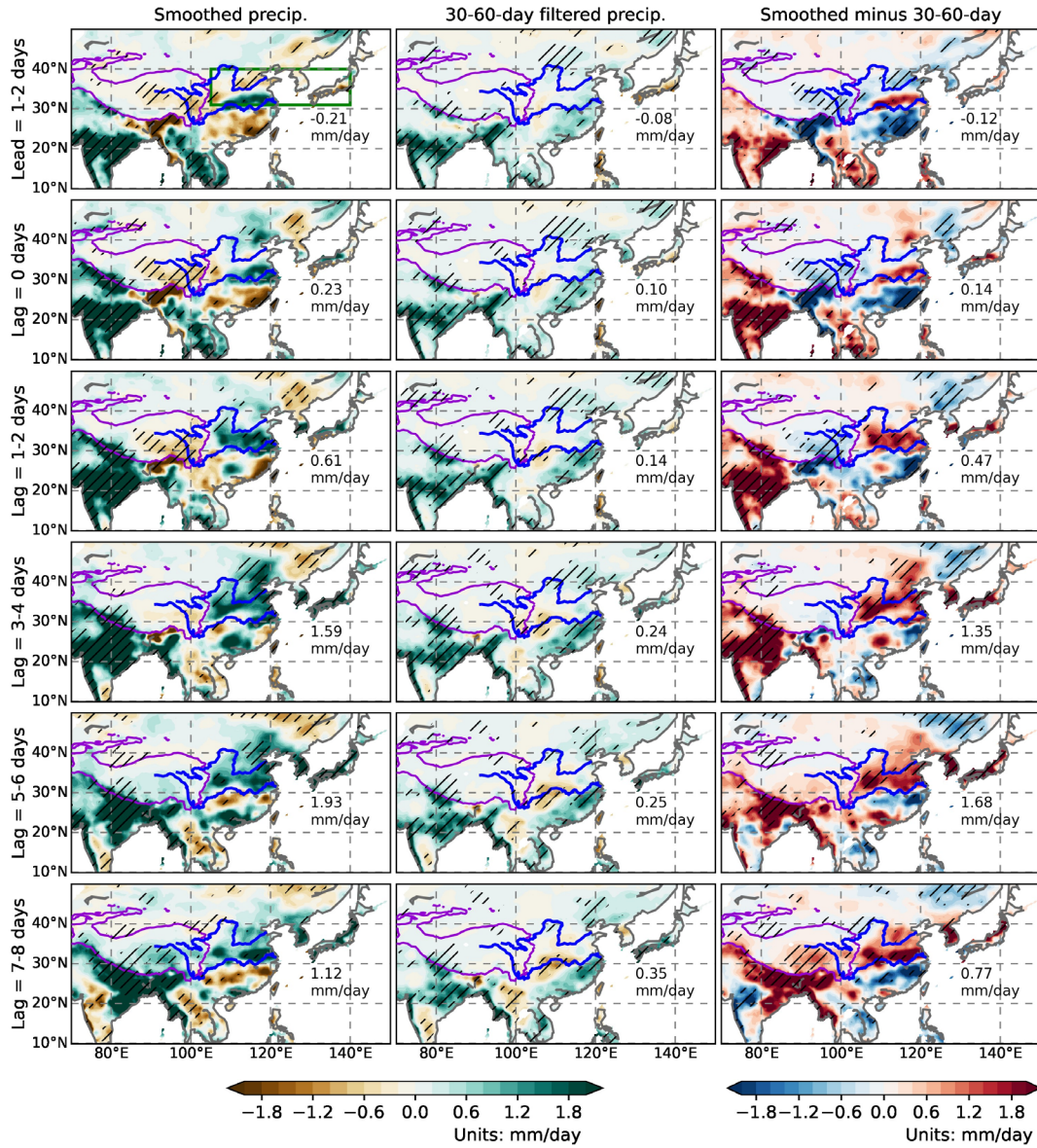


Figure 2-33 The ISM–EASM connection with 30–60-day bandpass filter. The lead–lag composite precipitation (unit: mm/day) of (left column) using the smoothed precipitation data as described in [section 2.3](#), and (middle column) using the filtered precipitation data with 30–60-day bandpass filter, and (right column) the difference between smoothed precipitation and filtered precipitation. The value on the lower right corner of each panel shows the area averaged precipitation anomalies over the wet band area (green rectangle). Stippled regions indicate anomalies that are statistically significant at the 90% confidence level based on the two-sided Student’s t test.

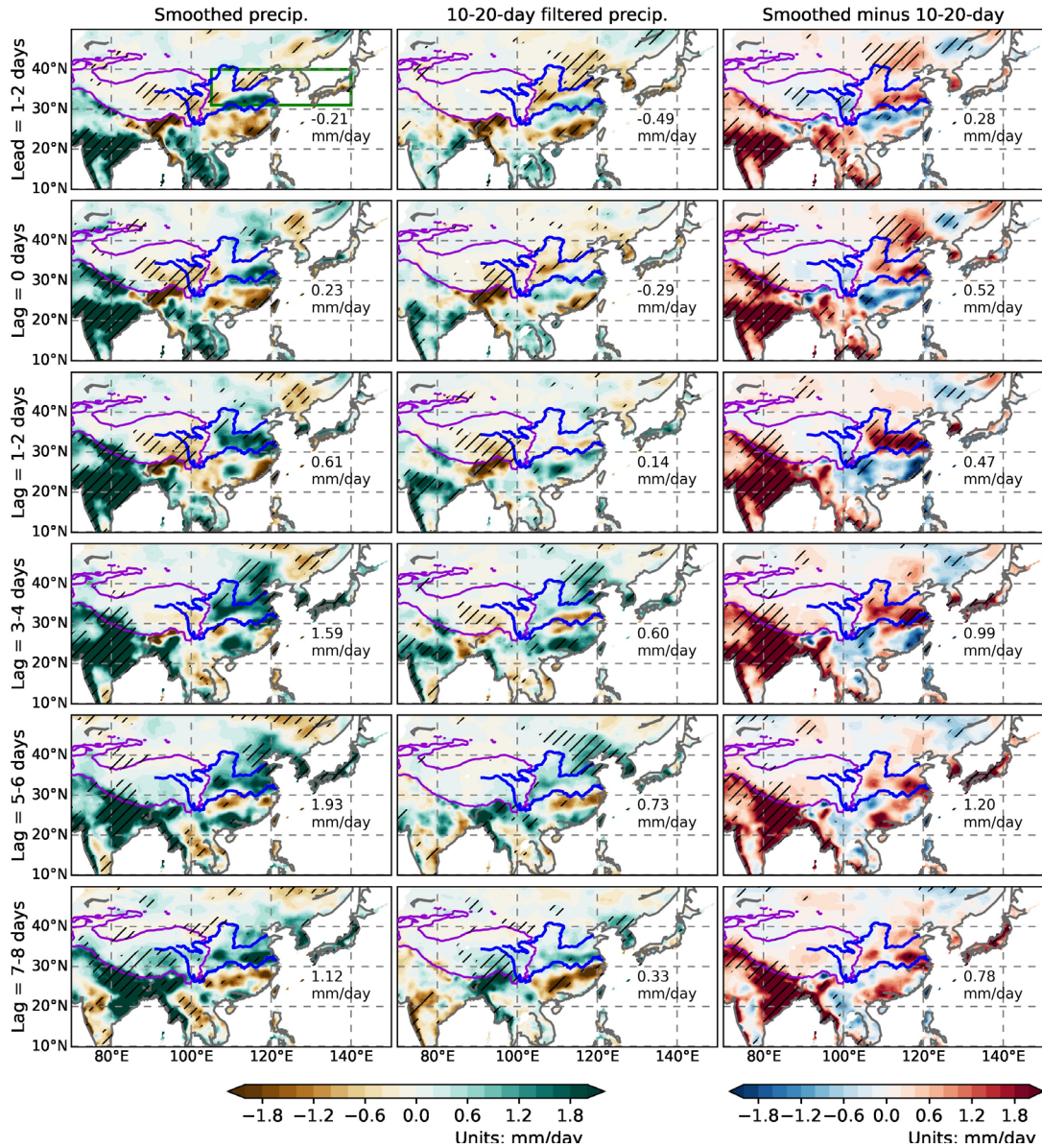


Figure 2-34 The ISM–EASM connection with 10–20-day bandpass filter. Similar to **Figure 2-33**, but with 10–20-day bandpass filter applied. Stippled regions indicate anomalies that are statistically significant at the 90% confidence level based on the two-sided Student’s t test.

To quantitatively evaluate the contribution of tropical ISOs to the EASM, the 30–60-day bandpass filtered wet band precipitation anomalies are calculated for each ISO phase in June for both BSISO and MJO (**Table 2-1**). The result shows that the influence of BSISO on the EASM wet band is overall weak, with maximum wet anomalies appearing in phase 5 of 0.46 mm/day, and the maximum dry anomalies in phase 2 of -0.63 mm/day. It has greater contributions to both dry band and ISM rainfall anomalies in contrast to that of wet band, which might be because their locations are closer to the tropical region. Although the MJO is absent during lag0 day to lag of 10 days, it has a potential greater influence on EASM than BSISO. In MJO phase 1, the wet band precipitation anomalies reach to -0.68 mm/day, while that in phase 6 is 0.45 mm/day. As for dry band precipitation anomalies, in MJO phase 3–5, the values are 0.82, 0.88,

and 1.24 mm/day, respectively. The MJO exhibits a comparable impact on ISM compared to the BSISO.

Table 2-1 Composite June 30–60-day bandpass filtered June precipitation anomalies (mm/day) of EASM wet band (Npr; 31°–40°N, 105°–140°E), EASM dry band (Spr; 20°–29°N, 105°–140°E), and ISM (10°–30°N, 70°–85°E) for each phase of BSISO and MJO when they are activate, based on the APHRODITE from 1958–2016.

		Phase 1	Phase 2	Phase 3	Phase 4	Phase 5	Phase 6	Phase 7	Phase 8
BSISO	Npr	-0.55***	-0.63***	-0.32***	0.06	0.46***	0.26***	0.21*	-0.08
	Spr	0.22	0.38**	0.68***	0.50***	0.18	0.03	-0.11	-0.06
	ISM	-1.27***	-1.05***	-0.10	0.92***	1.38***	0.81***	0.01	-0.79***
MJO	Npr	-0.68***	/	-0.01	0.28	-0.15	0.45***	0.28**	-0.33***
	Spr	-0.47*	/	0.82***	0.88***	1.24	-0.14	0.13	-0.62**
	ISM	-0.99***	/	0.02	0.93***	1.56	0.34	-0.55***	-1.06***

It should be noted that, these values are overall smaller than the values obtained from the lead–lag composite analysis, which is up to 1.93 mm/day during the lag of 5 days to lag of 6 days. Therefore, the influence of BSISO and MJO from the tropics might be weak on EASM during the lag days. However, the BSISO plays a vital role as one of the important triggers for the abnormal ISM heating (i.e., lag0). In conclusion, it is found that these results suggest that the influence of the tropical ISO on the connection between ISM and EASM is comparatively weaker than that of the mid-latitude CGT.

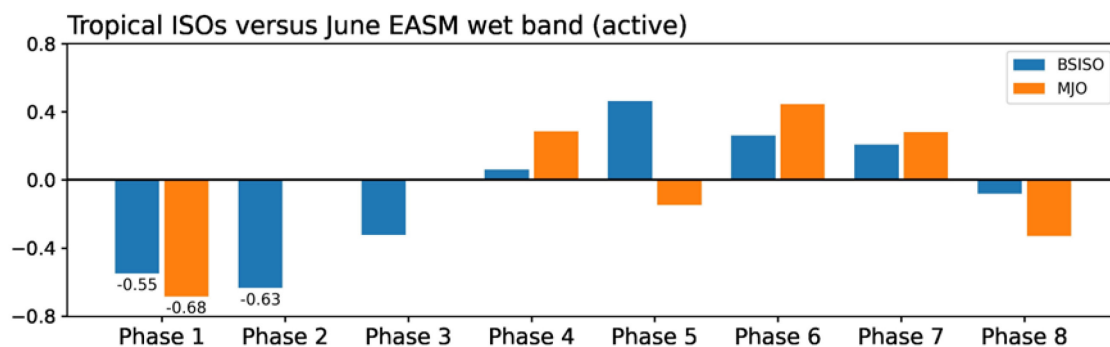


Figure 2-35 Relationship between EASM wet band and tropical ISOs. The composite 30–60-day bandpass filtered June precipitation anomalies (mm/day) of EASM wet band for each phase of BSISO and MJO when they are active. The absolute values greater than 0.5 mm/day are labelled.

2.5. Discussions

(i) The relationship between ISM and travelling wave

In [section 2.4.3](#), a transition of the 200-hPa wave pattern between lead of 10 days and lag of 10 days is found over East Asia, showing a bow-like longitude-time variation on the Hovmöller diagram (**Figure 2-14**). This variation corresponds to a binomial time variation (i.e., peak-valley-peak) of CGTI-PC1 in the same period. In this section, the possible interaction between ISM heating and this transition of wave pattern is firstly discussed. In [section 2.3](#) (**Figure 2-2**), the CGT reference pattern (i.e., CGTI-PC1, which is derived from interannual timescale) is set to track the subseasonal CGT variation, obtained via projection of the lead-lag composite 200-hPa meridional wind anomalies onto the CGT reference pattern. However, it is speculated that the low CGTI-PC1 near lag0 day (**Figure 2-12**; valley of the binomial time variation) is due to interaction or the travelling of the different wave trains in the westerlies jet stream. To answer this question, the EOF analysis is applied to the June daily 200-hPa meridional wind to explore the subseasonal wave activities, over the domain of 20° – 60° N, 0° – 150° E (same domain for CGT reference pattern). Hereafter, the EOF patterns and associated principal components from subseasonal 200-hPa meridional wind are referred to as SSV200-EOF and SSV200-PC, respectively. The EOF output show that in both JRA-55 and HIST (**Figure 2-36** and **Figure 2-37**), the two leading modes are two wave trains with orthogonal wave phases to each other, with the explained variance nearly identical. The explained variance of SSV200-EOF1 versus SSV200-EOF2 are 15.11% versus 12.15% in JRA-55, and 12.53% versus 11.64% in HIST, respectively. This result indicates that these two-wave train are the travelling waves dominant at a subseasonal time scale. The SSV200-EOF2 mode indeed resemble to the CGT reference pattern, in particular for the wave pattern over south Asia to the East Asia sector, with the anomalous southerly wind dominant over the East China. The SSV200-EOF1, on the other hand, corresponding to a wave train that with wave phase orthogonal to the CGT (i.e., SSV200-EOF2), with the anomalous southerly wind dominant over both the East China and Korean Peninsula, i.e., which is slightly shifted to the downstream of the jet stream in comparison of CGT.

The lead-lag composite 200-hPa meridional wind anomalies relative to lag0 are then projected onto these two leading modes for JRA-55 and HIST, respectively. The projected SSV200-PC2 variation from lead of 20 days to lag of 20 days shows a peak-valley-peak time variation, which exactly matched to the CGI-PC1 variation during the same period (**Figure 2-36e** and **Figure 2-37e**). The projected SSV200-PC1, conversely, is closer to a single peak time variation, maximized close to lead of 3 days. This result helps to explain the wave pattern transition between lead of 10 days and lag of 10 days relative to abnormal ISM heating, which shows a bow-like longitude-time variation. It is in fact composed of two wave trains with wave phases orthogonal to each other and travelling together.

To illustrate the relationships of these two wave trains with the abnormal ISM heating, the output of LBM simulation is investigated. The LBM is first regirded from original T42 to global grids with a horizontal resolution of 1.25° . Then the LBM simulated 200-hPa meridional wind anomalies are projected on to SSV200-EOF1 and SSV200-EOF2, respectively, based on JRA-55 (**Figure 2-38a**). The projected SSV200-PC1 from the LBM experiment shows a rapid response to the ISM heating, growing rapidly in first few days and stabilizing after 6 days in the simulation. Whereas the projected SSV200-PC2 of the LBM shows a somewhat delayed response, with

values being kept low for the first few days and then developing rapidly without reaching to equilibrium in the first 20 days in the LBM.

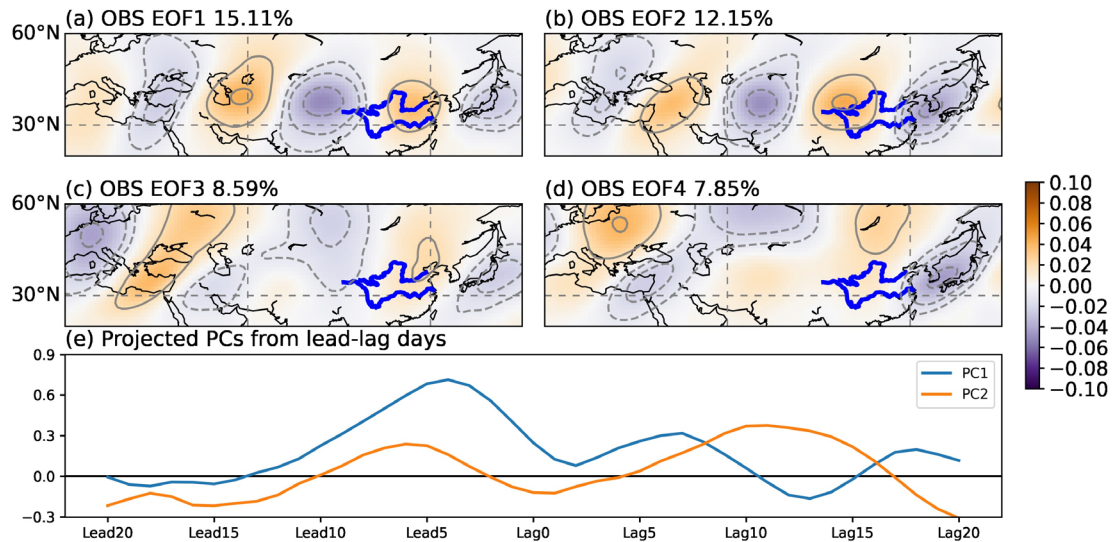


Figure 2-36 The JRA-55 leading patterns of June subseasonal wave train. The leading patterns of June subseasonal 200-hPa meridional wind (SSV200) based on JRA-55 for (a–d) first four leading SSV200 modes, and (e) the projected 200-hPa meridional wind anomalies from lead of 20 days to lag of 20 days for SSV200-PC1 and SSV200-PC2

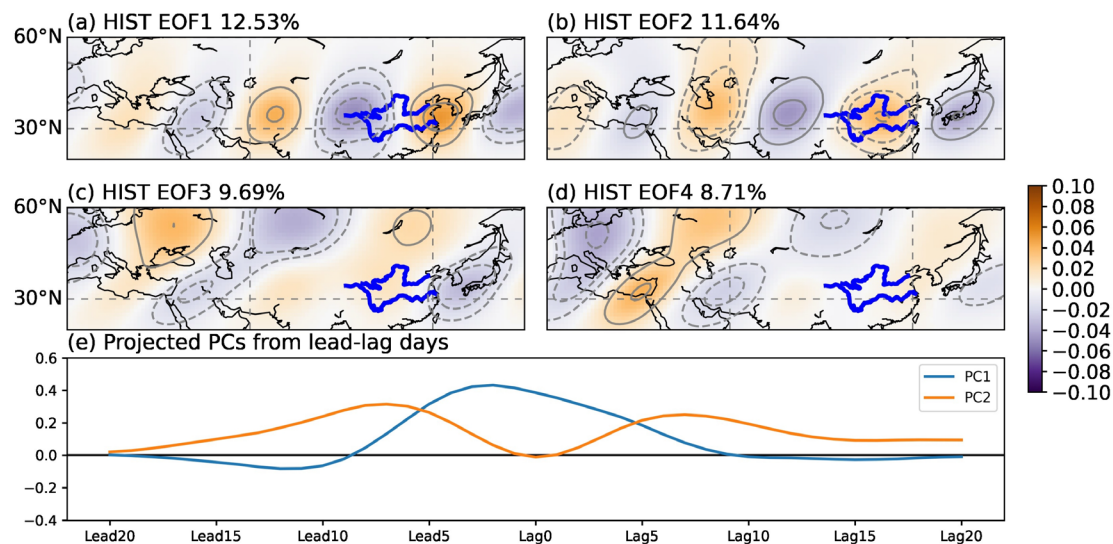


Figure 2-37 The HIST leading patterns of June subseasonal wave train. Similar to Figure 2-36, but for HIST.

To investigate the mechanism of the delayed response of SSV200–PC2 to ISM heating in the LBM, another LBM experiment is performed. Unlike the LBM experiment described and analyzed in [section 2.3](#), which assumed constant ISM heating forcing, the ISM heating forcing is removed on day 2 in the LBM simulation (hereafter, referred to as LBM–day2). Based on LBM–day2 experiment, the projected SSV200–PC1 decays quickly after day 2, while the projected SSV200–PC2 reaches its peak in the simulation near the day 9 and decays thereafter (**Figure 2-38b**). This provides the evidence that SSV200–PC1 and SSV200–PC2, the two dominant subseasonal Rossby train with orthogonal wave phase in the westerlies, are directly and indirectly forced by the ISM heating, respectively. This question may deserve more attention in the future, as the observed ISM heating variation is neither constant nor has it suddenly disappeared. More numerical experiments beyond the linear model are necessary.

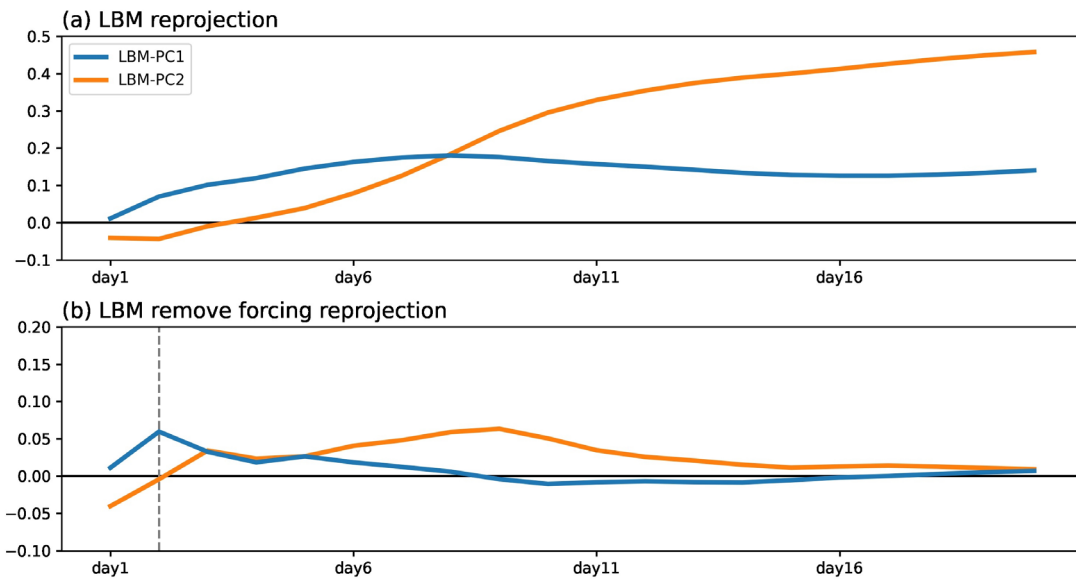


Figure 2-38 The time variation of SSV200 pattern in the LBM. The projected SSV200-PC1 and SSV200-PC2 for (a) LBM with constant ISM heating forcing and (b) the LBM experiment with ISM heating forcing removed after day 2 (LBM–day2). The vertical grey dash stands for day 2.

(ii) The ISM–EASM connection in other summer month

The CGT is the leading circulation mode prevailing from June to September over the Northern Hemisphere (see [section 1.2.1](#) and **Figure 2-2** for detail). The [section 2.4](#) only investigated the CGT variability in June. In this discussion, how to apply what we learn from June's situation to variations over other summer months is shown (i.e., June to September). To accomplish this, the ISMH thresholds are determined for each summer month based on their corresponding standard deviation (**Table 2-2**).

Table 2-2 ISMH threshold (K/day) and sample size for each summer month for JRA-55 and HIST.

Month	1 std. dev. values (JRA-55 / HIST)	Threshold (JRA-55 / HIST)	Events (JRA-55 / HIST)
Jun	1.06 / 1.10	1.00 / 1.00	29 / 2757
Jul	0.98 / 0.79	1.00 / 0.80	25 / 1991
Aug	0.94 / 0.72	1.00 / 0.70	22 / 2200
Sep	0.99 / 0.77	1.00 / 0.80	26 / 2187

The lead-lag composite analyses were once again performed but using adjusted thresholds for each summer month, based on their 1 standard deviation corresponding values computed from ISMH. The results suggest that, in contrast to June, the ISM diabatic heating during July to September is weakly or even not associated with the CGT circulation pattern at subseasonal time scale (**Figure 2-39** and **Figure 2-40**). A meridional dipole-like precipitation pattern was found in the July of observation, but not in the HIST. Meanwhile, the response of precipitation in August and September is generally much weaker than that in June, i.e., without showing a dipole-like precipitation pattern in the East Asia domain.

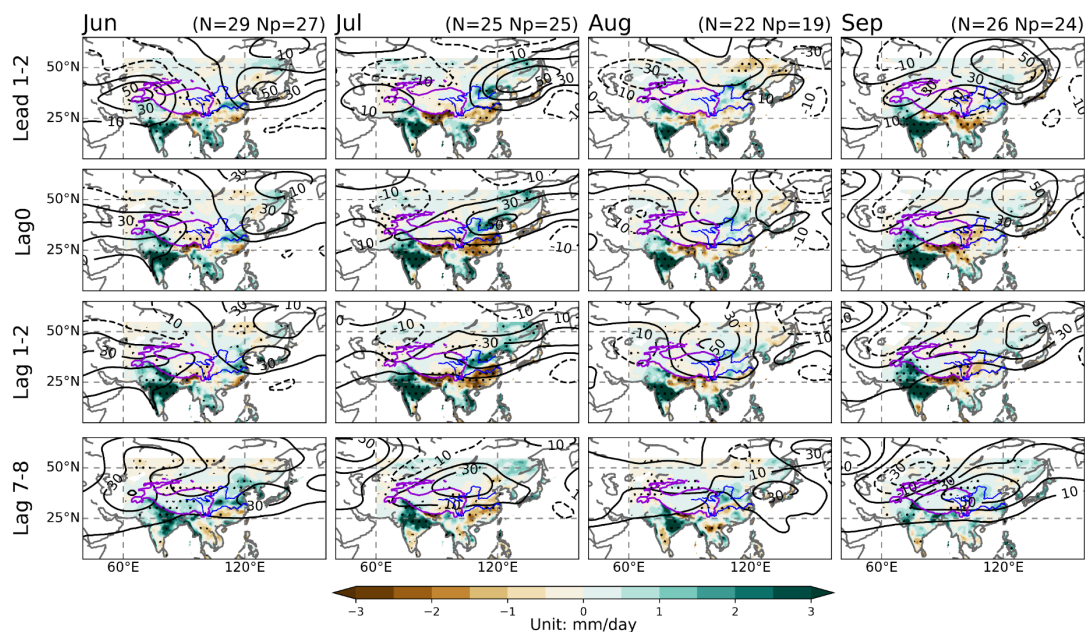


Figure 2-39 The observed ISM–EASM connection in each summer month. Evolution of the observed composite of 200-hPa geopotential height (contours; units: m) and precipitation (color shading; units: mm/day) from June to September. N and Np indicates samples used for 200-hPa geopotential height and precipitation, respectively. Stippled regions indicate anomalies that are statistically significant at the 90% confidence level based on the two-sided Student’s t test.

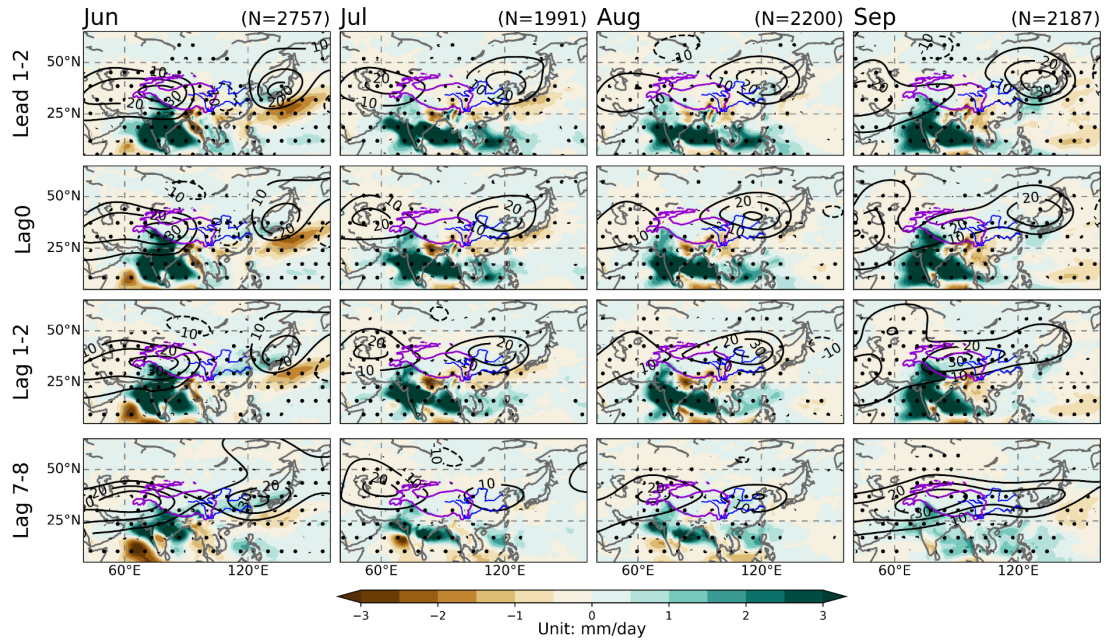


Figure 2-40 The simulated ISM–EASM connection in each summer month. Similar to **Figure 2-39**, but for HIST. Stippled regions indicate anomalies that are statistically significant at the 99% confidence level based on the two-sided Student’s t test.

To pursue whether the ISM can trigger East Asian CGT in other summer months, the LBM experiments were carried out to see the atmospheric response to the ISM diabatic heating. Five scenarios are considered (i.e., each month from May to September) with the 1979–2020 JRA-55 monthly mean climatological basic state. For each scenario, the same ISM diabatic heating of the June lag0 days averaged anomalies is imposed over the 10° – 30° N, 70° – 85° E region constantly (see **Figure 2-3** and [section 2.3](#) for more information). The LBM output shows that in May and June, the ISM diabatic heating triggers a regional CGT wave train pattern that propagates toward East Asia along the westerlies, causing positive meridional wind prevailing in the East Asia (**Figure 2-41**). In contrast, the regional CGT wave train is not observed in the LBM simulation for July and August. In September, although the Rossby waves can propagate to East Asia, the TN-WAF indicates that they are positioned farther north than those in May and June.

In conclusion, the ISM–EASM connection observed in June is not applicable to other summer months, especially July and August, when the CGT pattern is less coupled with the ISM heating.

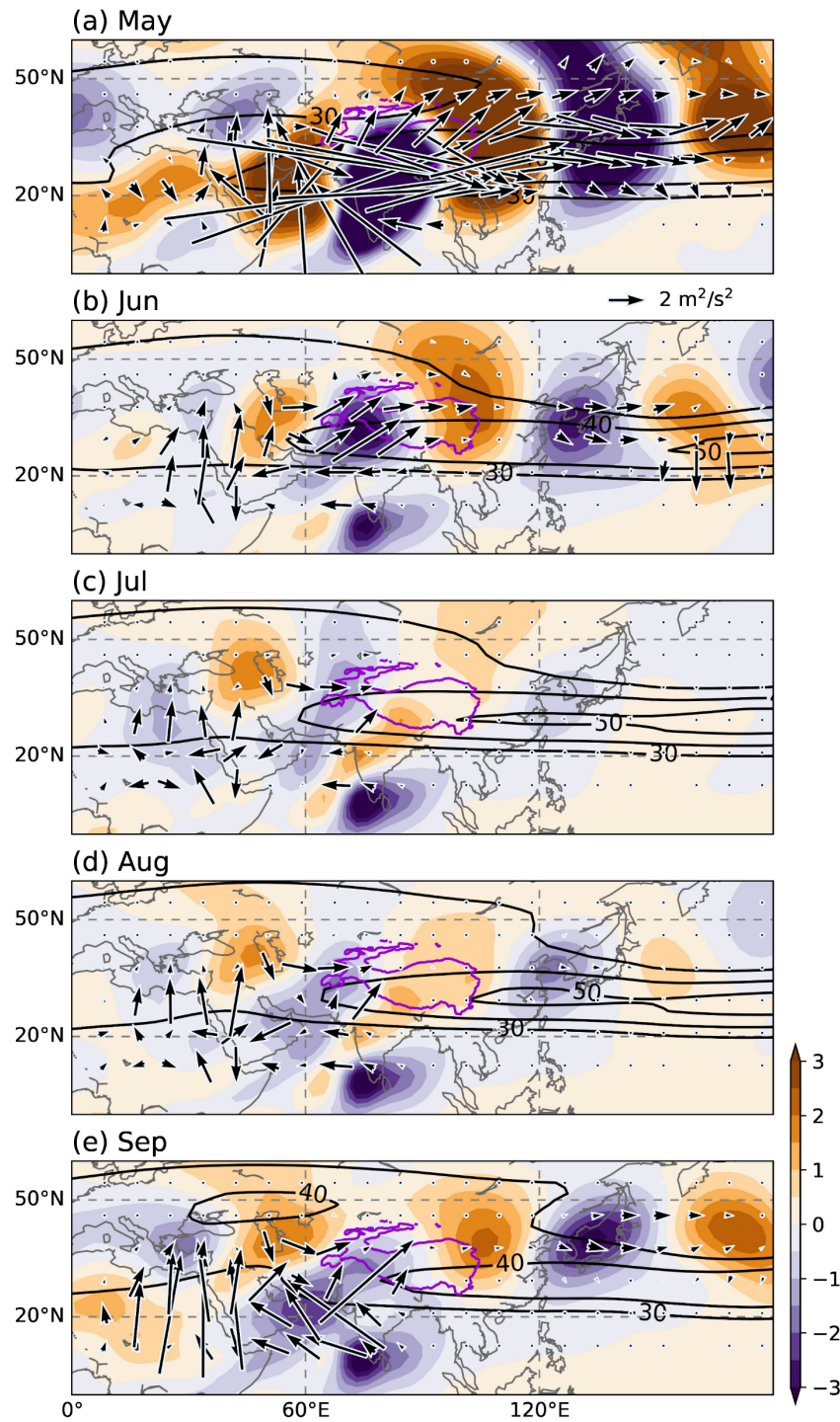


Figure 2-41 The LBM simulated wave train in response to constant ISM heating. The responses of the 200-hPa meridional wind (shading, units: m/s) and the TN-WAF (arrows) to the ISM diabatic heating simulated by LBM. Results are averaged from 30 to 50 days when LBM is equilibrium. (a) with May basic state was used, (b–e) similar to (a) but for June to September basic states were used, respectively.

(iii) The relationship between wet and dry bands in East Asia

In [section 2.4.2](#), it is conjectured that dry band in East Asia is not influenced by the Indian diabatic heating and associated CGT wave pattern. In this discussion, the relationship among wet band, dry band, and ISMH are presented in **Figure 2-42**, by computing lead-lag correlation of dry band precipitation index with respect to wet band precipitation index and ISMH. The results (both observation and HIST) indicate the dry band and wet band are likely to occur simultaneously (see also **Figure 2-22**). Therefore it is speculated that the development of wet band and dry band reflects a northward location/migration of the rain band of EASM.

On the other hand, the results also denote that the dry band tends to be more prominent prior than the ISM heating becomes vigorous. This is linked to the activity of low-level easterly winds before the ISM becomes active (**Figure 2-9, u-x**). This is out of scope in the current work, and needs the future investigation.

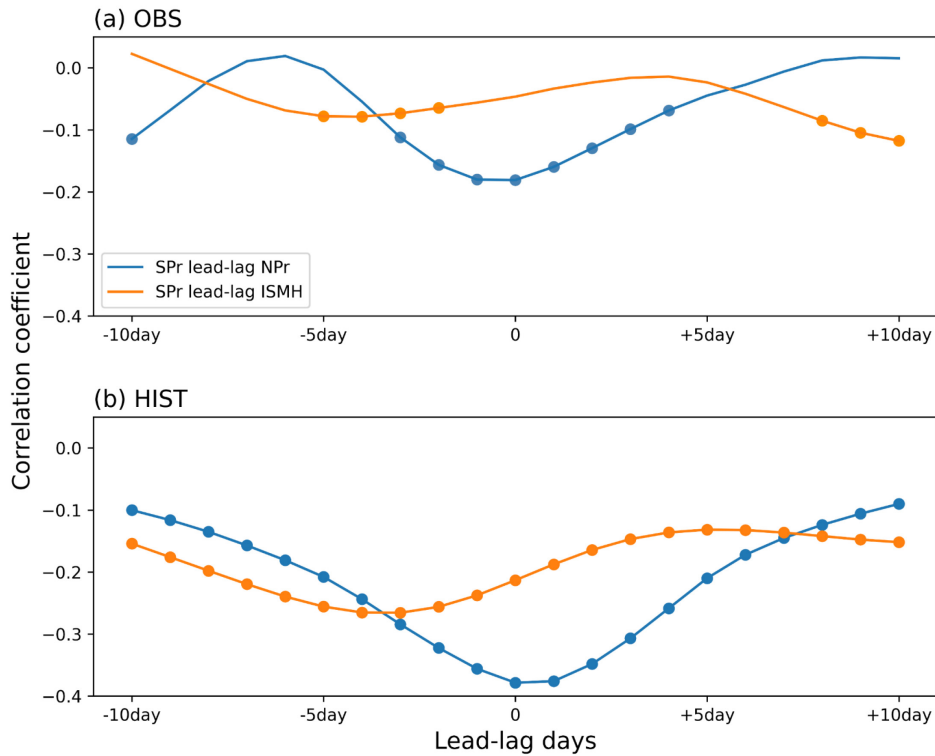


Figure 2-42 The relationship of wet and dry bands in East Asia. June subseasonal lead-lag correlations between wet band precipitation anomaly index and ISMH, respectively, relative to dry band precipitation anomaly index based on (a) observation data (APHRODITE) and (b) HIST data. SPr and NPr are for precipitation anomalies averaged over dry band and wet band regions, respectively. SPr lead-lag NPr means lead-lag correlations of SPr with respect to NPr, similarly for SPr lead-lag ISMH. Values with solid circle satisfy $p < 0.01$.

(iv) The relationship between CGT and Pacific-Japan pattern

The Pacific-Japan pattern (hereafter, PJ-pattern) is the dominant summer circulation pattern at lower level over the East Asia, associated to enhanced convection activities near the Philippines (Kosaka & Nakamura, 2010; Nitta, 1987). The PJ-pattern is identified as first leading mode of summer mean (June to August) 850-hPa relatively vorticity field over East Asia (0° – 60° N, 100° – 160° E; P. Xu et al., 2019). Following this criterion, the PJ-pattern is defined based on JRA-55 (**Figure 2-43**), characterized by lower-level negative vorticity anomalies over tropical western North Pacific near Philippines, positive vorticity anomalies cover mid-latitude East Asia, and negative vorticity anomalies near Far East Eurasia and Sea of Okhotsk. This is understood as a northward propagation of the Rossby wave excited by the convection activities near the Philippines (Nitta, 1987). In [section 2.4.5](#), a PJ-pattern-like pattern is shown to interact with the EASM dipole precipitation (**Figure 2-26c**). Therefore, in this subsection the relationship between EASM, CGT, and PJ-pattern is investigated.

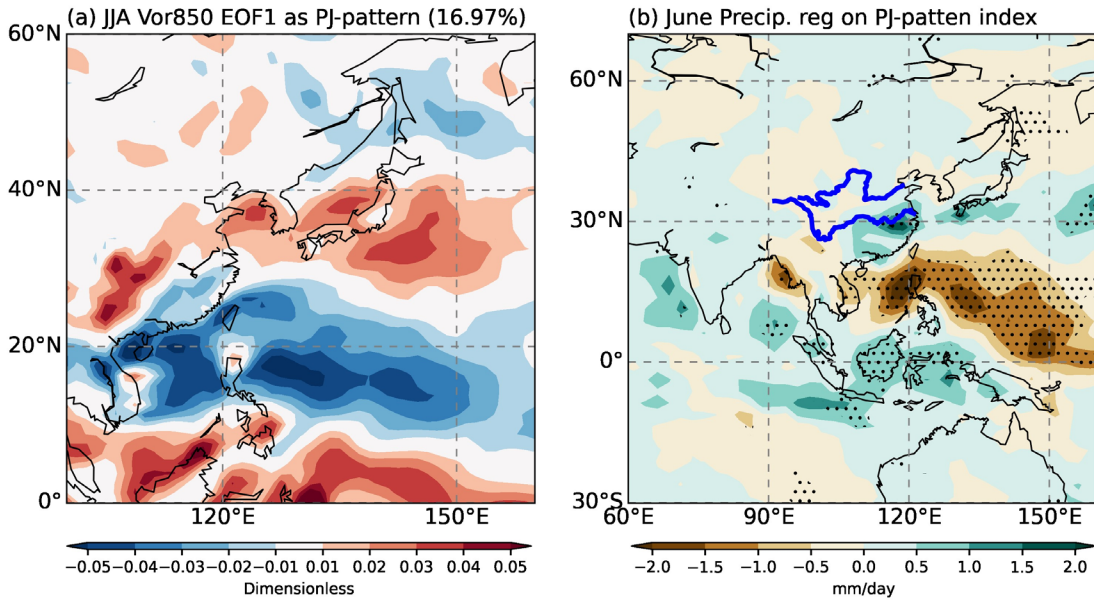


Figure 2-43 The PJ-pattern and associated precipitation anomalies. (a) The PJ-pattern derived from EOF of June to August 850-hPa vorticity anomaly, and (b) the PJ-pattern associated precipitation anomaly in June based on GPCP. Stippled regions indicate anomalies that are statistically significant at the 95% confidence level based on the two-sided Student's *t* test.

Subsequently, the June daily 850-hPa relatively vorticity anomalies from JRA-55 are projected onto PJ-pattern to obtain the daily PJ-index. The correlations between daily PJ-index, EASM wet band and dry band precipitation anomalies area-mean indices, ISM precipitation anomalies area-mean indices, and CGTI-PC1 are summarized in **Table 2-3**. The results indicate that PJ-pattern could influence the precipitation of EASM dry band and CGT, while it does not influence the EASM wet band and ISM precipitation (heating). Therefore, PJ-pattern could influence the dry band through altering the lower-level circulation while it is not reaching to wet band area in June. The high correlation between PJ-index and CGTI-PC1 is possibly due to the Rossby wave breaking induced by the CGT helping to promote convective activity

near the Philippines and the associated PJ-pattern (Takemura & Mukougawa, 2022). The correlation between daily CGTI-PC1 and the EASM wet band precipitation anomaly index is 0.15 ($p < 0.01$), with a partial correlation with removing the daily PJ-index still being 0.15 ($p < 0.01$). In conclusion, the PJ-pattern is possibly not involved in the subseasonal ISM-EASM wet band interaction, but it is one of the main candidates for the modulation of the EASM dry band at subseasonal time scale.

Table 2-3 The correlation coefficients between June subseasonal PJ-index, CGTI-PC1, precipitation indices of EASM wet band, EASM dry band, and ISM, from 1958 to 2015.

	EASM Npr	EASM Spr	ISM precip.	CGTI-PC1
PJ-index	0.03 ($p = 0.17$)	0.09 ($p < 0.01$)	0.04 ($p = 0.13$)	0.09 ($p < 0.01$)

(v) The symmetric influence of ISM heating

Similar to the method used to select the lag0 days (i.e., positive strong ISM heating), this subsection identifies the negative lag0 days (ISMH less than -1 K/day and with 5 or more consecutive days). The results obtained (**Figure 2-44**) are generally mirrored to those of positive ISMH (**Figure 2-9**), denoting the influence of ISM heating is symmetric regardless of the cooling or heating.

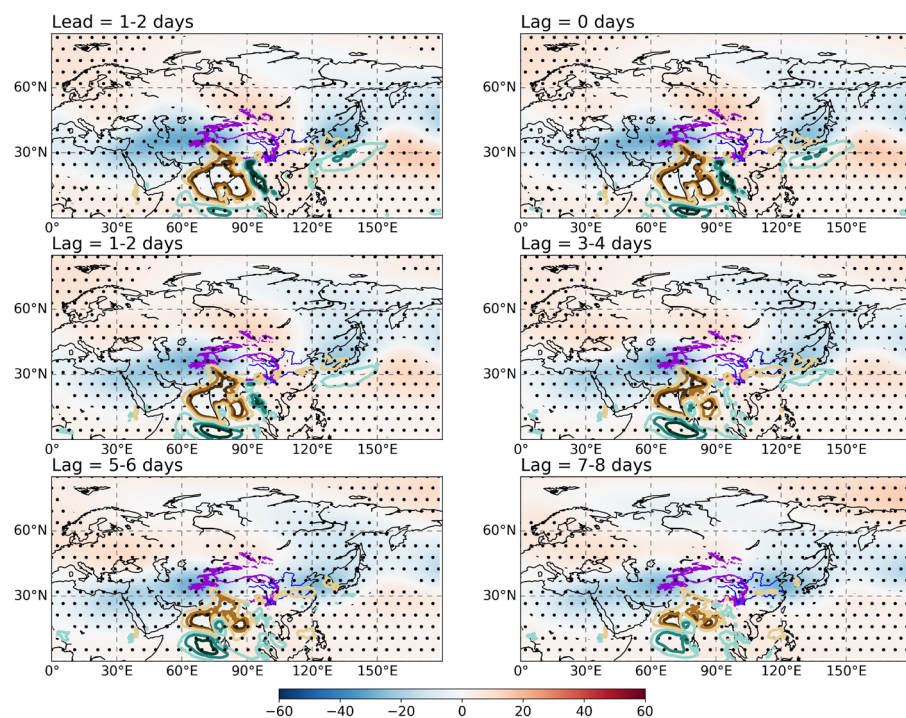


Figure 2-44 Simulated lead-lag characteristics of circulation and precipitation relative to below-normal Indian summer monsoon heating. Lead-lag composite at lead of 1–2 days, lag of 0 days, lag of 1–2 days, and lag of 7–8 days based on HIST. Shading stands for the 200-hPa geopotential height anomalies, contour indicates precipitation anomalies (unit: mm/day; with intervals of ± 1 , ± 2 , and ± 4 , positive in green contours while negative in brown contours). Dotted areas denote anomalies with $p < 0.01$ based on the two-sided Student's t test.

2.6. Summary and Conclusions

The investigation in [section 2.4](#) delves into the connections and relationships between the ISM, EASM, and the mid-latitude CGT on a subseasonal timescale, utilizing observational data and a large ensemble AGCM experiment (HIST). The results highlight the pivotal role of ISM's diabatic heating in shaping local CGT wave train that propagates towards East Asia and consequent influence of precipitation patterns over central East China, the Korean Peninsula, and southern Japan (i.e., the wet band). Meanwhile, a dry band on its south is seen, but it is less fluctuated and therefore is conjectured not influenced by the ISM.

The detailed sequence of ISM–EASM connection can be summarized as follows. The higher latitude CGT wave train travels from Western Europe into the ISM region, helping to establish the ISM convection, and the East Asian sector is not yet influenced by higher latitude CGT wave train at this point. The heating released by the condensation of the ISM precipitation subsequently initiated a local CGT wave train, which is then trapped within the westerly jet stream and resonating with a higher latitude CGT wave train. Thereafter, it propagates from the Indian subcontinent towards East Asia. At this stage, the Rossby wave in the westerly jet stream is orthogonal to the CGT pattern, with a clear wave phase difference. A few days later, the Rossby wave takes on a CGT pattern again, with more energetically in the East Asia, and extends horizontally in the westerly jet stream. The CGT index displays a distinct bimodality (high-low-high) during these progresses, and the power spectrum analysis on it suggests the quasi-biweekly oscillatory behavior of CGT. East Asia is predominantly influenced by mid-tropospheric warm advection and lower-tropospheric abundant moisture transport, which emerges as a critical factor in wet band formation. This is in response to the prevailing meridional winds established by the CGT. It is found that the CGT acts as a bridge between the ISM and EASM, and is originated from western Europe, interacts with the ISM diabatic heating, propagates to East Asia, and influences the East Asian precipitation.

Using K-means clustering, the interaction between ISM heating and the subtropical jet stream is illustrated. Four jet stream patterns are created by K-means clustering, then their differential impact on East Asian precipitation is investigated. The results suggest that ISM heating act as a trigger or reinforcer of the EASM rain band, depending on whether the meridional wind anomaly is preexisted prior than the strong ISM heating. This is confirmed by high and positive correlation between the EASM wet band and the southerly wind from the preceding days, as well as the lagged eastward horizontal wind. These factors are served as the trigger of EASM and the sign of the existed EASM wet band, respectively. The flat/wavy East Asian jet might be a sign of the EASM rainband occurrence/absent.

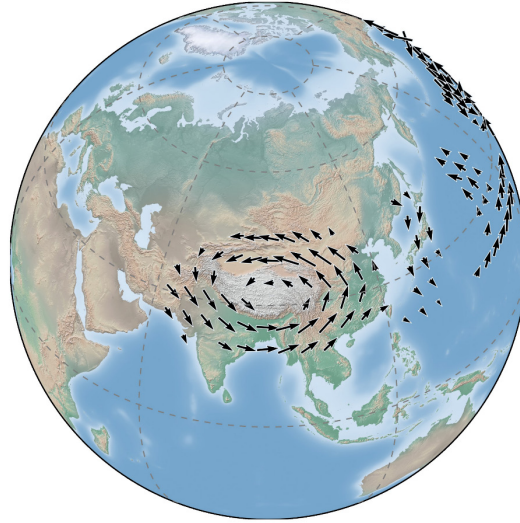
Meanwhile, there are evidence suggesting that the ISM–EASM connection that presented in the dissertation is less influenced by both ENSO and ISO. The reason is explained briefly as follows. The use of a lead–lag composite for ENSO conditions and the application of bandpass filters (30–60-day and 10–20-day) to precipitation demonstrate that the formation of East Asian wet band remains robust in the former case (i.e., ISM–EASM connection is independent of ENSO), while it seems to diminish in the latter (i.e., ISM–EASM connection is less influenced by ISO). Furthermore, we found that the EASM's wet and dry bands tend to occur simultaneously, indicating a preference for their concurrent occurrence. This helps explain the reason that previous research illustrated the good correspondence of the onset dates between the ISM and

EASM (see [section 1.2.4](#)). This is because the ISM heating activated a wet band over central East China, the Korean Peninsula, and southern Japan, which led to these regions entering the EASM season. (i.e., northward migration of the rainband).

The findings of this work provide a basis for ISM–EASM connection on the subseasonal timescale. More importantly, the large-ensemble simulation enables comparison of the temporal evolution of the atmospheric fields among various jet stream features, thereby enhancing understanding of the hydrological cycle in East Asia and benefitting subseasonal forecasting of heavy rainfall of the EASM. The present study leaves certain issues that appears important to be understood in the future; for instance, further studies are needed on the factors that control the dry band over East Asia, and the processes behind ISM heating including how its timespan from initiation to peak is determined, as well as how the other impacts are possibly involved in ISM–EASM connection.

3

Mid-latitude forcing II: TP



Modulation of Tibetan Plateau thermal forcing on East Asian summer monsoon

Brief summary

This chapter investigates atmospheric teleconnections in response to Tibetan Plateau (TP) land memories during early summer. Land surface processes exert an important role in energy exchange between land and atmosphere, influencing the climate system and its predictability. We explore the physical process of how TP land surface temperature anomalies in May influence the subsequent June precipitation in East Asia and global atmospheric circulations. The maximum covariance analysis for the reanalysis data confirms the possibility of circum-global wave train that travels from TP to North America in both May and June according to TP temperature anomalies. The cold land anomalies in TP, in particular in central to western TP, play a crucial role in the excitation of the wave train as soil temperature anomaly tends to persist from May to June due to memory effect of the soil. A linear baroclinic model experiment and the analysis of large ensemble GCM experiment demonstrate that TP temperature anomaly can emanate the circum-global teleconnection in phase with the one detected in the reanalysis data.

3. Impact of Tibetan Plateau thermal forcing on the modulation of East Asian summer monsoon

3.1. Introduction

Land surface processes play a crucial role in the climate system and its prediction. Soil moisture, soil temperature, snow cover, and vegetation states are important land attributes which controls energy transfers between land and atmosphere that significantly impact the climate system (Diallo et al., 2022; Erdenebat & Sato, 2016; Koster et al., 2004; Nakamura et al., 2019; Portmann et al., 2022; Sato & Nakamura, 2019; Xue et al., 2018, 2021; Yasunari et al., 1991). The slowly evolving states of the land surface processes regulates surface hydroclimate and subsequently enhance climate predictability through land-atmosphere interactions (Flanner, 2005; Z. Guo et al., 2006; Koster et al., 2006). Land-atmosphere interactions are not only considered to be dominantly local but also have nonlocal effects, affecting downstream or remote weather and climate through various mechanisms. The nonlocal impacts of anomalous land surface anomalies have been highlighted in recent studies, demonstrating that land surface processes can influence both local and remote regions through the generation of the Rossby wave train and associated circulation anomalies, which can have a delayed downstream influence on the weather and climate (W. Li et al., 2018; Teng et al., 2019; H. Xu et al., 2022; Jingqiu Yang et al., 2021). In recent years, the popularity of studying land surface processes in the climate system has grown due to advancements in modeling capabilities, improved observations, and an increased understanding of the role of land in climate variability (Robertson & Vitart, 2018). Understanding and accurately representing land surface processes are crucial for improving climate predictions. Thus, capturing the complex interactions between land and atmosphere will lead to more skillful forecasts and a better understanding of the climate system.

The Tibetan Plateau (TP), with the highest elevation on Earth, its thermal forcing associated with land surface processes has long been considered to dominate the weather and climate of its surrounding regions (Yuzhi Liu et al., 2020; Sato, 2005; G. Wu et al., 2015). The interaction between TP and drought/flood events of the East Asian summer monsoon (EASM) has received much attention as one of the major challenges in climate prediction (Y. Zhang et al., 2019). Through an idealized aqua planet experiment with Eurasia continent and TP, J.-H. Son et al., (2019) found the thermal forcing over TP tuned by altering TP surface albedo can modulate the East Asian precipitation. The thermal forcing of TP could explain approximately 14% of the EASM precipitation intensity, which is as substantial as the land-sea thermal contrast, although the TP orographic forcing might be the primary factor of the monsoon formation. This conclusion in terms of the relationship between TP thermal forcing and EASM precipitation, is consistent with the findings in Wang et al., (2008) which suggested that observed rising surface air temperature over TP under ongoing climate change could lead to lesser precipitation in North China while strengthening precipitation belt over the region from South China to Korean Peninsula. Their AGCM and LBM experiment found that TP temperature can induce upper- and lower-level Rossby waves toward East Asia. Majority of earlier efforts have devoted to study EASM activity and physical processes behind it in mid-summer when EASM matures. However, recent studies highlighted the possibility of seasonal prediction of EASM activity because land surface processes of TP establish a unique seasonal transition from spring to early summer as it lies in mid-latitude and its high altitude. For instance,

the new evidence indicates spring-to-summer TP thermal forcing could be controlled by subsurface temperature (SUBT; Diallo et al., 2022; Xue et al., 2018, 2021). It was found that imposed cold TP SUBT in the model acts as persistent cold forcing over TP, enabling a delayed impact on EASM precipitation in a case study of year 2003. The substantial heat anomaly stored in the subsurface layer over the TP allows to retain the surface thermal anomalies and gradually release it in the subsequent months (Ye Liu et al., 2020). If such seasonal transition involves the phase change of soil moisture, anomalous thermal forcing can persist more efficiently from spring to summer (Xu et al., 2022).

Here, we aim to study physical process of how May land surface temperature anomaly affects the subsequent June atmospheric circulations which is known to have potential impact on East Asian precipitation, based on the analysis of multiple events under different climatic conditions. In particular, we focus on how the TP SUBT anomalies could persist within the land surface-atmosphere coupling system.

3.2. Objectives

The anomalies of the cold TP soil imposed in the numerical model proved to be able to add value to the EASM forecast for the year 2003, based on the discussion of [section 1.3.3](#) and [section 3.1](#). These results provide an insightful understanding of initialization of land component in enhancing the EASM prediction, which motivates more researchers start from 2003 case study to explore a general relationship of TP soil enthalpy and EASM, which may require more observational data on this aspect to reveal their general relationship.

Here, we provide the observation evidence of how May TP land surface temperature anomaly affects the subsequent June atmospheric circulations and EASM, and its possible underlying physical processes.

3.3. Data and Methods

The monthly and pentad mean atmospheric circulation and hydrological parameters were derived from JRA-55, comprising analyzed variables of temperature, geopotential height, wind speed, and total column integrated water vapor flux. The monthly and pentad mean soil temperature, sensible heat flux, and latent flux data from ERA5-Land (Muñoz-Sabater et al., 2021) are used, with a horizontal resolution of 9 km and the used time span of 1979–2020, which provides valuable insights of spatiotemporal land-atmosphere interaction over TP region. The ERA5-Land features a four-layer representation of soil, with layer depths of 0–7, 7–28, 28–100, and 100–289 cm. In situ monthly soil temperature data from 15 meteorological stations with depths of 0, 5, 10, 15, 20, 40, 80, 160, and 320 cm over the period of 1981–2005, which was provided by Xue et al. (2021) are also used. Gridded ($0.5^\circ \times 0.5^\circ$) monthly 2-m air temperature data (T2m) were obtained from the Climate Anomaly Monitory System (GNCH_CAMS; Fan & van den Dool, 2008), and monthly precipitation GPCP are analyzed from 1979–2020. The monthly mean snow cover is analyzed from 1979–2020 from the National Oceanic and Atmospheric Administration (NOAA) Northern Hemisphere snow cover extent climate data record (NH SCE CDR; Estilow et al., 2015). The pentad mean daily

T2m maximum (T2max) and minimum (T2min) data from NOAA climate prediction center (CPC) are also analyzed as from 1979–2020 (CPC Global Unified Temperature; Xie et al., 2007). The values of snow cover extent and temperature related variable are linearly detrended to eliminate the impact of global warming.

3.4. Results

3.4.1. Strategy of singular value decomposition

In this section, to focus on the potential influence of land memories on the EASM, the autocorrelation of TP T2m areal mean index (25°–40°N, 70°–105°E with elevation above 2500 m) is computed from 1979 to 2020, with a linear detrend to eliminate global warming (**Figure 3-1**). The highest autocorrelation occurs between May and June ($r = 0.46$, $p < 0.01$), and is then mostly concentrated in the winter season (September to February; with $r > 0.40$, $p < 0.01$). In winter season, the vast majority of the area of TP are covered by snowpack. Due to its high elevation and associated lower temperatures, once the snow cover formed, it is able to persist across the cold seasons, which is known as an important factor of the persistence of snow cover anomalies and associated temperature anomalies (Z. Wang et al., 2019). Therefore, it is not surprising that winter TP T2m shows a high and positive autocorrelation.

However, the snow cover mostly vanishes in the late spring (April to June), with the snow cover is limited to western TP in May (Xiao & Duan, 2016). For this reason, it is worth to explore the possible mechanism of the high autocorrelation between May and June. It is speculated that this is because the soil memory arises from the soil enthalpy, and it might be the important factor of the delayed impact of TP T2m on EASM.

The SVD analysis was conducted to assess the relationship between May TP T2m and EASM precipitation. The comparisons included May TP T2m versus concurrent May EASM precipitation (i.e., same month), May TP T2m versus EASM precipitation at different lag of months (1–3) as well as summer mean (June to August) EASM precipitation. This SVD map (**Figure 3-2**) created by homogeneous regression shows the May TP T2m is strongly coupled with EASM precipitation in the concurrent and subsequent few months, with various EASM precipitation patterns. It should be noted that, it is found that the SVD between May TP T2m and June EASM precipitation present an ISM–EASM connection, implying a relation to the investigation in the [chapter 2](#). In conjunction with the high autocorrelation of TP T2m between May and June, the relationship between May TP T2m and June EASM precipitation is determined as a main target for investigation.

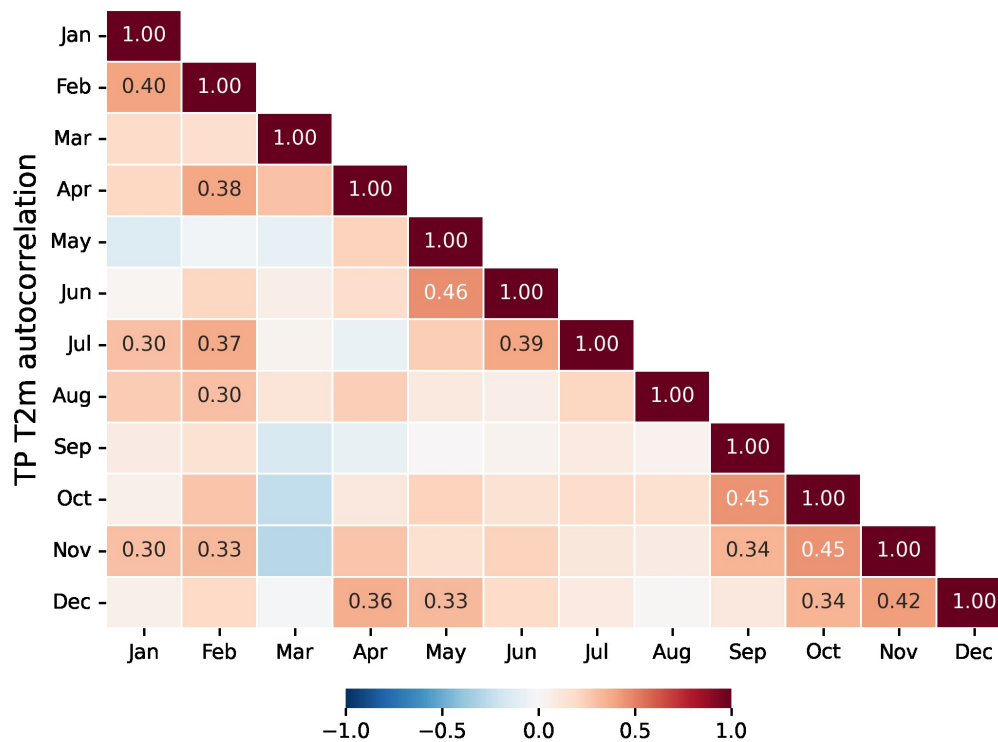


Figure 3-1 The autocorrelation of TP T2m. The heat map showing the autocorrelation of detrended TP T2m from 1979 to 2020, where the values denote label correlation coefficient with $p < 0.05$.

Based on this result, the T2m from TP region (defined as 25°–50°N, 70°–105°E with elevation above 2500 m) and East Asian precipitation from domain of 20°–40°N, 105°–135°E from 1979 to 2020 are inputted as the left and right fields of the SVD, respectively. SVD is an extension of principal component analysis. The spatial patterns obtained from SVD represent coherent spatial structures that capture the dominant modes of variability, while the corresponding principal components represent the associated temporal behavior of each spatial pattern. The homogenous regression is used to show internal relations within a field by regression upon its own principal component score.

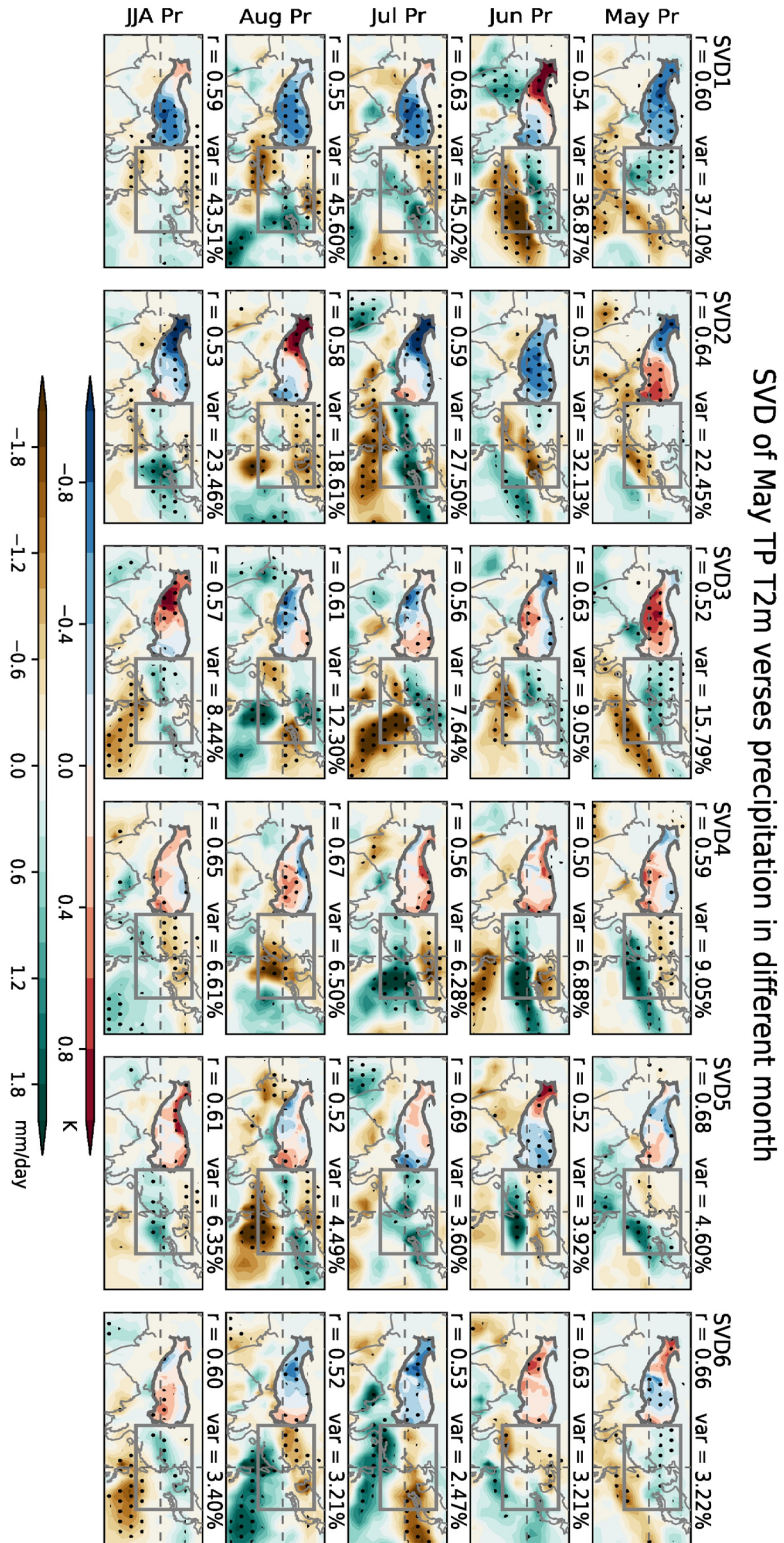


Figure 3-2 The SVD between TP T2m and EASM in summer. The homogeneous regression pattern of 6 leading SVD modes (columns) for May TP T2m versus EASM precipitation at different lag (0–3) of months. Stippled regions indicate anomalies that are statistically significant at the 95% confidence level based on the two-sided Student's t test. The grey contour stands for the plotted domain and SVD input domain for TP, and rectangle indicated the input domain for EASM precipitation.

3.4.2. The coupled pattern revealed by the singular value decomposition

The two leading modes of the SVD reveal the distinct differential regional influences, i.e., surface air temperature anomalies in different areas of the TP coupled to different precipitation patterns in East Asia. In **Figure 3-3** the temperature (left singular vector) in the first mode shows the potential impact from the anomalous warm western TP, where the temperature anomalies are seemingly embedded in the large-scale wave train-like forcing across the Eurasia. The second mode, instead, shows the cold temperature over the entire TP (**Figure 3-4a**). The explained variances for the first and second modes are 36.87% and 32.13%, respectively, meaning their contributions are comparable. Meanwhile, the explained variance of the third mode sharply reduced to 9.05%. Therefore, the first two modes are the most dominant modes at interannual time scale. Here, the second SVD mode is selected for further investigation, since we focus more on the thermal forcing from the TP.

The second SVD mode (hereafter, SVD2) shows cold TP coupled to a horizontal “wet-dry-wet” pattern in East Asia (**Figure 3-4**), with greater precipitation in the Yangtze River basin and the northwestern Pacific, and less precipitation in southern China and southern Japan. The scores between May TP T2m (i.e., left singular vector) and June East Asian precipitation (i.e., right singular vector) are significantly correlated, with correlation coefficient of 0.55 ($p < 0.01$). The scores for the left singular vector and the right singular vector of the SVD2 are 0.69 and 1.72 for the year 2003, and -2.39 and -1.75 for the year 1998, respectively. These scores indicate our analysis can successfully extract the year of typical TP forcing investigated in Xue et al., (2018) and Diallo et al., (2022). To understand the coupling mechanism behind the SVD2, the following sections will conduct a composite analysis.

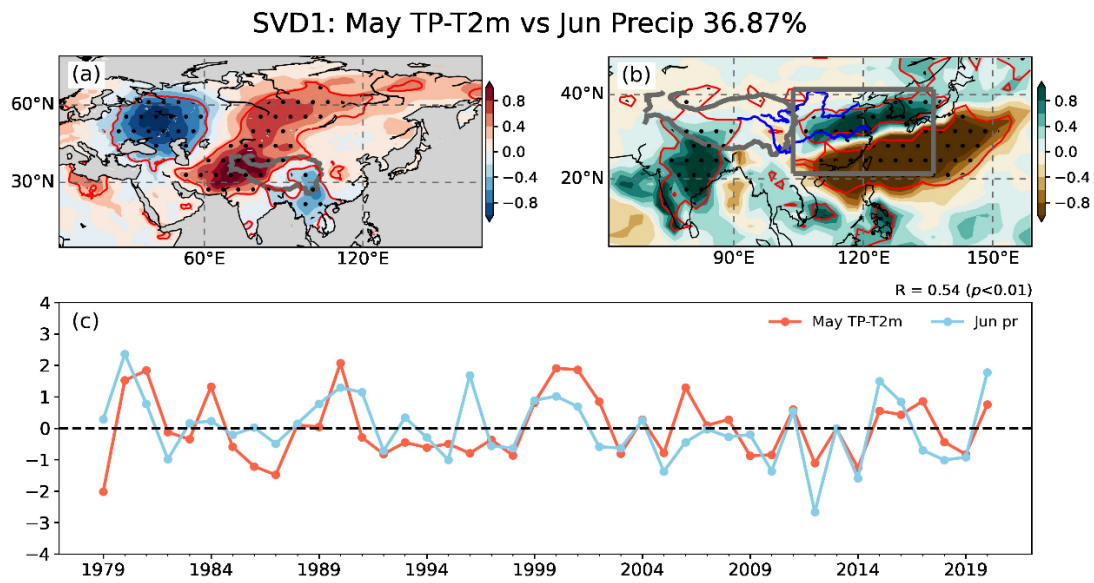


Figure 3-3 The leading coupling mode between May TP T2m and June EASM. The homogeneous regression pattern of second SVD mode of (a) T2m anomalies and (b) precipitation anomalies over East Asia and (c) their corresponding principal component scores. Red contour and dotted regions denote anomalies significant at the 90% and 95% confidence level, respectively, based on the two-sided Student's *t* test. The grey contour in (a) and grey rectangle in (b) denote the domain of May TP T2m and June East Asian precipitation, respectively, inputted to the SVD analysis. Blue lines in (b) represent the Yellow River (north) and the Yangtze River (south).

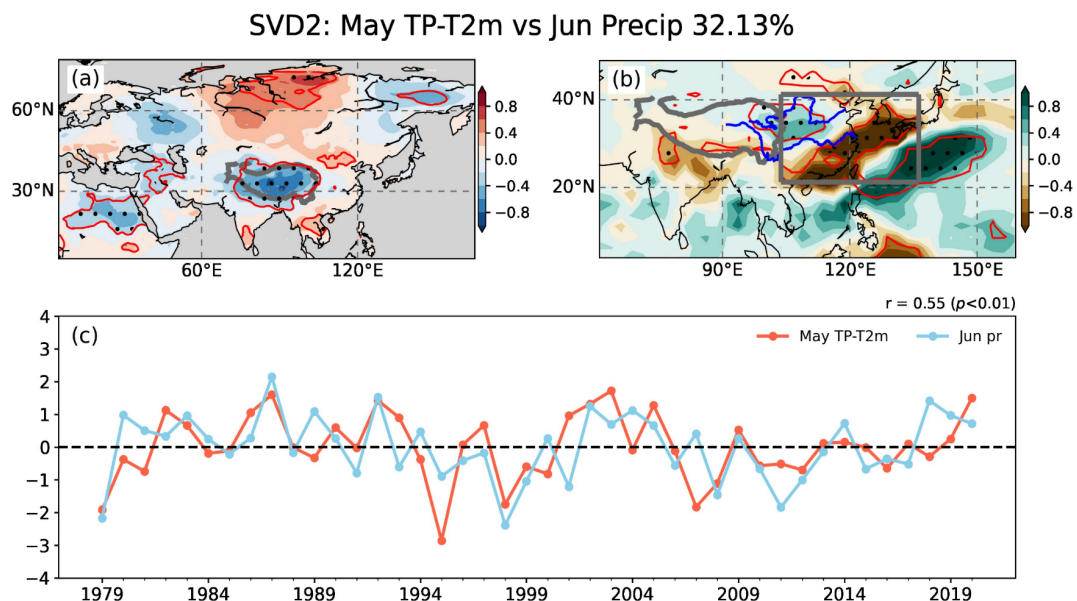


Figure 3-4 The second coupling mode between May TP T2m and June EASM. As in Figure 3-3, but for second SVD mode.

3.4.3. Land surface features and large-scale circulation features

According to SVD2, the composite analysis is conducted based on SVD score. We selected years whose scores for the left singular vector and the right singular vector are both greater than 0.5 or both lower than -0.5 (totally 15 years are selected), and composite was made as the difference of averages between the former (greater than 0.5) and latter (lower than -0.5). It is found that cold TP in May is accompanied by a strong negative geopotential height anomaly at 200-hPa over TP region (**Figure 3-5b**). The associated cyclonic circulation near TP region might serve as a wave source and excite a wave train originating from TP and travelling across northeast China, far eastern Eurasia, the North Pacific Ocean, and finally terminating to US near the Rocky Mountain (**Figure 3-5a**). This wave train indeed resembles to the one known as TP-Rocky Mountain Circum-global (TRC) wave train proposed by Xue et al. (2022). The TRC wave train further persists until June, but notably the wave train appears to be initiated in the western TP region rather than the entire TP as that in May **Figure 3-6d**). As a high elevation area, TP has a strong and effective land memories (Ye Liu et al., 2020; Y. Zhang et al., 2019). The persistent TRC wave train is speculated to be maintained by TP soil memories, rather than snow cover (**Figure 3-5**), which is elaborated in the next section.

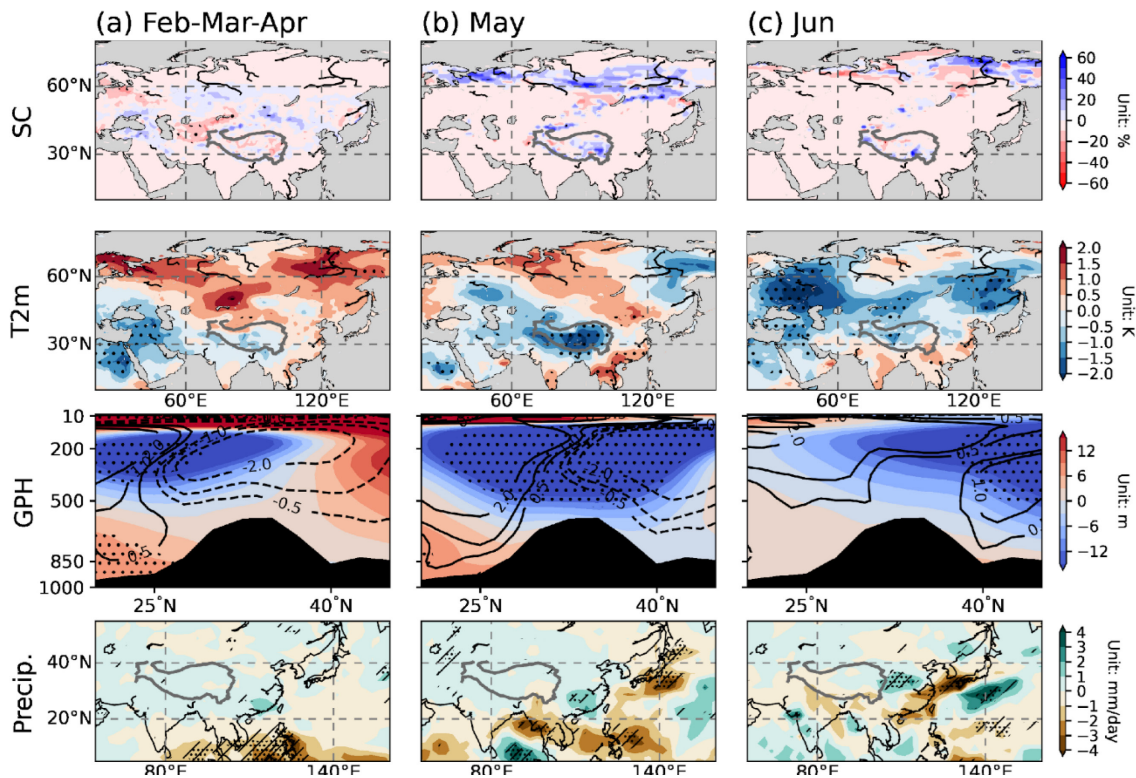


Figure 3-5 The composite land surface conditions of SVD2. The composite difference about SVD2 scores between high ($> +0.5$) and low (< -0.5) years of snow cover, T2m, cross section of geopotential height (zonal mean from 70° – 105° E), and precipitation (rows from top to bottom, respectively). (a) averaged from February to April, (b) May, and (c) June. The stippled and hatched regions denote anomalies significant at the 90% and 95% confidence level, respectively, based on the two-sided Student's t test.

Meanwhile, there is no clear signal of lower-level circulation for East Asia in May. This might be attributable to the latitudinal position of the wave train pathway which propagates far north from East Asia in May (**Figure 3-6a**). In contrast, in June Japan and its surrounding oceans are dominated by the negative geopotential height anomalies in both the upper and lower troposphere, corresponding to a part of TRC pathway from TP toward East Asia. The negative geopotential height anomalies in the lower troposphere drive the water vapor flux from Japan toward southeast China. As this direction is opposite to the climatological water vapor transport, precipitation is reduced in East Asia (**Figure 3-6b**, East Asian dry band). At the same time, the Yangtze River basin is covered by significant negative potential vorticity anomalies in the upper troposphere (**Figure 3-6d**), which appear to be linked dynamically to the upper-level divergence, creating a suitable condition to enhance precipitation (i.e., **Figure 3-6**, positive precipitation anomalies near the Yangtze River basin).

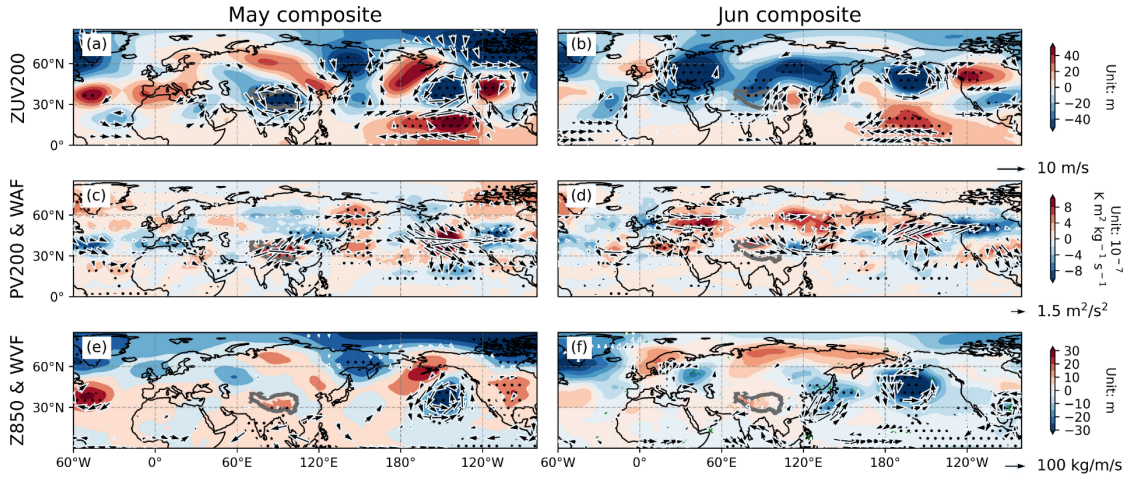


Figure 3-6 The composite circulation patterns of SVD2. The composite difference about SVD2 scores between high ($> +0.5$) and low (< -0.5) years of (a) May 200-hPa geopotential height (shading) and horizontal wind (arrows), and (c) May 200-hPa potential vorticity anomalies (shading) and wave activity flux (arrows), and (e) May 850-hPa anomalies of geopotential height (shading) and water vapor flux (arrows). (b), (d), and (f) are similar to (a), (c), and (e) but for the composite of June. Dotted regions denote anomalies significant at the 90% confidence level, respectively, based on the two-sided Student's *t* test. The insignificant vectors ($p > 0.1$) of wind and water vapor flux are omitted.

3.4.4. Role of the Tibetan Plateau soil temperature

Possible interactions between the atmosphere and land over TP are presented in **Figure 3-7** which illustrates vertical profile of area-averaged temperature over TP region (i.e., 25° – 50° N, 70° – 105° E). The rapid cooling of the tropospheric and surface air above the TP occurred in late April and early May. This is presumably owing to the negative geopotential height anomalies that dominated over TP as shown in **Figure 3-5** and **Figure 3-6**. The anomalous atmospheric forcing of colder air temperatures above

TP gradually influences the land surface and penetrates deep into the soil layers, eventually leading to the TP land surface and soil temperatures at depths of up to 1 meter significantly colder in May (**Figure 3-7** and **Figure 3-9**). The anomalous atmospheric temperature shrinks rapidly after late May and ceases by early June. In contrast, the soil temperature anomaly remained to be negative until the end of June. These processes can be interpreted as that the cold air temperature in May is memorized in the soil over TP, in particular within central-western TP (**Figure 3-9**). The low soil temperature in western TP and associated turbulent heat flux at the surface (sensible heat flux plus latent heat flux) might in turn influence the local circulation over TP in mid-June, forcing the atmosphere to cool down again. In particular, the mid-June sensible heat flux over western TP is nearly insignificant, while the anomalies of the turbulent heat flux arise from the greater contribution of latent heat flux. Such latent heat flux in mid-June is expected to help TP surface cooling and potentially excite and maintain the TRC wave train toward East Asia (**Figure 3-9**).

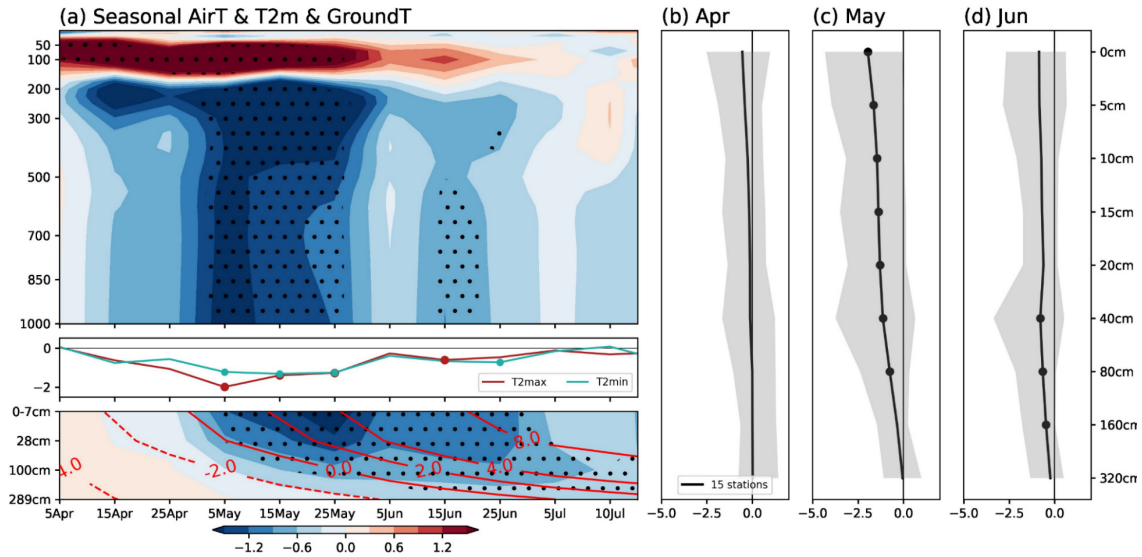


Figure 3-7 The TP soil memories of SVD2. Soil to air vertical profile of area-averaged temperatures over TP region (25° – 50° N, 70° – 105° E and for areas above 2500 m elevation). Values represent composite difference about SVD2 scores between high ($> +0.5$) and low (< -0.5) years and the drawing is for every two pentads (10-day periods) for (a) air temperature (upper panel), daily maximum T2m together with daily minimum T2m (T2max and T2min; middle panel), and soil temperature (bottom panel, shading) with its climatological values (bottom panel, red contour). (b), (c), and (d) are vertical profile of soil temperature anomalies in April, May, and June, respectively, based on the average of 15 in situ stations. Dot on the cross sections and lines denote anomalies significant at the 90% confidence level, respectively, based on the two-sided Student's t test.

The vertical profile of diabatic heating (thermodynamic equation) and its associated terms over TP are examined to illustrate the large-scale land-atmosphere interaction (**Figure 3-8**). In May, the column of the atmospheric cooling is dominant by the evident anomalous temperature advection over the lower troposphere above TP (near 500-hPa; **Figure 3-8a**). While in June, interestingly, the cooling is not regulated by the temperature advection (**Figure 3-8b**). The negative vertical heat diffusion denotes the downward transport of heat, with the vertical pressure velocity being positive (downward motion). Therefore, the vertical pressure gradient of temperature anomaly is negative, which means temperature decreases with the pressure increase. This is speculated to be controlled by the cold land surface and soil. This situation suggests that warmer air is transported downward. This is common under stable atmospheric conditions where a layer of warmer air traps cooler air near the surface, leading to a downward heat transport. Actually, June TP is an anomalous heat sink, as shown by the negative diabatic heating (negative apparent heat source; **Figure 3-8b**).

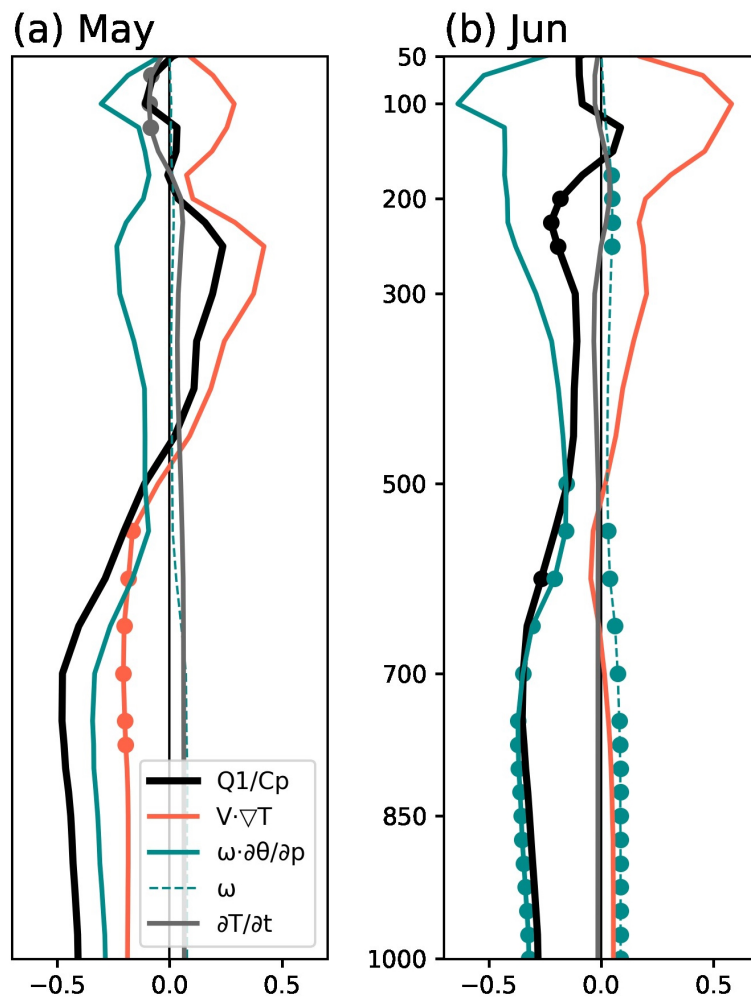


Figure 3-8 The vertical profile of TP diabatic heating and associated decomposed terms. The vertical profile of each term (K/day) from the thermodynamic equation over TP region (25°–50°N, 70°–105°E and for areas above 2500 m elevation), composite based on SVD2 of (a) May, and (b) June. Dot on lines denote anomalies significant at the 90% confidence level, based on the two-sided Student's t test.

The 15 stations over TP also support the land memory effect revealed above. The negative temperature anomaly in the soil layer is largest in May in consistent with the anomalous cold air temperature. The cold anomalies in the ground temperature persist until June, especially in the deep soil layers of 40–160 cm, due to the soil memories in that level. Meanwhile, we anticipate that the reanalysis (ERA5-Land) might overestimate the soil memories in the upper soil layers because soil temperature anomaly is insignificant in in-situ data (**Figure 3-7d**) unlikely to the reanalysis (**Figure 3-7a**). Xu et al. (2022) pointed out the overestimation of soil memory is associated with soil water phase as frozen soil with solid water has higher specific heat than the soil with liquid water. Since May corresponds to the period when climatological soil temperature varies across the freezing level (**Figure 3-7a**; red contour), soil water phase should be diagnosed accurately, otherwise the energy exchange within soil layer contains error. For that, more in situ observations with careful validation will be needed in the future.

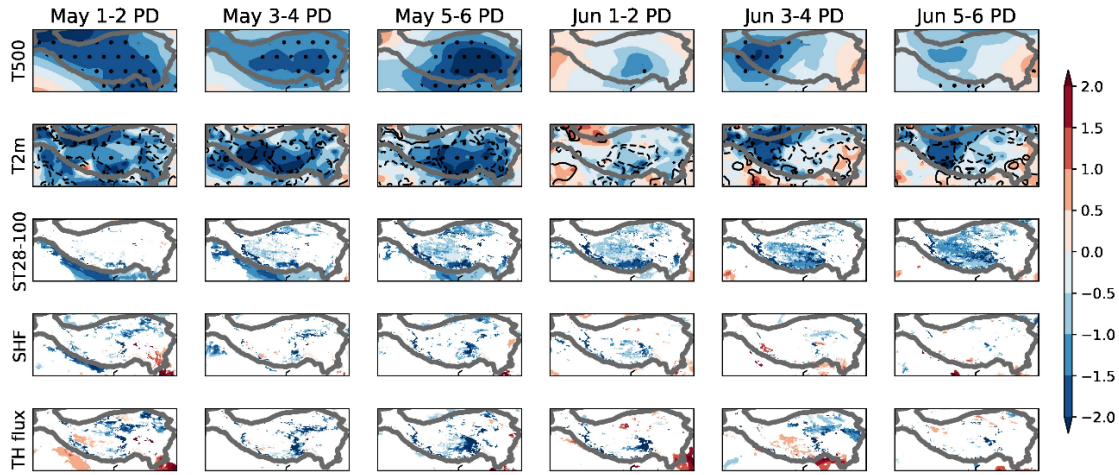


Figure 3-9 The TP soil memories of SVD2. The composite difference about SVD2 scores between high ($> +0.5$) and low (< -0.5) years of of 500-hPa temperature (unit: K), minimum T2m (shading; contour for maximum T2m; unit: K), soil temperature (unit: K), sensible heat flux (unit: W/m^2), and turbulent heat flux (unit: W/m^2 ; rows from top to bottom, respectively) displays for each 2 pentads (PD) from May to June (columns from left to right, respectively). The stippled regions denote anomalies significant at the 90% confidence level based on the two-sided Student's t test. The shading for soil temperature, sensible heat flux, and turbulent heat flux with $p < 0.1$.

Composite analysis about SVD2 also shows spatially homogeneous features of negative temperature anomalies in May of both atmosphere and land surface (**Figure 3-10**), and in situ observations mostly agree with this result. There are 5 out of 15 stations, distributed across the TP, showing significantly negative temperatures in May. Among these 5 stations, 3 of them still have significant anomalies in June. These stations mostly lie near central-western TP, in good agreement with SUBT anomalies investigated using the reanalysis data (**Figure 3-10**). The June monthly-mean fields for air temperature and T2m show cold anomalies near the central-western TP, suggesting

the dominance of land memories in this region (**Figure 3-10**).

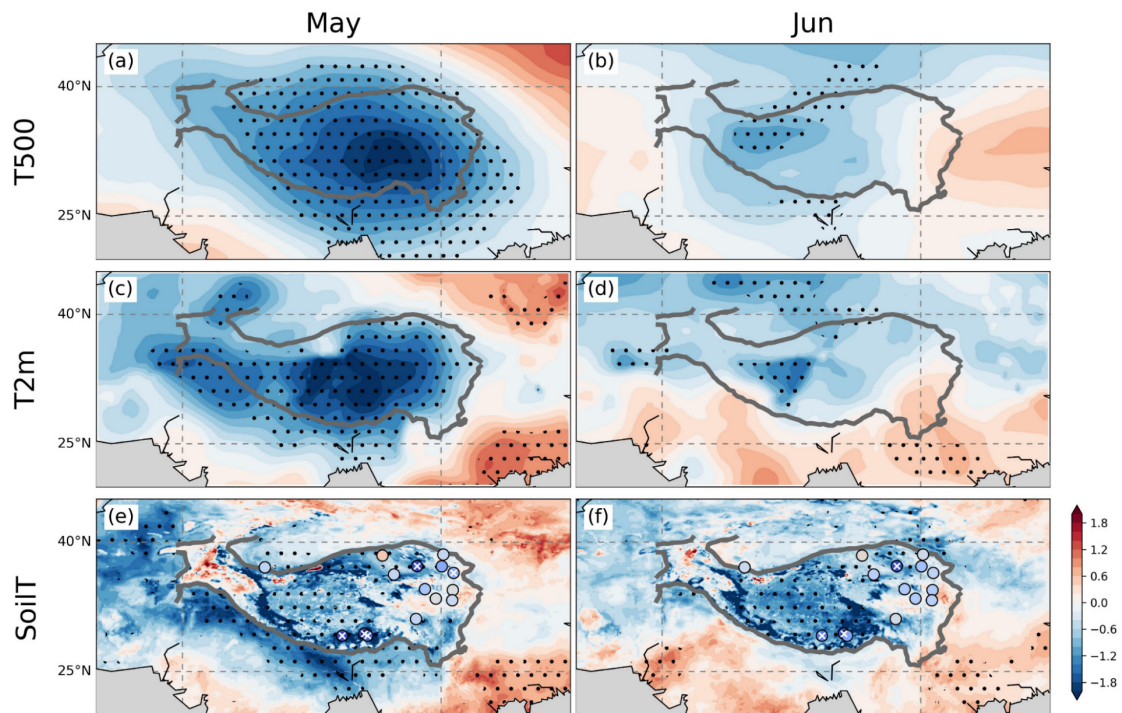


Figure 3-10 The TP soil memories of SVD2. Spatial pattern of composite difference about SVD2 scores between high ($> +0.5$) and low (< -0.5) years of (a) May 500-hPa temperature, (c) May T2m, and (e) May soil temperature of 7–100 cm (shading) and in situ soil temperature of 40–160 cm (circles). The (b), (d), and (f) are similar to (a), (c), and (e) but for the result from June. Dotted regions and circles with white crosses denote their difference significant at the 90% confidence level, respectively, based on the two-sided Student's *t* test.

3.5. The AGCM experiments

A sensitivity experiment is also carried out by using an AGCM named Model for Prediction Across Scales-Atmosphere (MPAS-Atmosphere), version 8.0 (Skamarock et al., 2012). The MPAS-Atmosphere are coupled with Noah land surface model (F. Chen & Dudhia, 2001; Ek et al., 2003), with totally 4 vertical soil layers of 10, 30, 60, and 100 cm thick. The MPAS-Atmosphere is set with a horizontal resolution of approximately 120 km and 37 vertical levels of the atmosphere. The boundary conditions were generated from COBE-SST2 monthly mean data (Hirahara et al., 2014), and the initial conditions were from National Centers for Environmental Prediction final analysis (NCEP-FNL) data sets (NCEP, 2000). The simulations were integrated for about 2 months from 30 April to 31 July in each year, from 2000 to 2012.

Two simulations were conducted assuming the cold TP (EXP_CTP) and the warm TP (EXP_WTP) conditions. The soil temperature anomalies obtained from **Figure 3-7c** is imposed at the first model integration step over the TP region, with the imposed

values are doubled (**Figure 3-11a**), following previous study (H. Xu et al., 2022; Xue et al., 2018). Meanwhile, the soil water phases (i.e., liquid or ice) are altered accordingly based on the soil temperature is above 273.15K or not (**Figure 3-11b**).

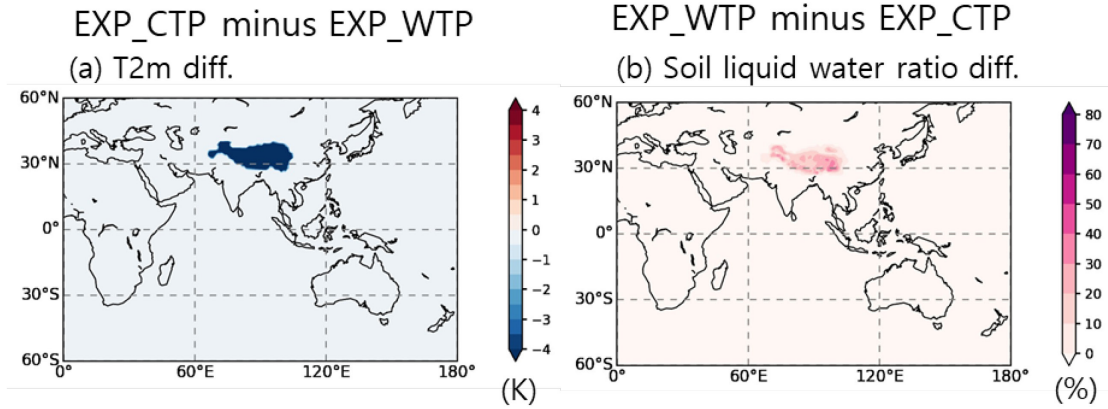


Figure 3-11 The difference of soil conditions between EXP_CTP and EXP_WTP in the first model integration step. Spatial pattern of first model integration step of (a) T2m difference using EXP_CTP minus EXP_WTP, and (b) ratio of soil liquid water relative to total soil water difference using EXP_WTP minus EXP_CTP. The patterns are exact same for simulation in each year from 2000 to 2012.

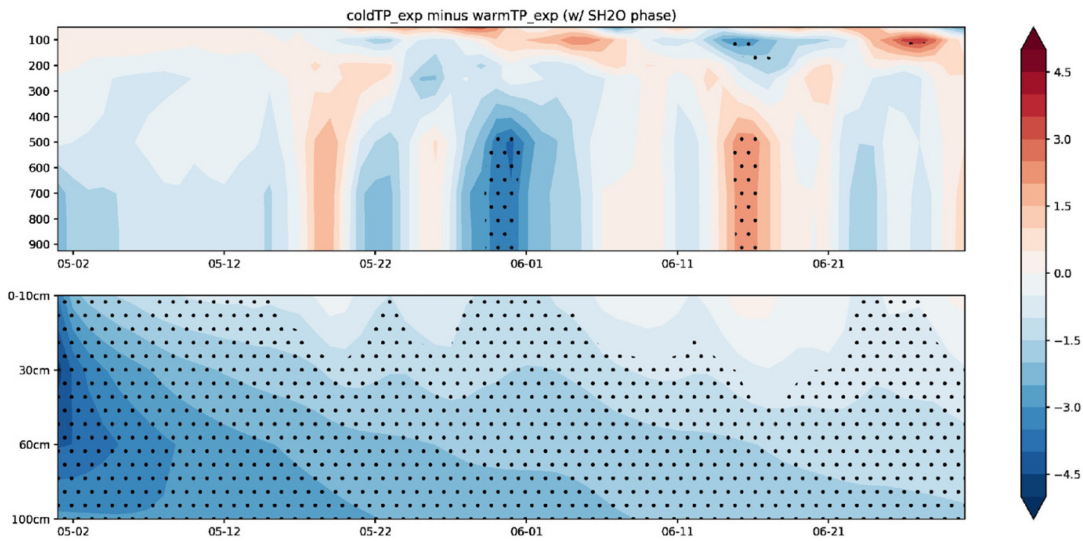


Figure 3-12 The difference of air and soil temperature between EXP_CTP and EXP_WTP. Time evolution of TP air temperature (upper panel) and soil temperature (lower panel) difference from 1 May to 31 July, using EXP_CTP minus EXP_WTP from 13 ensemble members (i.e., from 2000 to 2012). Dotted regions and circles with white crosses denote their difference significant at the 90% confidence level, respectively, based on the two-sided Student's t test.

Based on the simulations from 2000 to 2012, i.e., a total of 13 ensemble members are created, considering different atmospheric circulation coupled with abnormal TP land and soil temperatures. The results show that the imposed temperature forcing is stored in the soil, while the air temperature is not significantly cold in May (**Figure 3-12**). These results indicate that the current results of the MPAS-atmosphere will need to be carefully tuned in the future. (Xue et al., 2021).

3.6. Discussions

In this section, we discuss how TP land memories could force the TRC wave train. Related to the cold anomalies over the entire TP in May and over the central-western TP in June, the upper atmospheric circulation shows positive vorticity anomalies over the whole TP and over the central-western TP for May and June, respectively (**Figure 3-10a and b**). **Figure 3-13** shows the vertical distributions of the vorticity differences between May and June. The largest anomaly of vorticity in May appears near 200-hPa and extends across entire TP, while those for June thicken downward up to near 400-hPa, and are mostly located near the central-western TP. These features again indicate that the formation of the vorticity anomalies is different according to the monthly evolution, as we speculated in [section 3.4.3](#). Meanwhile, the 700-hPa vorticity pattern shows a horizontal tripolar vorticity anomaly centered over the Yangtze River basin, southern China–Southern Japan, and northwestern Pacific (**Figure 3-13d**). This pattern indicates the lower-level convergence in the Yangtze River basin and Northwest Pacific while divergence extend from the south China to south Japan. Therefore, it is suggested low-level circulation also dynamically plays a role on the precipitation anomalies.

We performed the LBM (Watanabe & Kimoto, 2000) experiments here to provide more information between the anomalous June vorticity in the central-western TP and the wave train. The LBM was set to the resolution of T42L20, which is 128×64 grids horizontally and 20 sigma levels vertically. The basic state used is the June climatological values averaged from 1979 to 2020 derived from JRA-55. The idealized vorticity forcing is prescribed in the LBM (**Figure 3-14**), based on our observational results from **Figure 3-13**.

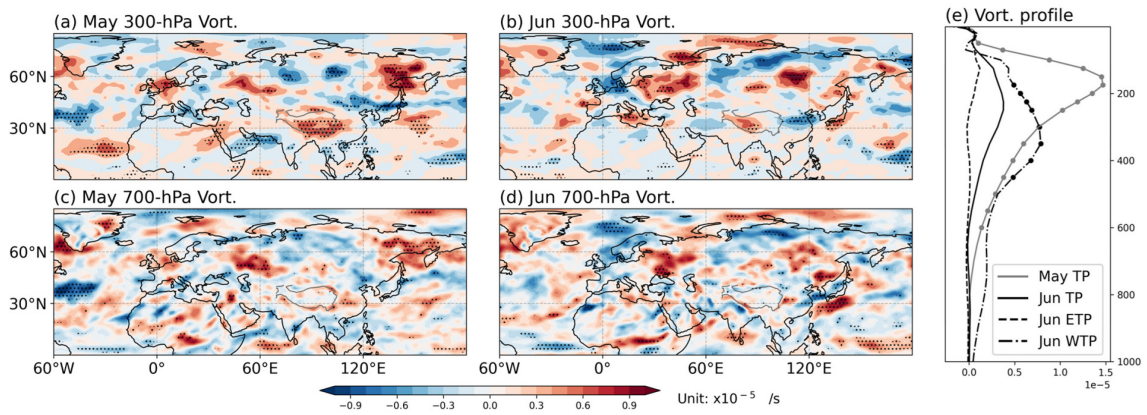


Figure 3-13 The vorticity field related to SVD2. Spatial pattern of composite vorticity difference about SVD2 scores between high ($> +0.5$) and low (< -0.5) years of (a) May 300-hPa vorticity, and (b) June 300-hPa vorticity. The (c) and (d) are similar to (a) and (b), but for 700-hPa vorticity. (e) shows the vertical profile of vorticity area-averaged from entire TP (25° – 40° N, 70° – 105° E), Eastern TP (ETP; 25° – 40° N, 90° – 105° E) and central-western TP (WTP; 25° – 30° N, 80° – 90° E) with elevation above 2500 m. Dotted regions and dotted lines denote anomalies significant at the 90% confidence level, respectively, based on the two-sided Student's *t* test.

The LBM results show that in response to the western TP vorticity forcing, a TRC-like wave train is excited in the upper troposphere (**Figure 3-15**). Interestingly, the wave train simulated by LBM is similar to the one obtained from the composite analysis about SVD2 in **Figure 3-13**. Their common features are cyclonic circulations at the central-western TP, southern Japan, and the US continent near the Rocky Mountains, and the anticyclonic circulation near the Yangtze River basin and the western coast of Canada. Their similarity confirms the role of the land memory effect in maintaining the TRC-like wave train over two months and its possible impact on the downstream weather and climate.

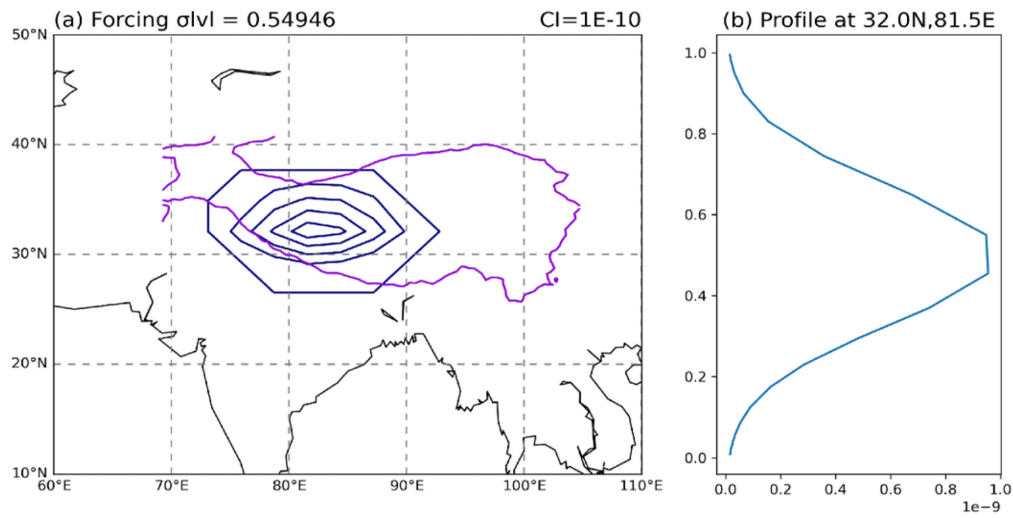


Figure 3-14 The vorticity forcing used in LBM. The idealized vorticity pattern as forcing that used in LBM.

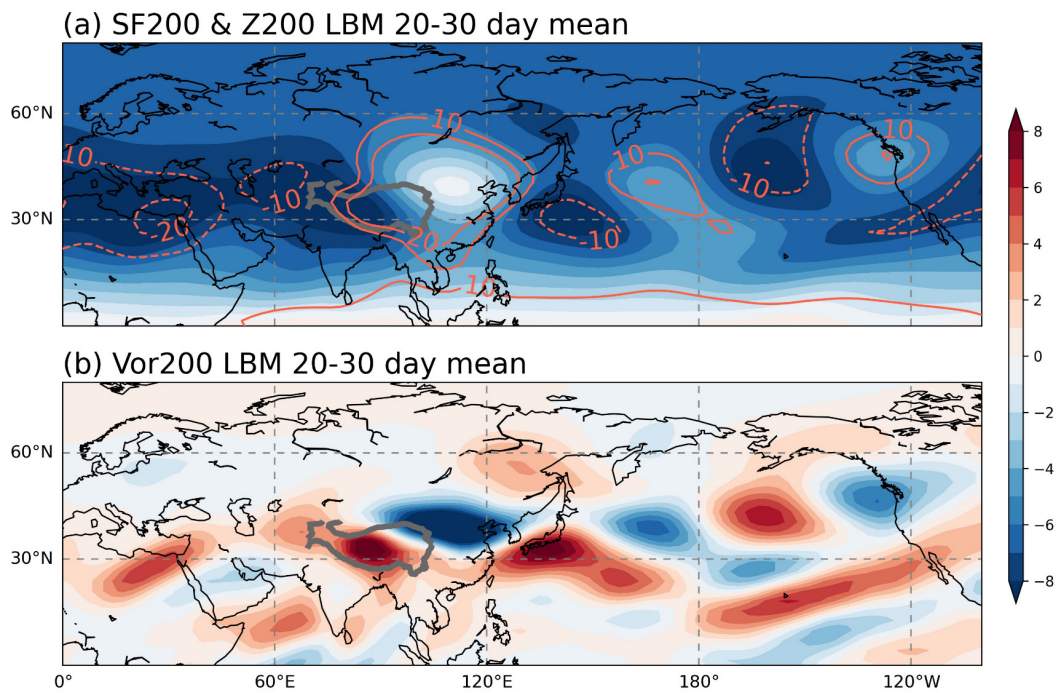


Figure 3-15 The LBM simulated response. The responses of 200-hPa circulation simulated by LBM. The outputs are averaged from 20 to 30 days when LBM reached equilibrium. (a) 200-hPa stream function (shading; units: $\times 10^6 \text{ m}^2/\text{s}$) and geopotential height (contour). (b) 200-hPa vorticity (units: $\times 10^{-6} \text{ s}^{-1}$).

The wave train emanating from TP thermal anomaly is examined by the LBM experiment and is found to be similar to the result derived from the reanalysis dataset. To further convince these results, a large-ensemble AGCM simulation (HIST) was used to examine how May SUBT anomaly could affect the subsequent June TRC teleconnection. The variables are concatenated along their time axis to produce a total 6100-year time series (i.e., 100 members with 61 years). An index of standardized detrended May T2m was created for HIST experiment by area-averaging the detrended T2m over TP region (elevation above 2500 m), and then, the associated variables were regressed on this index to study their association with TP temperature.

The regression results are presented in **Figure 3-16**, to illustrate the memory of the cold TP. Please note that the sign was reversed for better comparison with the SVD result. The AGCM large ensemble also shows that the central-western TP is the region with strong land temperature memories, evidenced by persistent SUBT (figure is omitted). The regression analysis for 200-hPa vorticity replicates the TRC-like wave train whose geographical locations of anomalies are almost in phase with our composite result using reanalysis (**Figure 3-13b**) and the LBM experiment (**Figure 3-15**). Therefore, it is very likely that, in AGCM ensemble experiment, land surface temperature anomalies in May are coupled with a TRC-like wave train. These results from multiple different approaches provide confidence that the circum-global teleconnection in the observed result is not a coincidence. However, the regression analysis for June precipitation in HIST shows a chaotic nature with a Pacific-Japan like feature, which is different from our composite result in **Figure 3-5**. Further investigations are needed in future to relate TRC-wave train and other factors modulating precipitation variability.

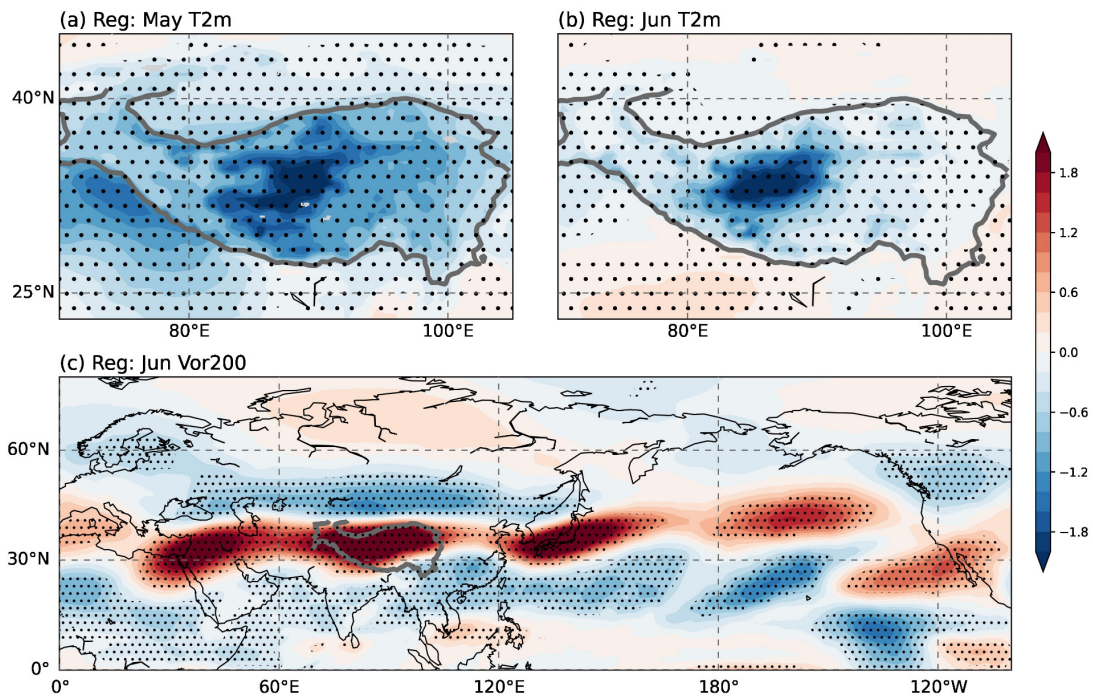


Figure 3-16 The HIST simulated response. The regressed patterns upon the sign-reversed TP May T2m index in HIST. (a) May T2m (units: K), (b) June T2m (units: K), and (c) June 200-hPa vorticity (units: $\times 10^{-6} \text{ s}^{-1}$). Dotted regions denote anomalies significant at the 99% confidence level, based on the two-sided Student's *t* test.

3.7. Summary and Conclusions

In this study, we reported a feature and possible mechanism of May TP soil memories affecting subsequent June TRC teleconnection. The TP soil is able to memorize the cold anomalies after it is created by the forcing from the atmosphere, and the cold anomalies in soil can persist until the next month. Through the turbulent heat flux as one of the interaction processes between land and atmosphere, the stored ground temperature anomaly might provide preferential conditions for the air to cool again. Both reanalysis and large-ensemble simulations show these processes are more evident in the central-western TP, accompanying with clear vorticity anomaly in the upper troposphere. The role of this vorticity anomaly was investigated by LBM experiment in which vorticity forcing was prescribed in the central-western TP, and it is found that the vorticity anomaly can trigger a TRC teleconnection under June basic state. The June TRC teleconnection exhibits enhanced anomalies in cyclonic circulation over the western TP, southern Japan, and near the US continent of the Rocky Mountains, and the anti-cyclonic circulation (i.e., divergence) over Yangtze River basin which might affect precipitation there.

Although the June TRC teleconnections in reanalysis and HIST are similar, the precipitation response looks different. Meanwhile, how the cold land surface coupled to the vorticity anomalies is yet unclear, which needs the verification by AGCM experiment with carefully tuning. The reasons and mechanisms must be understood in the future studies. The issue to be discussed is the relevance of TRC teleconnection to the year-to-year variability of the jet stream, because the jet stream controls the wave propagation from western TP to East Asia. The future effort in this line could help figure out the chaotic nature of precipitation response to TP forcings.

4

Discussions and outlook

4. Discussions and outlook

In order to understand whether the consideration of ISM subseasonal variations is important to improve the prediction of EASM, especially compared to the monthly mean ISM precipitation and CGTI from Q. Ding & Wang, (2005), the forecast skill of June EASM wet band is evaluated. The presence or absence of the June EASM wet band is determined by the June EASM wet band precipitation anomalies above or below 0 mm/day, respectively. The presence of monthly CGT, abnormal ISM monthly precipitation anomalies, and lag0 day are determined by $\text{CGTI-PC1} > 1.0$, standardized ISM precipitation anomalies > 1.0 , and successfully identified lag0 day, respectively. After obtaining those values, the hit rate, false alarm, and threat score are computed (**Table 4-1** and **Table 4-2**). The hit rate is the proportion of hits to the total of hits and misses. The false alarm rate is the ratio of false alarms to the total of hits and false alarms. The threat score is the ratio of hits to the sum of hits, misses and false alarms.

Table 4-1 Observed forecast skill in June of EASM wet band for lag0, CGTI-PC1, and abnormal ISM monthly precipitation between 1958 and 2015.

	Lag0	CGTI-PC1	ISM precip.
Hit (count)	15	11	7
Miss (count)	12	16	20
False alarm (count)	12	8	4
Hit rate (%)	55.56	40.74	25.93
False alarm ratio (%)	44.44	42.11	36.36
Threat score (%)	38.46	31.43	22.58

Table 4-2 HIST forecast skill in June of EASM wet band for lag0, CGTI-PC1, and abnormal ISM monthly precipitation (totally 6100 years, from 1951 to 2011 with 100 ensemble members).

	Lag0	CGTI-PC1	ISM precip.
Hit (count)	1683	908	725
Miss (count)	1213	1988	2171
False alarm (count)	1074	782	251
Hit rate (%)	58.11	31.35	25.03
False alarm ratio (%)	38.96	46.27	25.72
Threat score (%)	42.39	24.69	23.04

The forecasting skill of lag0, namely, taking into account the sub-seasonal variations of the ISM, is the best of these three factors. It has the highest hit rate and

threat score due to the highest succeeding forecast (i.e., hit), although the false alarm ratio is not outstanding especially in the observation. The CGTI-PC1 has the second-best prediction, while using the ISM mean precipitation is the worst. The results support the speculation in [section 2.4.5](#), that is, the signal of the contribution of subseasonal ISM heating in the interannual EASM is enhanced or obscured when several processes coexist, which does not necessarily determine the physical link between ISM and EASM.

The impact of CGT on mid-latitude weather and climate, as discussed in [chapter 2](#), has become a prominent topic in recent years. It exerts influence not only on weather and climate extremes like heavy precipitation and heatwaves in East Asia but also plays a crucial role in shaping the hydrological cycle in the mid-latitude region, such as the mass balance of the glaciers in TP (see [section 1.2.3](#) for detail). Given its status as first leading summer circulation mode in Eurasia and its substantial impact, the predictability of the CGT raises significant concerns. Sugimoto et al. (2022) discovered that a crucial aspect for accurately simulating the interannual variations of June EASM precipitation lies in accurate prediction of the CGT, using regional climate model simulations. The further investigation shows the bias in the regional climate model might be greatly influenced by the cumulus (or convective) parameterization schemes, namely, the performance of Kain–Fritsch scheme (Kain & Kain, 2004) is better than Grell-3D ensemble (Grell & Dévényi, 2002), with other settings nearly identical (Sugimoto et al., 2022). This simulated result indicates that the better representative of ISM precipitation in June is the key to improve the East Asian weather and climate prediction, similar to our claim in [section 2.4.5](#). Therefore, the mechanism by which some cumulus parameterizations work well to represent ISM precipitation while others do not is another interesting question to illustrate. Apart from this, a recent study also indicates that the ISM heating in different sub-regions could trigger different circulation patterns (Gupta et al., 2023). Specifically, the circulation patterns associated with the heating in the northern and southern Indian subcontinents differ significantly. This would also be an interesting topic to see how this difference could perturb the ISM–EASM connection. Finally, comprehending the initiating trigger of the European-CGT is another crucial aspect that can enhance the forecasting of both the ISM and EASM, which is less investigated and understood at this moment.

The connection between TP soil enthalpy and EASM, as preliminary presented in [chapter 3](#), is also a hotspot in recent years. This topic is under the Global Energy and Water Exchanges (GEWEX) of the program named “Impact of Initialized Land Temperature and Snowpack on Subseasonal-to-Seasonal Prediction” (LS4P; Xue et al., 2022). Consistent results show a realistic initialization of the land component in land–atmosphere coupled AGCM leads to the better prediction of temperature and precipitation for the 1–3 months (Dirmeyer & Halder, 2017; Shukla et al., 2019). However, the models are unable to represent land conditions over TP with significant bias, due to its complex terrain and land–atmosphere interaction, as well as rare availability of TP in situ measurement (Flanner, 2005; Xue et al., 2022). This significantly complicates the assessment of model performance and the identification of biases for enhancing soil initialization. The result in [chapter 3](#) provided the substantial observation basis and preliminary modelled result, which is helpful for the understanding on memory of the TP soil enthalpy. In subsequent investigations, an intensified emphasis should center on recognizing model deficiencies, specifically those related to thermal forcing. Multiple studies suggest that these inadequacies are primarily rooted in the representation of land surface schemes governing frozen-

thawing processes and the radiation budget (Lee et al., 2019; K. Li et al., 2022).

Finally, the interaction between CGT and TP thermal forcing is an intriguing subject. TP thermal forcing plays a significant role in initiating the ISM onset by modulating the land-sea thermal gradient (Senan et al., 2016). Understanding the interaction of these two thermal forcings at a subseasonal timescale and their respective contributions to the EASM are desired issues, which have currently received less attention.

5

Conclusions

5. Conclusions

The EASM is identified as structure with northeastward elongated rainband which result in spatiotemporally continuous rainfall, affecting East China, the Korean Peninsula, and Japan. The EASM exhibit great fluctuations at both interannual and subseasonal time scales. The EASM and have the potential to induce significant consequences such as severe flooding or drought to East Asian countries. These events can significantly impact local food production, water security, and the associated socioeconomic impact. Timely and precise predictions of the EASM offer a viable means of mitigation to such extreme events. Over the past few decades, extensive efforts have been made to understand the fundamental physics and dynamical prediction of monsoon systems, with some significant progress, but still unsatisfactory. For example, the 2020 abnormal EASM was unpredictable even with advanced operational forecast model. Therefore, delving deeper into our understanding of EASM beyond current knowledge is necessary, which may provide information to the initialization strategy for operational forecast model. Based on these considerations, this dissertation provides new insights on interaction between EASM and mid-latitude forcings.

The ISM is a well-known Rossby wave source in summer. However, the debate persists regarding its interaction with the CGT and EASM at the interannual timescale. This is primarily due to the non-stationary correlation of precipitation within the ISM and EASM domains. Therefore, in [chapter 2](#), we delve into the intricate connection between the Indian ISM and EASM during June, at a shorter time scale (i.e., subseasonal time scale). To ensure the robustness of the result, this investigation relies on a comprehensive analysis drawing from observations, large-ensemble simulation output (HIST), and simplified numerical simulations. The findings indicate that the continuous propagation of the CGT wave train from western Europe toward the ISM could be associated with the onset of the ISM. Thereafter, the substantial condensation heating released from ISM precipitation perturbs the air surrounding the westerly jet stream, triggering a Rossby wave trapped within the westerly jet stream. This local Rossby wave propagating to the East Asian sector is found orthogonal to the CGT. Few days later, as this local wave train extends along the westerly jet stream, and it gradually reverts to a CGT pattern. Throughout this process, the CGT index (CGTI-PC1) displays a bimodal variation, exhibiting biweekly periodicity according to spectral analysis. The wave train triggered by the ISM results in poleward meridional wind prevailing in East Asia, bringing warm advection and abundant moisture transports to the EASM, contributing to the formation of the wet band. Further analysis suggests that ISM heating may act as a trigger or reinforcer of the EASM rain band, contingent on whether the meridional wind existed prior to the strong ISM heating. Moreover, the subseasonal ISM-EASM connection appears to be less influenced by ENSO and ISO. Exploration of this connection into other summer months reveals challenges in directly applying the results from June. The discrepancy may stem from the distinct seasonal jet locations, as depicted in responses shown by LBM. Namely, the ISM heating might not be able to trigger CGT wave train in July and August. The results in [chapter 2](#) strongly suggest that precise prediction of the ISM is of significant importance in prediction of the EASM, since these two monsoon systems are bridged by the CGT.

In [chapter 3](#), the evidence of potential connection between TP soil enthalpy and EASM are presented. This has become a hot topic in recent years, as it has been established that accurate representation of land parameters is crucial in forecast models.

The land conditions over the TP, however, exhibit significant biases. Hence, we are motivated to investigate the impact of TP land surface processes. The observational results show that the TP soil enthalpy has the capacity to retain the cold temperature forcing from the atmosphere. This capability potentially contributes to reinforcing cold atmospheric conditions in the subsequent month, through releasing the latent heating. This is supported by in situ data from the meteorological stations. The soil memory effect is notably pronounced in the central western TP, potentially instigating a wave train that propagates along the westerly jet stream, consequently influencing EASM precipitation. The large-ensemble simulation also supports the notion that the central western TP might exhibit a more substantial soil memory compared to other parts of the TP. The results confirm the potential delayed impact of TP soil memory, providing value-added insights for initializing strategies of the seasonal forecasting. Those result are summarized as **Figure 5-1**

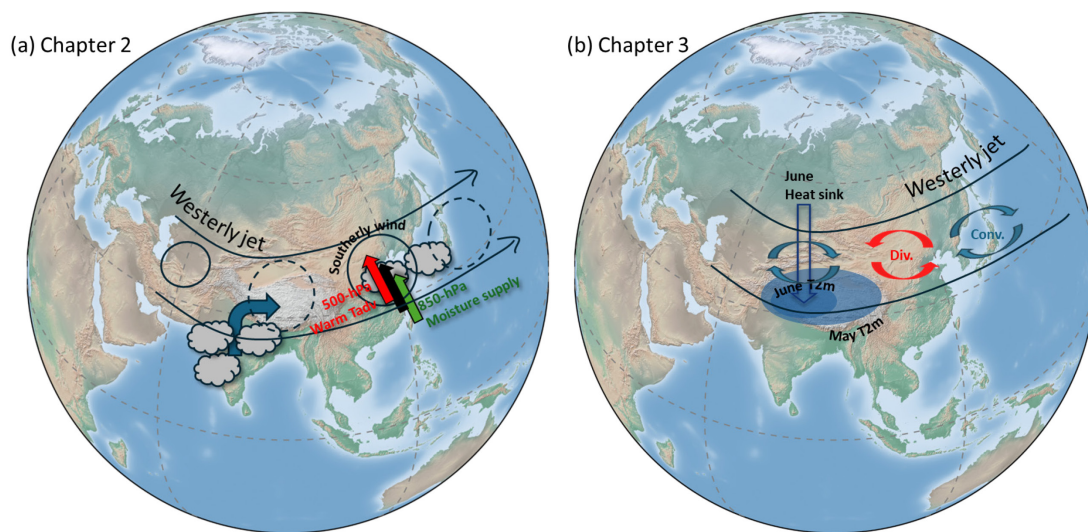


Figure 5-1 Schematic summary for Chapters 2 and 3. The Schematic figure showing the mechanisms of (a) subseasonal ISM–EASM connection bridged by CGT wave train, and (b) heat sink over central-western TP and associated retained cold soil memory, and its coincidence with the large scale upper-level divergence and convergence.

The impacts of two mid-latitude forcings, the CGT and TP land surface processes, on EASM activity are examined. This dissertation provides hints and evidence for better understanding of EASM. Based on the results, it is proposed that improved representations of ISM precipitation and TP soil processes are crucial for enhancing predictability of EASM. The improved prediction of EASM contributes to more effective mitigation strategies for the socio-economic challenges faced by East Asian countries. Future research endeavors should explore interactions with processes originating from the lower atmosphere (e.g., IOD, PJ-pattern, etc.), their interconnections, and optimal initialization strategy methods for their representation in numerical simulations.

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