Land–sea thermal contrast associated with summer monsoon onset over the Chao Phraya River basin

Tomohito J. Yamada1, Sourabh Shrivastava2, Ryosuke Kato2,3

Received: 4 January 2022 / Accepted: 26 June 2022 / Published online: 21 July 2022
© The Author(s) 2022

Abstract
Earlier onset of the Southeast Asian summer monsoon (SAM) was observed over the Chao Phraya River basin in Thailand using Thai Meteorological Department-derived high-resolution merged rainfall data from 1981 to 2016. SAM variability depends on numerous local and global factors, including thermal conditions over the Bay of Bengal (BoB) and Tibetan Plateau (TbT). Despite tremendous past research efforts, the effect of thermal heat contrast on the SAM remains unclear. Using observational and reanalysis datasets, we found that the absolute value of total heat over the BoB was increasing. However, the interannual variability of total heat was greater over the TbT. Changes in surface temperature (±1.5°C), air thickness (±20 m), and geopotential height over the TbT were associated with the timing of SAM onset. The results also suggested that significant changes in air thickness are driven by surface temperature differences over the TbT, while changes in the integrated apparent heat source and integrated apparent moisture sink of ±100 W m⁻² resulted in anomalous convective activities over the BoB and mainland of the Indochina Peninsula in years of early and late SAM onset. At the intraseasonal timescale, Madden–Julian oscillation (MJO) was observed over the Indian Ocean and Western Hemisphere for 4–10 days in years of early SAM onset. The opposite situation was found for years of late SAM onset, with MJO located over the Western Pacific Ocean and Maritime Continent.

1 Introduction

Thailand receives approximately 80% or more of its annual total rainfall between May and October according to a Thai Meteorological Department (TMD, 2021) during the South Asian summer monsoon (SAM). However, the variability of SAM onset has been raised as a topic of concern (Zhou et al. 2019) due to its effects on agriculture, human population, and industry. For example, rice production strongly depends on the parameters of SAM as well as its onset. In addition, the SAM affects more than 69.6 million people in Thailand (The World Bank, 2019), particularly in the Chao Phraya River basin (CPB), which covers approximately 30% (157,925 km²) of the mainland area of Thailand. Moreover, SAM onset has major effects on the gross domestic product of Thailand.

The characteristics of SAM onset have been extensively studied (Zhang et al. 2002; Takahashi and Yasunari, 2006; Ding, 2007; Kuraji and Arthorn, 2011; Kajikawa et al. 2012; Nguyen-Le, 2015; Shrivastava et al., 2017; Tanaka et al. 2019). Using a rainfall index for 1951–1996, Zhang et al. (2002) defined the mean onset date over the central Indochina Peninsula (ICP) as approximately the 26th pentad of the year. Kajikawa et al. (2012) found that the SAM has arrived earlier over the ICP in the three decades from 1979 to 2008 based on Climate Prediction Center merged analysis of precipitation (CMAP) at 2.5°×2.5° spatial resolution. Minoura et al. (2003) discussed the mechanism of SAM onset and characterized the ICP region as having relatively slow onset based on European Centre for Medium-Range Weather Forecasts reanalysis data with a spatial resolution of 2.5°×2.5° from 1982 to 1993. Using ground observations, a trend of earlier SAM onset was also identified by Kuraji and Arthorn (2011) and Tanaka et al. (2019) in the Mae Chaem watershed of the ICP. In recent decades, long-term high-resolution gauge rainfall data for the ICP have become available for research purposes. Therefore, our first question...
is whether SAM onset has continued to shift earlier in recent decades or whether this trend has changed.

Extensive efforts have been made to determine the relationships between SAM onset and various contributing factors (e.g., Minoura et al. 2003; Kiguchi et al. 2016). Takahashi and Yasunari (2006) determined climatological mean values of rainfall over 50 years and showed a distinct climatological monsoon break occurring at approximately the 33rd pentad over the ICP, with a dramatic change in large-scale monsoon circulation in the seasonal progression. Kiguchi et al. (2016) analyzed premonsoon rainfall to identify its relationship with SAM onset, but found no clear relationship. Minoura et al. (2003) discussed the mechanism of SAM onset, determined that onset of the Indian summer monsoon (ISM) was slower than SAM onset, and suggested that the onset dates of ISM and SAM should be distinguished. The question of whether global teleconnections affect SAM onset over the ICP remains unanswered. Takahashi and Yasunari (2006) identified a trough over the Bay of Bengal (BoB) during early SAM onset; at the same time, weak southwest-erlies appeared over the eastern Indian Ocean and high precipitation was subsequently recorded over the ICP. Many efforts have focused on SAM onset measured based on the El Niño–Southern Oscillation (ENSO) and Indian Ocean dipole, and the relationship between SAM onset timing and large-scale variability modes has been documented (Wu and Wang, 2000; Wang et al. 2001; Ding, 2007). Wu and Wang (2000) identified a link between ENSO and SAM onset. Their research indicated that physical processes occurring on land and in the ocean are critically responsible for reducing land–ocean thermal contrast. The Madden–Julian oscillation (MJO) may also affect SAM onset at the intraseasonal scale (Chi et al. 2015). The MJO is one of the dominant modes of intraseasonal variability in the tropics (Madden and Julian, 1972). The eastward-moving MJO is associated with deep convection and circulation from eastern Africa, the Indian Ocean, and the Maritime Continent to the Western Pacific (Madden and Julian, 1971, 1972). When the MJO moves to the eastern and western portions of the Indian Ocean, it impacts SAM onset (Chi et al. 2015). The MJO, which is influenced by thermal conditions over the Indian Ocean, is not associated with either late or early SAM. The relationships of SAM onset with local factors also remain unclear. Therefore, we explored whether the intensity and position of the MJO can influence SAM onset.

Heat contrast between the Asian landmass and the tropical Indian Ocean is essential to changes in SAM onset (Kajikawa et al. 2012). The relationship between sea surface temperature (SST) over the BoB and rainfall over the ICP has not been fully clarified (Kanae et al. 2002). Therefore, some studies have focused on the effect of land–sea thermal contrast on the SAM (Wu and Wang, 2000; Nguyen-Le et al., 2015). For establishment of the SAM, the Tibetan Plateau (TbT) is a high-elevation heat source (Yanai et al. 1992; Abe et al. 2013) that strongly impacts the transition of SAM rainfall (Wang et al. 2001; Sato and Kimura, 2007). Yanai et al. (1992) revealed that increased temperature over the eastern TbT during SAM onset was mainly driven by diabatic heating. Minoura et al. (2003) provided strong evidence of a relationship between thermal contrast and SAM onset over the ICP. Zhang et al. (2017) found a significant thermal difference based on the atmospheric heat source and suggested that SAM onset lags behind the time of thermal change. Moreover, the importance of sensible heat flux in the western TbT and latent heat release over eastern India and the adjoining region was demonstrated by Flohn (1957). After analyzing observational data, Xu and Chan (2001) found that both sensible and latent heat fluxes determine thermal conditions over the TbT. Numerous studies have investigated the apparent heat source (Q1) and apparent moisture sink (Q2), as well as their links to large-scale atmospheric circulation (Yanai et al. 1973; Schumacher et al. 2008; Xing et al. 2016). Despite these advances, many important and basic questions remain unanswered, such as how thermal conditions over the western TbT influence and interact with SAM onset and how Q1 and Q2 behave during SAM onset over the BoB and ICP. Therefore, an effort has been made to explain the possible mechanism driving SAM onset timing over the CPB. SAM onset over the CPB begins in May. Here, we focus on the first three pentads of May (details provided in the following section).

The remainder of this paper is organized as follows. In “Sect. 2,” we describe the data and methods used in this study. In “Sect. 3,” the results are presented and discussed. The conclusions of the study are provided in “Sect. 4.”

2 Data and methods

2.1 Observational and reanalysis data

We primarily used TMD-derived high-resolution merged rainfall data with 0.05°×0.05° spatial resolution to investigate the climatology and rainfall fraction corresponding to early and late onset, and plotted pentad rainfall over the ICP between 1981 and 2016. This dataset is obtained with a network of daily rain gauges operated by the TMD. To investigate temperature variability over East Asia, we used the high-quality gauge-based temperature dataset APHRODITE V1808 (Yasutomi et al. 2011), specifically the monsoon Asia version of APHRODITE. We also used the Optimum Interpolation SST (OISST) data product of the National Centers for Environmental Information (NCEI) and National Oceanic and Atmospheric Administration (NOAA).
to 2016 (Reynolds et al. 2007). Three-dimensional atmospheric data from The Modern Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2), were used for the period of 1981–2016. Three-hourly averaged data for air temperature, wind circulation, omega (vertical pressure velocity), relative humidity, and geopotential height were downloaded (GMAO, 2021). MERRA-2 was developed using the Goddard Earth Observing System (GEOS-5.12.4) atmospheric data assimilation system with a spatial resolution of 0.5°×0.625° (Gelaro et al. 2017). We used MERRA-2 because MERRA computes the horizontal discretization of the model on a latitude–longitude grid, whereas MERRA-2 uses a cubed-sphere grid; moreover, the variable used to control moisture in MERRA-2 aligns with recent versions of the Gridpoint Statistical Interpolation (GSI) and differs from that used in MERRA. Compared to the MERRA system, the background error statistics for MERRA-2 show smaller standard deviations for many variables. Meanwhile, both larger and smaller correlations were obtained for MERRA-2, depending on the variable, latitude, and vertical level. Moreover, MERRA included no new satellite observation sources after the introduction of NOAA-18 in 2005. MERRA-2 includes numerous additional satellite observations, both before and after that mission (Gelaro et al. 2017). Two real-time multivariate MJO indices (RMMs; RMM1 and RMM2) introduced by Wheeler and Hendon (2004) were used to define the various phases of MJO (RMM, 2021).

2.2 Methodology

2.2.1 Definitions of South Asian monsoon onset date, early and late years

Pentad (5-day simple mean) data of rainfall and other atmospheric variables were constructed for the 36-year period. The methodology and definition used for estimation of SAM onset were adopted from Wang and Ho (2002) over 98°E–103°E and 13°N–20°N (CPB), where the definition of the SAM onset pentad is based on a relative rainfall rate exceeding 5 mm day−1 after deduction of the January mean occurring during the boreal summer (May–September). Zhang et al. (2002) found that most early and late years correspond to the definition of Wang and Ho (2002). To avoid sampling errors, in our subsequent analysis, equal sample sizes of the nine earliest (1986, 1988, 1994, 1995, 1999, 2000, 2001, 2002, and 2007) and nine latest (1983, 1987, 1991, 1992, 1993, 1997, 2010, 2013, and 2015) onset years were used.

Here, we selected rainfall events during the first three pentads in May, as the CPB receives approximately 50% or more of its rainfall events in the first three pentads of May (Fig. 1a). In early/late onset years, the fraction of total rainfall during this period was 60%/40% (Fig. 1b–1c). We also identified the change in monsoon onset (Fig. 1d, discussed in “Sect. 3”). Figure 1e shows daily mean rainfall and the difference in rainfall amount. This difference occurs before the first three pentads of May; after the third pentad, the difference becomes negligible in all three categories (average, early, and late). Therefore, based on our analysis, we found that the first three pentads in May are a reasonable period for differentiating early and late onset years.

2.2.2 Apparent heat source (Q1) and apparent moisture sink (Q2)

We calculated atmospheric Q1 and Q2 based on thermodynamic equations. The equations used are as follows (Yanai et al. 1973):

\[
Q_1 = c_p \frac{\partial T}{\partial t} - c_p (\omega \sigma - \mathbf{v} \cdot \nabla T)
\]  

(1)

\[
Q_2 = - \left( L \frac{\partial q}{\partial t} + L \mathbf{v} \cdot \nabla q + L \mathbf{w} \frac{\partial q}{\partial p} \right)
\]  

(2)

Here, the variables are air temperature (T), time (t), horizontal wind vector (V), vertical wind velocity (w), specific heat at constant pressure \( (C_p = 1004.64 \text{ J (K kg)}^{-1}) \), latent heat of condensation \( (L) \), specific humidity \( (q) \), static stability \( (\sigma) \), pressure \( (p) \), and the horizontal gradient operator \( (\nabla) \).

The vertically integrated forms of Q1 and Q2 can be written as follows:

\[
< Q_1 > \approx LP + S + < Q_R >,
\]  

(3)

\[
< Q_2 > \approx L(P - E).
\]  

(4)

Here, \( Q_R \) is the radiative heating rate and \( P, S, \) and \( E \) are the precipitation rate, sensible heat flux, and evaporation rate per unit area at the surface, respectively (Yanai et al. 1973, 1992).

\[
< >= \frac{1}{g} \int_{p_b}^{p_t} \left( \mathbf{v} \right) dp.
\]  

(5)

Here, \( p_t \) (300 hPa) is pressure at the top level and \( p_b \) (1000 hPa) is pressure at the surface.

2.2.3 Madden–Julian oscillation

The MJO is a 30–60-day oscillation, known as an intra-seasonal oscillation. The MJO can modulate the timing and strength of the SAM (Jones and Carvalho, 2002), as described by Wheeler and Hendon (2004). We selected the approximate location of the MJO (Donald et al. 2004) shown in Fig. 2 and noted that the MJO travels east at 4–10 days...
per phase. Therefore, we used a 10-day period to assess the intensity and location of the MJO during SAM onset.

3 Results and discussion

3.1 Fluctuations and climatological features of South Asian Monsoon onset

To address the first research question, we plotted the difference in SAM onset pentad (5-day simple mean) estimated using the method of Wang and Ho (2002) between the periods of 1999–2016 and 1981–1998 over each grid in northern Thailand (Fig. 1a–c). Our results suggested that SAM onset in the CPB trended earlier (by approximately two to four pentads) over the last 2 decades (Fig. 1d).

We aimed to elucidate the climatological features of surface and atmospheric variables during SAM onset. We used average wind vectors (850 hPa), specific humidity (850 hPa), temperature (land and ocean), and geopotential height (500 hPa) for the 36-year period from 1981 to 2016. At 850 hPa, the prominent pattern of wind systems was cross-equatorial flow driven by southwesterlies in the first three pentads of May, and these southwesterlies crossed the ICP and moved toward the South China Sea (Figure S1a). Northwesterlies arising from northwestern India and the region adjoining the ICP were also dominant during the onset period, especially over the northern BoB. Specific humidity (g kg⁻¹) was lower over the BoB (< 10 g kg⁻¹) than the ICP (> 10 g kg⁻¹) (Figure S1b). Meanwhile, a warmer ICP was observed, with surface temperatures above 28 °C, while the SST of the BoB was above 25 °C during the first...
three pentads of May (Figure S1c). Moreover, the TbT, with an average temperature of <10 °C, showed warming in the regions of northwestern India, Pakistan, and Afghanistan. The 500-hPa geopotential heights, which exceeded 5850 gpm over the BoB and 5750 gpm over the TbT and in the trough over the ICP, are shown in Figure S1d.

3.2 Climatology and associated signals of early and late South Asian monsoon onset

To identify the signals associated with the change in SAM onset over the CPB, air thickness anomalies were plotted for the first three pentads of May because atmospheric thickness between 200 and 500 hPa in the monsoon region is an essential parameter (Li and Yanai, 1996; Kajikawa et al. 2012). Figure 3a shows the climatology of air thickness, which was calculated as the height difference between 200 and 500 hPa. In this plot, air thickness exceeded 6800 m over the BoB, much greater than that over the TbT, suggesting that air thickness increased over the BoB and associated regions, including the ICP, during the SAM onset period. The anomalous pattern of air thickness is also identified here in Fig. 3b and 3c, and highly significant (95%) changes in air thickness were present over the western TbT, northwestern India, Pakistan, the BoB, and the ICP (Fig. 3b). No specific threshold value for atmospheric thickness has been defined in previous research, but we found that a difference in air thickness of 30 m or greater over the TbT relative to the BoB led to early SAM onset. Meanwhile, in Fig. 3c, decreased air thickness (up to 30 m) over central India, TbT, and the BoB identified during late onset. This pattern also indicates that local factors in the BoB influence SAM onset.

Specifically, the longitudinal profile of heat contrast between land and ocean was plotted for each pentad of the year (shown in Fig. 4a for pentads 27–37). Figure 4a also shows the climatological averages for early and late SAM onset years, defined as the average pentad when the thermal contrast between 27.5°N–37.5°N (land) and 5°N–15°N (ocean) reaches zero. We found that the large shift in thermal heat contrast at 98°E–100°E over the ICP occurred earlier by three pentads, with a standard deviation of ±4 pentad, in early SAM onset years. In years with late SAM onset, the delay of this shift was approximately one pentad relative to the climatological average, with a standard deviation of ±2 pentad. These results indicate that land (TbT) is the dominant source area of factors causing early SAM onset, followed by the BoB. Figures 4b and 4c plot the difference in thermal contrast in early and late years relative to the climatological average. In early onset years, significant anomalies of 90% greater thickness were found in the upper
troposphere along the western TbT between the 25th and 28th pentads (May), which were associated with positive (negative) changes of more than ±1.5–2.0 °C in surface temperature anomalies (Yanai et al. 1992). At this time, the ICP receives its first period of rainfall (Zhang et al. 2002), but an air thickness increase (decrease) over 60–80°E (western TbT) of ±20 m can affect SAM onset. In contrast, during the full monsoon period (May–October), large changes in air thickness were not observed.

Here, we analyzed the anomalous pattern of various atmospheric variables such as 850 hPa wind (Fig. 5a–5b), specific humidity (g/kg) (Fig. 5c–5d), surface temperature (°C) (Fig. 5e–5f), and 500 hPa geopotential height (m) (Fig. 5g–5h). At a geopotential height of 500 hPa (Fig. 5g), positive anomalies were found over the western TbT, northwestern India, Pakistan, and Afghanistan in early years, while negative anomalies were found over the BoB. The increase in 500-hPa geopotential height over the TbT during early onset years was related to warming of the upper troposphere along the western TbT between the 25th and 28th pentads (May), which were associated with positive (negative) changes of more than ±1.5–2.0 °C in surface temperature anomalies (Yanai et al. 1992). At this time, the ICP receives its first period of rainfall (Zhang et al. 2002), but an air thickness increase (decrease) over 60–80°E (western TbT) of ±20 m can affect SAM onset. In contrast, during the full monsoon period (May–October), large changes in air thickness were not observed.

Here, we analyzed the anomalous pattern of various atmospheric variables such as 850 hPa wind (Fig. 5a–5b), specific humidity (g/kg) (Fig. 5c–5d), surface temperature (°C) (Fig. 5e–5f), and 500 hPa geopotential height (m) (Fig. 5g–5h). At a geopotential height of 500 hPa (Fig. 5g), positive anomalies were found over the western TbT, northwestern India, Pakistan, and Afghanistan in early years, while negative anomalies were found over the BoB. The increase in 500-hPa geopotential height over the TbT during early onset years was related to warming of the upper troposphere along the western TbT between the 25th and 28th pentads (May), which were associated with positive (negative) changes of more than ±1.5–2.0 °C in surface temperature anomalies (Yanai et al. 1992). At this time, the ICP receives its first period of rainfall (Zhang et al. 2002), but an air thickness increase (decrease) over 60–80°E (western TbT) of ±20 m can affect SAM onset. In contrast, during the full monsoon period (May–October), large changes in air thickness were not observed.

Here, we analyzed the anomalous pattern of various atmospheric variables such as 850 hPa wind (Fig. 5a–5b), specific humidity (g/kg) (Fig. 5c–5d), surface temperature (°C) (Fig. 5e–5f), and 500 hPa geopotential height (m) (Fig. 5g–5h). At a geopotential height of 500 hPa (Fig. 5g), positive anomalies were found over the western TbT, northwestern India, Pakistan, and Afghanistan in early years, while negative anomalies were found over the BoB. The increase in 500-hPa geopotential height over the TbT during early onset years was related to warming of the upper troposphere along the western TbT between the 25th and 28th pentads (May), which were associated with positive (negative) changes of more than ±1.5–2.0 °C in surface temperature anomalies (Yanai et al. 1992). At this time, the ICP receives its first period of rainfall (Zhang et al. 2002), but an air thickness increase (decrease) over 60–80°E (western TbT) of ±20 m can affect SAM onset. In contrast, during the full monsoon period (May–October), large changes in air thickness were not observed.
troposphere, possibly explaining the strong northwesterlies toward the ICP observed in early onset years (Fig. 5g). The opposite pattern was identified in late onset years, with the western and central TbT having negative anomalies while the Arabian Sea, BoB, and South China Sea showed positive anomalies of 500-hPa geopotential height associated with delayed SAM onset over the CPB (Fig. 5h). From these findings, the thermal conditions of the BoB and TbT show
opposite trends but correlated each other. Therefore, clarifying the roles of other atmospheric and surface patterns is essential.

In Fig. 5e and 5f, cold (warm) surface temperatures and SST anomalies are shown over the ICP and BoB in early (late) onset years, as discussed previously (Li and Yanai, 1996; Xing et al. 2016). The dipole pattern (temperature anomalies over the TbT and BoB) indicates favorable climatic conditions and high pressure over northwestern India, Pakistan, and the western TbT in early onset years; a low-pressure system apparently formed over the BoB with intense westerlies, leading to early onset. Cyclonic/anticyclonic circulation in the lower troposphere is present in early/late onset years north of the BoB (Xing et al., 2016). Moreover, the strong energetic southwesterly monsoon approached the ICP and transported moisture from the ocean toward land (Fig. 5a), which caused an increase in specific humidity over the ICP in early onset years (Fig. 5c). In contrast, weaker winds (Fig. 5b) and lower specific humidity (Fig. 5d) were found in years of late SAM onset.

After analyzing thermal contrast and other atmospheric variables, $<Q1>$ and $<Q2>$ W m$^{-2}$ were averaged for the period of 1981–2016, and the results are shown in Fig. 6. In addition, Fig. 7 summarizes MJO position and intensity in years of early (Fig. 7a) and late (Fig. 7b) SAM onset over the CPB. Figure 6a and 6b shows, climatologically, the TbT and mainland ICP became more important heat sources during SAM onset compared to the ocean (Taylor et al. 2003). Figures 6c–6f show anomalies of $<Q1>$ and $<Q2>$, and anomalous increases in both $<Q1>$ and $<Q2>$ over mainland ICP exceeding 80 W m$^{-2}$ were apparent in early SAM onset years (Fig. 6c and 6d). Strong convection driven by southwesterly wind was pushed by a strong MJO (Fig. 7a) toward the Indian Ocean and Maritime Continent. The
dominant position of the MJO during early SAM onset was in phases 1, 2, 3, and 4 (75% of total early onset events). However, in late SAM onset years, significant changes in $<Q_1>$ and $<Q_2>$ occurred over the BoB (Fig. 6e and 6f). Our results also reflect anomalous activities caused by late SAM onset; one such effect was decreased latent heat that led to a reduction in specific humidity and weakening of low-level winds, all of which were driven by the weak MJO and its position (Fig. 7b) during late SAM onset. Here, only two events (15%) showed the MJO over the Indian Ocean and Maritime Continent. Moreover, of those two events, one event was associated with a weak MJO. The mechanism underlying this finding is that when the MJO remains in phases 1, 2, 3, and 4, convective activities over the ICP increase (Fig. 1), whereas when it moves toward the east and crosses the Maritime Continent, convection over the ICP may be suppressed due to increased upward motion over the western Pacific.

4 Conclusion

In recent decades, earlier SAM onset has been observed over the CPB through analysis of TMD rainfall. We described the cause of the observed change in SAM onset over the CPB. The major findings of this study are as follows. First, in association with early/late onset, warmer/colder surface conditions ($\pm 1.5$ °C or greater changes) over the western TbT and adjacent regions were detected. Anomalous changes of $\pm 20$ m in air thickness over the TbT were noted during the first three pentads of May in early/late onset years, driven by rising/falling temperatures. In climatological terms, air thickness was greater over the BoB, but the changes in air thickness were greater over the TbT than the BoB. During SAM onset, TbT becomes a heat source; investigation of $<Q_1>$ and $<Q_2>$ showed positive (> 150 W m$^{-2}$) changes over the BoB, with stronger convective activities associated with intense MJO during early SAM onset. Strong MJO positioned over the Indian Ocean and Maritime Continent increased stormy conditions. Years with warmer conditions over TbT led to the initiation of stronger MJO. This strengthened MJO spent around 4–10 days (Fig. 7a) over the Indian Ocean, causing stormy conditions that led to early SAM onset. In contrast, negative (approximately $-100$ W m$^{-2}$) anomalies were found in late onset years, driven by weakened southwesterlies over the BoB. As the southwesterlies weaken, the transport of moisture over the ICP decreased sharply, while the ICP cooled and ceased to provide a heat source; these conditions do not favor SAM onset. In conclusion, thermal conditions over the TbT are dominant during early SAM onset, but several other factors such as weakened southwesterlies, low specific humidity over the BoB, and cooling of the ICP suppress SAM onset.

The significance of various atmospheric variables during SAM onset is very important. Therefore, this study is essential to accurate prediction of SAM onset based on thermal conditions over the TbT and BoB. In addition to changes in SST patterns, a warmer climate may cause changes in the amplitude of MJO, which will also affect SAM onset. Therefore, we recommend identifying MJO impacts during SAM onset under climate change conditions.

Supplementary Information The online version contains supplementary material available at https://doi.org/10.1007/s00704-022-04131-3.

Author contribution In this study, the research question, funding, and writing were done by Dr. TJY. Dr. SS contributed the data analysis and writing as well as research question, and Mr. RK helped for the data collection and analysis.

Funding This study was supported by the MEXT-Program for the advanced studies of climate change projection (SENTRAN) Grant Number JPMX07 + e-rad, JSPS KAKENHI (Grant Number 22H01594), JST/JICA (SATREPS), and ArCS.

Data availability The rainfall datasets analyzed during the current study are not publicly available. However, the data shall be requested...
to the Thai Meteorological Department (TMD). Other datasets are publicly available such as surface temperature is from APHRODITE V1808 (Yasunomi et al. 2011), OSI SST is from NOAA (Reynolds et al. 2007), MERRA-2 is from https://gmao.gsfc.nasa.gov/ (Gelaro et al. 2017), RMM is from Wheeler and Hendon (2004), and Fig. 2 is from https://www.climate.gov/ with credit to Fiona Marton.

Code availability All codes are developed using Python 3.6 except the <Q1> and <Q2> calculation. The code is downloadable from the NCL library https://www.ncl.ucar.edu/.

Declarations

Ethics approval This research is purely original and this research does not harm to the society or any individual knowingly and unknowingly.

Consent to participate Informed consent was obtained from all individual participants included in this study.

Conflict of interest The authors declare no competing interests.

Open Access This article is licensed under a Creative Commons Attribution 4.0 International License, which permits use, sharing, adaptation, distribution and reproduction in any medium or format, as long as you give appropriate credit to the original author(s) and the source, provide a link to the Creative Commons licence, and indicate if changes were made. The images or other third party material in this article are included in the article’s Creative Commons licence, unless indicated otherwise in a credit line to the material. If material is not included in the article’s Creative Commons licence and your intended use is not permitted by statutory regulation or exceeds the permitted use, you will need to obtain permission directly from the copyright holder. To view a copy of this licence, visit http://creativecommons.org/licenses/by/4.0/.

References

Abe M, Hori M, Yasunari T, Kitoh A (2013) Effects of the Tibetan Plateau on the onset of the summer monsoon in South Asia: the role of the air-sea interaction. J Geophys Res Atmos 118:1760–1776. https://doi.org/10.1002/jgrd.50210

Chi Y, Zhang F, Li W, He J, Guan Z (2015) Correlation between the onset of the East Asian subtropical summer monsoon and the eastward propagation of the Madden–Julian oscillation. J Atmos Sci 72:1200–1214. https://doi.org/10.1175/JAS-D-14-0038.1

Ding Y (2007) The variability of the Asian summer monsoon. Journal of the Meteorological Society of Japan. Ser II 85B:21–54. https://doi.org/10.2151/jmsj.85B.21

Donald A, Meinke H, Power B, Wheeler M, Ribbe J (2004) Forecasting with the Madden-Julian oscillation and the applications for risk management, new directions for a diverse planet. Riet 126:24–35

Flohn H (1957) Large-scale aspects of the “summer monsoon” in South and East Asia Journal of the Meteorological Society of Japan. Ser II 35A:180–186. https://doi.org/10.2151/jmsj1923.35A.0_180

Gelaro R, McCarty W, Suárez MJ, Todling R, Molod A, Takacs L, Randles CA, Darmenov A, Bosilovich MG, Reichle R, Wargan K, Coy L, Cullather R, Draper C, Akella S, Buchard V, Conaty A, da Silva AM, Gu W, Kim G, Koster R, Lucchesi R, Merkova D, Nielsen EJ, Partyka G, Pawson S, Putman W, Rienecker M, Schubert DS, Sienkiewicz M, Zhao B (2017) The modern-era retrospective analysis for research and applications, version 2 (MERRA-2). J Clim 30:5419–5454. https://doi.org/10.1175/JCLI-D-16-0758.1

Global Modelling and Assimilation Office (2021) Modern era retrospective analysis for research and applications version 2 (MERRA2). Last access March 2021.

Jones C, Carvalho L (2002) Active and break phases in the South American monsoon system. J Clim 15:905–914

Kajikawa Y, Yasunari T, Yoshida S, Fujinami H (2012) Advanced Asian summer monsoon onset in recent decades. Geophys Res Lett 39:1–5. https://doi.org/10.1029/2011GLO150540

Kanai S, Oki T, Musiake K (2002) Principal condition for the earliest Asian summer monsoon onset. Geophys Res Lett 29:34–36. https://doi.org/10.1029/2002GL015346

Kiguchi M, Matsumoto J, Kanai S, Oki T (2016) Pre-monsoon rain and its relationship with monsoon onset over the Indochina Peninsula. In Frontiers in Earth Science &amp; Earth System Sciences Vol. 4, p. 42. https://www.frontiersin.org/article/https://doi.org/10.3389/feart.2016.00042

Kuraji K, Arthorn B (2011) Rainfall variability over mountainous area in Northern Thailand. JPS International Forum “Climate Changes in Monsoon Asia” (CCMA), 6–7 Jan. 2011, 45–46

Li C, Yanai M (1996) The onset and interannual variability of the Asian summer monsoon in relation to land–sea thermal contrast. J Clim 9:358–375. https://doi.org/10.1175/1520-0442(1996)009&3c0358:TOAIVO%3e2.0.CO;2

Madden RA, Julian PR (1971) Detection of a 40–50 day oscillation in the zonal wind in the tropical Pacific. J Atmos Sci 28:702–708. https://doi.org/10.1175/1520-0469(1971)028-c0702:DOADOI%3e2.0.CO;2

Madden RA, Julian PR (1972) Description of global-scale circulation cells in the tropics with a 40–50 day period. J Atmos Sci 29:1109–1123. https://doi.org/10.1175/1520-0469(1972)029-c1109:DOASC%3e2.0.CO;2

Minoura D, Kawamura R, Matsuura T (2003) A mechanism of the onset of the South Asian summer monsoon. J Meteorol Soc Jpn 81:563–580. https://doi.org/10.2151/jmsj.81.563

Nguyen-Le D, Matsumoto J, Ngo-Duc T (2015) Onset of the rainy season in the eastern Indochina Peninsula. J Clim 28:5645–5666. https://doi.org/10.1175/JCLI-D-14-00373.1

Real-time Multivariate MJO indices (2021) RMM data. Last access June 28, 2021. http://www.bom.gov.au/climate/mjo/graphics/rmm.74toRealtime.txt

Reynolds RW, Smith TM, Liu C, Chelton DB, Casey KS, Schlax MG (2007) Daily high-resolution-blended analyses for sea surface temperature. J Clim 20:5473–5496. https://doi.org/10.1175/2007JCLI11824.1

Sato T, Kimura F (2007) How does the Tibetan Plateau affect the transition of Indian monsoon rainfall? Mon Weather Rev 135:2006–2015. https://doi.org/10.1175/MWR3386.1

Schumacher C, Ciesielski PE, Zhang MH (2008) Tropical cloud heating profiles: analysis from KWAJEX. Mon Weather Rev 136:4289–4300. https://doi.org/10.1175/2008MWR2275.1

Shrivastava S, Kar SC, Sharma AR (2017) Inter-annual variability of summer monsoon rainfall over Myanmar. Int J Climatol 37(2):802–820. https://doi.org/10.1002/joc.4741

STMD (The Thai Meteorological Department) (2021) Climate data chart. https://www.tmd.go.th/en/climate.php?FileID=7. Last access June 21, 2021.

Takahashi HG, Yasunari T (2006) A climatological monsoon break in rainfall over Indochina – A singularity in the seasonal march of the Asian summer monsoon. J Clim 19(8):1545–1556. https://doi.org/10.1175/JCLI3724.1

Tanaka N, Tantasirin C, Aranyabhaga N, Thitirojanawat P, Sapahomokh S, Arunpraparut W, Kuraji K (2019) Long-term ecologically data at Kog-Ma, Mae Moh and rainfall data at Mae Chaem watershed in northern Thailand. The University of Tokyo Forests Press, Tokyo, pp 125–140
Taylor CM, Ellis RJ, Parker DJ, Burton RR, Thornicroft CD (2003) Linking boundary-layer variability with convection: a case-study from JET 2000. Q J R Meteorol Soc 129(592):2233–2253

The World Bank (2019) Population, total–Thailand. https://data.worldbank.org/indicator/SP.POP.TOTL?locations=TH. Last access June 21, 2021

Wang B, Ho L (2002) Rainy season of the Asian-Pacific summer monsoon. J Clim 15:386–398. https://doi.org/10.1175/1520-0442(2002)015<3086:RSOTAP>2.0.CO;2

Wang B, Wu R, Lau KM (2001) Interannual variability of the Asian summer monsoon: contrasts between the Indian and the western North Pacific-East Asian monsoons. J Clim 14:4073–4090. https://doi.org/10.1175/1520-0442(2001)014<4073:IVOTAS>2.0.CO;2

Wheeler MC, Hendon HH (2004) An all-season real-time multivariate MJO index: development of an index for monitoring and prediction. Mon Weather Rev 132:1917–1932. https://doi.org/10.1175/1520-0493(2004)132<1917:AIAMIR>2.0.CO;2

Wu R, Wang B (2000) Interannual variability of summer monsoon onset over the western North Pacific and the underlying processes. J Clim 13:2483–2501. https://doi.org/10.1175/1520-0442(2000)013<2483:IVOSMO>2.0.CO;2

Xing N, Li J, Wang L (2016) Effect of the early and late onset of summer monsoon over the Bay of Bengal on Asian precipitation in May. Clim Dyn 47:1961–1970. https://doi.org/10.1007/s00382-015-2944-z

Xu J, Chan JCL (2001) First transition of the Asian summer monsoon in 1998 and the effect of the Tibet-tropical Indian Ocean thermal contrast. Journal of the Meteorological Society of Japan. Ser II 79:241–253. https://doi.org/10.2151/jmsj.79.241

Yanai M, Estensersen S, Chu JH (1973) Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. Journal of Atmospheric Sciences 30:611–627. https://doi.org/10.1175/1520-0469(1973)030<611:DOBPO>2.0.CO;2

Yanai M, Li C, Song Z (1992) Seasonal heating of the Tibetan Plateau and its effects on the evolution of the Asian summer monsoon. Journal of the Meteorological Society of Japan. Ser II 70:319–351. https://doi.org/10.2151/jmsj1965.70.1B_319

Yasutomi N, Hamada A, Yatagai A (2011) Development of a long-term daily gridded temperature dataset and its application to rain/snow discrimination of daily precipitation. Glob Environ Res 15:165–172

Zhang Y, Li T, Wang B, Wu G (2002) Onset of the summer monsoon over the Indochina Peninsula: climatology and interannual variations. J Clim 15(22):3206–3221. https://doi.org/10.1175/1520-0442(2002)015<3206:OOTSMO>2.0.CO;2

Zhang Y, Fan G, Hua W, Zhang Y, Wang B, Lai X (2017) Differences in atmospheric heat source between the Tibetan Plateau-South Asia region and the southern Indian Ocean and their impacts on the Indian summer monsoon outbreak. Journal of Meteorological Research 31:540–554. https://doi.org/10.1007/s13351-017-6042-5

Zhou ZQ, Xie SP, Zhang R (2019) Variability and predictability of Indian rainfall during the monsoon onset month of June. Geophys Res Lett 46:14782–14788. https://doi.org/10.1029/2019GL085495

Publisher’s Note Springer Nature remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.