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**Revealing the relationship between ENSO and the annual cycle of the East Asia
and western Pacific monsoon using a low-level circulation weather-type calendar**

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ABSTRACT

26 This paper proposes a novel approach to identifying the annual cycle and variability of the East Asia
27 and western Pacific (EA-WP) monsoon. A low-level circulation weather type (WT) calendar is
28 constructed based on nine WTs obtained from K-means cluster analysis for 46 years (1979-2024) of
29 daily 850-hPa wind data. The WT daily occurrence frequency reveals climatological features of the
30 monsoon annual cycle and seasonal progression. The WT-based summer and winter monsoon indices
31 well represent key monsoon characteristics. A strong East Asian summer monsoon (EASM)
32 corresponds to more frequent occurrences of WT featuring low-level anticyclonic circulation over the
33 western North Pacific, while a strong East Asian winter monsoon (EAWM) corresponds to more
34 frequent occurrences of WTs characterized by low-level northerly winds over the South China Sea and
35 the Philippine Sea. The variability of EASM and EAWM is synchronized during the developing phase
36 of ENSO events. Strong (weak) EA-WP monsoon years are marked by strong (weak) EASM followed
37 by strong (weak) EAWM, whereas no clear relationship is found between EASM and its preceding
38 EAWM. Approximately, 86% (76%) of the strong (weak) EA-WP monsoon years are associated with
39 La Niña (El Niño) winters. The period from mid-July to late-August is identified as a critical window
40 for the synchronization of monsoon and ENSO. Observational evidence suggests that strong and weak
41 air-sea coupling represent two distinct states of EA-WP monsoon years. Under strong air-sea coupling,
42 monsoon variability is significantly enhanced, with EA-WP monsoon and ENSO functioning as two
43 interconnected components of a single coupled system.

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1. Introduction

Weather and climate over the Asia-Australia-Pacific region (20°S-60°N, 50°E-180°) are dominated by a grand monsoon system formed by the East Asian, South Asian and western North Pacific monsoons in the Northern Hemisphere, and the Indonesian-Australian monsoon in the Southern Hemisphere. Climatological features of the grand Asian-Australian monsoon (A-AM) system exert profound societal impacts, influencing the daily lives of nearly 60% of the global population. The monsoon system exhibits a pronounced annual cycle associated with a tropical convective maximum that migrates meridionally and zonally across the Indian and Pacific sectors. This migration broadly follows the seasonal displacement of the intertropical convergence zone (ITCZ): from northern summer to winter, convection shifts southward and eastward, and from northern winter to summer it migrates northwestward (Webster et al. 1998; Chao and Chen 2001; Geen et al. 2020). Cheng et al. (2024) referred to this migrating convective maximum as the Intertropical Convective Cell (ITCC) and identified eight distinct phases during its seasonal progression along the Asian–Maritime Continent land bridge.

Earlier studies Meehl (1987, 1994) demonstrated that strong summer monsoon convection over India and Southeast Asia tends to persist from boreal summer into boreal winter, subsequently evolving into enhanced convection within the South Pacific Convergence Zone (SPCZ). Consequently, a stronger SPCZ during boreal winter is often preceded by a stronger Asian summer monsoon. Years characterized by enhanced convection persisting from boreal summer through boreal winter are commonly referred to as strong annual-cycle years. Meehl (1987) further suggested that intensified SPCZ convection enhances subsidence over the tropical Indian Ocean, thereby contributing to a weaker Asian summer monsoon in the following year. This mechanism underlies the biennial tendency of tropical Indo-Pacific convection, which can occur independently of ENSO, although ENSO-related anomalies may amplify it. The pronounced biennial signal in A–AM precipitation (Meehl, 1987, 1993)

was subsequently confirmed by Wang et al. (2008) using season-reliant empirical orthogonal function (S-EOF) analysis applied to monsoon-year precipitation (Yasunari, 1991).

In contrast, studies focusing on East Asia suggest a more complex cross-season relationship between the East Asian summer monsoon (EASM) and winter monsoon (EAWM). Wu and Chan (2005a, 2005b) examined the relationship between Asian summer and winter monsoons during the period of 1958–1999 and found that a discernible linkage between the summer and winter monsoons emerges primarily when the monsoons are modulated by ENSO. Specifically, the summer monsoon strength unlikely tends to be weak during El Niño developing years and it unlikely tends to be strong during the following year, while the winter monsoon tends to be weak during El Niño winters. When the winter monsoon is coupled with during La Niña developing years it tends to be stronger, while the monsoon during the following summer tends to be weaker. They concluded that the biennial oscillation evident in tropical monsoons (Meehl 1987, 1993) is not clearly manifested over East Asia. Subsequent work by Chen et al. (2013) demonstrated that the EASM–EAWM linkage is mediated by the El Niño-induced western North Pacific anticyclone (WNPAC), which persists from boreal winter into the following summer. More recently, Yu et al. (2021) showed that the in-phase relationship between strong (weak) summer and winter monsoons arise from distinct mechanisms. The transition from a strong EASM to a strong EAWM is associated with a summer WNPAC and a winter western North Pacific cyclone (WNPC), whereas the weak-to-weak transition involves a summer WNPC and a winter WNPAC. ENSO plays a dominant role in the in-phase transitions, which are more pronounced during El Niño than La Niña events. Related findings were reported by Wu et al. (2014) for the South China Sea (SCS), where in-phase winter-to-summer rainfall anomalies occur more frequently during ENSO decaying years, while out-of-phase summer-to-winter anomalies are more common during ENSO developing years. The in-phase relationship was attributed to the east-west circulation coupled with the SST contrast between the equatorial central Pacific and the western Pacific, while the out-of-phase relationship was attributed to the Rossby-wave response to local SST anomalies over the Philippine Sea and the equatorial western North Pacific.

Most previous studies have relied on analyses of monthly or seasonal mean wind and precipitation anomalies, or on leading modes derived from empirical orthogonal function (EOF) techniques. Consequently, the role of daily low-level circulation in shaping the cross-seasonal linkage between the EAWM and EASM remains insufficiently explored. Addressing this gap is critical for improving seamless prediction across weather and climate time scales. Using 40 years (1979-2018) of daily 850-hPa wind data, Qian et al. (2021) demonstrated that the development of the EASM from April 1 to July 31 can be characterized by eight weather types (WTs). The seasonal evolution of these WTs is gradual yet intermittent, capturing essential weather-within-climate information. Interannual variations in WT frequency and associated precipitation anomalies reveal systematic modulation by ENSO.

Building on Qian et al. (2021), the present study extends the analysis to 46 years (1979–2024) of year-round daily 850-hPa wind fields over an expanded Asia–Australia–Pacific domain (20°S-60°N, 50°E-180°E). The primary objective is to investigate the relationship between daily circulation weather types and variability in the monsoon annual cycle. Specifically, this study addresses the following questions:

1. Can the monsoon annual cycle and its variability be succinctly represented using a weather-type calendar derived from cluster analysis of daily low-level circulation patterns?
2. What cross-season linkage of the EAWM and EASM are revealed within such a weather-type framework?
3. How does ENSO influence the cross-season linkage between the EAWM and EASM?

The data and methodology are described in Section 2. Section 3 presents the weather-type classification and the resulting low-level circulation calendar. Section 4 examines the monsoon annual cycle and EASM–EAWM linkage as depicted by the weather-type calendar. Section 5 discusses the role of ENSO, and Section 6 summarizes the main findings and conclusions.

2. Data and Method

2.1 Data

Atmospheric circulation variables and precipitation are from the fifth version of European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA5) for the period 1979-2024. ERA5 data are provided at a horizontal resolution of $0.25^\circ \times 0.25^\circ$. Daily fields are calculated as the mean of four synoptic times (0000, 0600, 1200, and 1800 UTC).

Deep convection is identified using the interpolated daily outgoing longwave radiation (OLR) version 1.2 with the spatial resolution of $1^\circ \times 1^\circ$ (Lee 2011). Daily sea surface temperature (SST) data are taken from the NOAA Extended Reconstructed Sea Surface Temperature dataset, version 5 (ERSST v5; Huang *et al.* 2017).

2.2 Weather Typing Analysis

The weather typing analysis procedure follows that of Qian et al. (2021); a schematic overview is shown in Fig. 1. Additional methodological details can be found in Roller et al (2016). The analysis consists of three major steps. First, the dimensionality of the daily 850-hPa wind fields is reduced using EOF analysis. A total of 72 EOF modes are retained, accounting for approximately 75% of the total variance. Second, the reduced-dimension data are partitioned using a K-means clustering algorithm, and the within-cluster similarity and between-cluster separation are evaluated. The number of clusters, K , is prescribed a priori and varied from 2 to a sufficiently large number (up to 12). For each K , the clustering procedure is repeated N times with different initial conditions, yielding an $N \times K$ matrix of anomaly correlation coefficients (ACCs) among cluster centroids that is the central point of a cluster represented by the mean value of the data points within the cluster. Third, the optimal number of clusters is determined using the classifiability index (CI; Michelangeli et al. 1995). For a given cluster i of partition P (denoted as P_i) and the centroids of cluster j of partition Q (denoted as Q_j), where $1 \leq i, j \leq K$ and partition $Q \neq P$. After obtaining all pairs of $ACC(P_i, Q_{j=1-K})$, then choose the maximum

ACC scores between P_i and each cluster j in a partition Q , and assign the average of all Q partitions to $ACC[P, i]$, which is the average ACC score between P_i and all the other portions. CI is defined as the average of all elements in ACC matrix. The higher CI means that clustering in k groups has higher average ACC among all kinds of partitions through N iterations. In practice the optimal number of clusters is identified as the value of K corresponding to the maximum CI following the first local minimum (the “elbow” criterion). In this study, the CI values computed using $N = 200$ iterations are shown in Fig. 2. A clear elbow occurs at $K = 9$, indicating that nine clusters represent the optimal classification of daily 850-hPa circulation patterns.

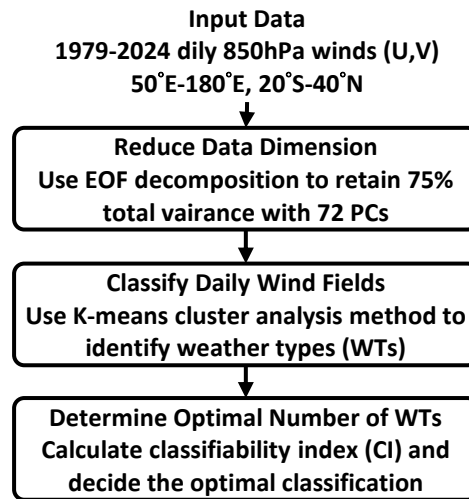


Fig. 1 Schematic illustration of the input data and major analysis steps of weather typing procedure.

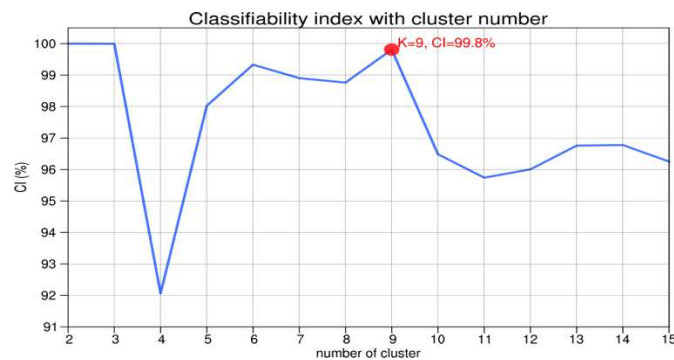


Fig. 2 The Classifiability Index (CI) of the K-means clustering analysis applied to 46 years (1979–2024) of daily 850-hPa wind fields over the Asia–Australia–Pacific region (20°S–60°N, 50°E–180°E). The optimal number of clusters is nine.

2.3 Monsoon index

The monsoon indices used in this study follow Yu et al. (2021). A larger East Asian summer monsoon index (EASMI) corresponds to a more pronounced Meiyu–Baiu front and enhanced monsoonal rainfall, while a larger East Asian winter monsoon index (EAWMI) indicates stronger northeasterly monsoonal flow and colder conditions over Southeast Asia.

The EASMI is calculated as the June–August mean 850-hPa zonal wind averaged over (5°–15°N, 90°–130°E) minus that averaged over (22.5°–32.5°N, 110°–140°E). This index was originally proposed in Wang and Fan (1999) as the DU2 index and was shown to be strongly correlated with low-level southerly shears averaged over the combined areas (5°S–5°N, 90°–120°E) and (5°–15°N, 120°–145°E). It has been widely adopted as a representative measure of East Asian summer monsoon intensity.

The EAWMI is calculated as the December–February mean 850-hPa meridional wind averaged over (20°N–40°N, 100°E–140°E), following Yang et al. (2002). This index is highly correlated with variations in the upper-level East Asian jet stream. The sign of the EAWMI is reversed in this study so that larger values correspond to stronger northeasterly monsoonal flow.

2.4 El Niño and La Niña events

El Niño and La Niña events are identified using the operational definition of the National Oceanic and Atmospheric Administration Climate Prediction Center (NOAA CPC). The Oceanic Niño Index (ONI) is defined as a 3-month running mean SST anomalies in the Niño 3.4 region (5°S–5°N, 170°W–120°W), calculated relative to a 30-year climatology updated every five years. A warm (cold) ENSO episode is identified when the ONI exceeds +0.5°C (falls below –0.5°C) for at least five consecutive

overlapping seasons. Warm episodes are referred to as El Niño events, and cold episode as La Niña events.

3. Spatial pattern and seasonal evolution of the Weather Types

As described in in Section 2, the 46-year (1979–2024) daily 850-hPa wind (UV850) fields over the Asia–Australia–Pacific region (50°E–180°E, 20°S–40°N) are classified into nine weather types (WTs). These weather types are denoted WT1–WT9, with the numbering reflecting the sequence of their preferred occurrence during the annual cycle. Composite 850-hPa circulation patterns for each WT are shown in Fig. 3. WT1–WT3 and WT8–WT9 are characterized by easterly zonal flow in the 0°–15°N latitude band over the Indian Ocean, whereas WT4–WT7 exhibit westerly zonal flow in the same region. The easterly regimes are associated with southward cross-equatorial flow along the western boundaries of both the Indian and Pacific Oceans, while the westerly regimes correspond to northward cross-equatorial flow. The cross-equatorial flow is substantially stronger over the western Indian Ocean than over the South China Sea and the Philippine Sea.

Fig. 4 presents a low-level circulation calendar depicting the daily occurrence of WTs from 1979 to 2024. The WT sequence evolves gradually but intermittently, with stepwise transitions from one type to another. These transitions reflect weather-within-climate variability regulated by the monsoon annual cycle. A notable feature is the seasonal contrast in WT dominance: several months are characterized by the prevalence of a single WT, whereas boreal winter months (December–February) exhibit pronounced diversity, with multiple WTs occurring frequently. The percentage occurrence of each WT on each calendar date (Fig. 5) highlights this wintertime complexity. During December–February, no single WT exceeds 60% of the total occurrence. WT9, WT1 and WT2 alternate in dominance throughout the winter, although WT9 is relatively more frequent during mid-December. March represents a transitional period, with the first half of the month characterized by frequent occurrences of WT9 and WT1–WT3, followed by a marked increase in WT3 and a decline in WT1 and WT2 during the latter half of the month.

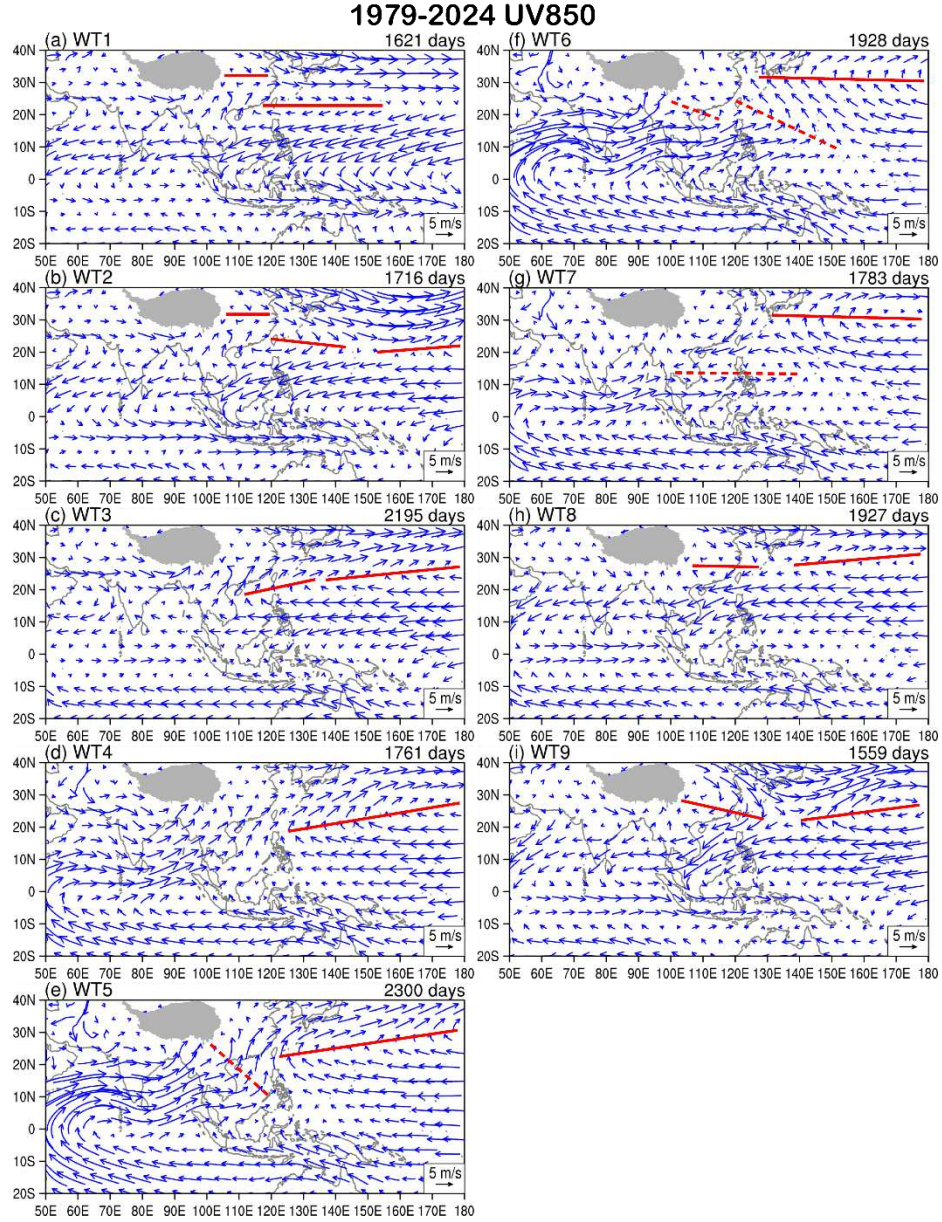


Fig. 3 Composite 850-hPa horizontal winds (vectors, m s^{-1}) over the Asian-western Pacific region for (a) WT1, (b) WT2, (c) WT3, (d) WT4, (e) WT5, (f) WT6, (g) WT7, (h) WT8, and (i) WT9. The total number of days for each weather type is indicated in the upper right corner of each panel. Cyclonic troughs and anticyclonic ridges are denoted by red dashed and solid lines, respectively. Gray shading indicates topography above 4000 m.

From April to November, the seasonal progression is characterized by six distinct peaks in WT occurrence: WT3 in April, WT4 in May, WT5 during June-July, WT6 in August, WT7 from late-

September to October, and WT8 in November. This orderly progression of dominant WTs provides a compact representation of the monsoon annual cycle. The preferred timing of each WT (Figs. 4 and 5) closely follows the seasonal evolution of solar insolation, with consistent phase locking from year to year.

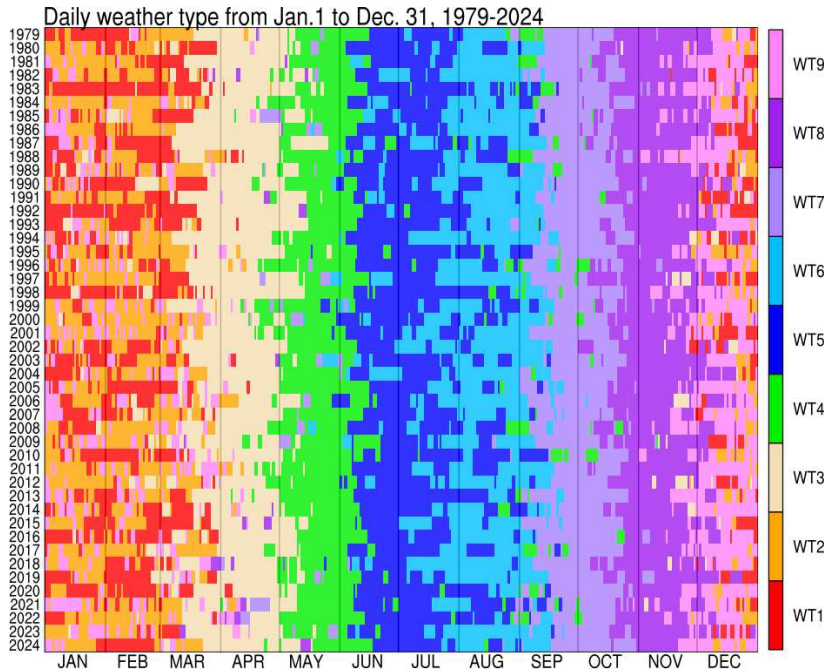


Fig. 4 Daily occurrence of weather types from 1 January 1979 to 31 December 2024. The y-axis increases downward from 1979 (top) to 2024 (bottom), and colors denote different weather types.

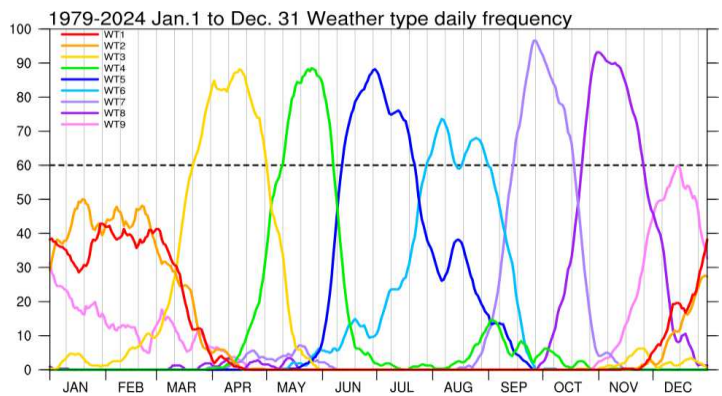


Fig. 5 Percentage occurrence of each weather type for each calendar date over the 46-year period (1979–2024). The dashed line indicates a 60% occurrence threshold.

To further elucidate the differences among WTs, composite anomalies of UV850 and precipitation are shown in Fig. 6. Anomalies are computed by subtracting a WT-weighted monthly mean from the composite mean of daily fields for each WT. The weights are derived from the climatological monthly frequency of each WT over the 46-year period. This WT-weighted monthly mean represents the climatological annual cycle implied by the daily WT occurrences. WT1 (Fig. 6a) exhibits circulation and precipitation anomalies characteristic of boreal winter dry conditions. The Australia–Indonesia–Indochina land bridge is anomalously dry, whereas the western Indian Ocean and western South Pacific are anomalously wet. The circulation anomalies feature easterlies over the tropical Indian Ocean and westerlies over the Pacific. In the meridional direction the circulation anomalies feature northerly over the Philippine Sea and southerly over Australia and the western South Pacific, with low-level convergence near Indonesia and Papua New Guinea. Enhanced equatorial westerlies east of 150°E are associated with intensified convection over the near-equatorial western Pacific, indicative of a strengthened ITCZ.

The anomalous patterns of WT2 (Fig. 6b) are largely opposite to those of WT1. WT2 and WT9 show similar precipitation structures, although WT9 exhibits wetter anomalies north of the equator and stronger northerly flow extending from continental East Asia to the South China Sea and across the equator (Fig. 6i). The northerly anomalies are linked to an anticyclonic anomaly over southern China and a cyclonic anomaly centered over northern Japan for WT9, while the center for WT2 is shifted eastward toward the southwestern Aleutian region. Both WT2 and WT9 represent strong East Asian winter monsoon conditions, characterized by intensified northeasterlies along the East Asian coastal regions, a deepened Aleutian Low, and enhanced rainfall over the Maritime Continent, northern Australia, and the western Pacific. Both WT2 and WT9 have similar equatorial circulation, whereas WT2 exhibits stronger monsoon over Australia.

During boreal summer, WT4 (Fig. 6d) features anomalous cyclonic circulation accompanied by enhanced Meiyu–Baiu rainfall extending from southern China northeastward through Taiwan and the Bonin Islands to the northwestern Pacific. Over tropical Indian Ocean, anomalous cyclonic circulation

over southeastern Arabian Sea and eastern Indian Ocean produce equatorial westerly anomalies. These westerlies converge with anomalous easterlies from the Pacific over the Maritime Continent, generating wet anomalies. To the north, a broad region of suppressed precipitation extends from the northern Indian subcontinent eastward to the South China Sea and the western Pacific warm pool.

WT5 (Fig. 6e) is characterized by an anomalous anticyclone over the subtropical western North Pacific, associated with dry conditions near the anticyclone center and enhanced precipitation over surrounding regions, including South Asia, East Asia, and the Maritime Continent. In contrast, WT6 (Fig. 6f) exhibits an anomalous cyclonic circulation over the subtropical western North Pacific with enhanced rainfall near its center and drier conditions in the surrounding regions. WT7 (Fig. 6g) represents a weakened and southeastward-shifted version of WT6, with wet anomalies over the northern Indian subcontinent and Bay of Bengal displaced southward from approximately 20°N (WT6) to 15°N. WT8 (Fig. 6h) marks a rapid transition in off-equator tropical zonal wind anomalies over South Asia (10°-20°N), from westerly to easterly. Over the equatorial Indian Ocean, cyclonic (counterclockwise) anomalies enhance cross-equatorial flow along the western boundary and produce wet conditions near the circulation center. The evolution of these anomalous circulation patterns is consistent with the seasonal WT frequency distribution (Fig. 5), with WT8 occurring most frequently from mid-October to late November.

1979-2024 Prec. & UV850 Anomaly

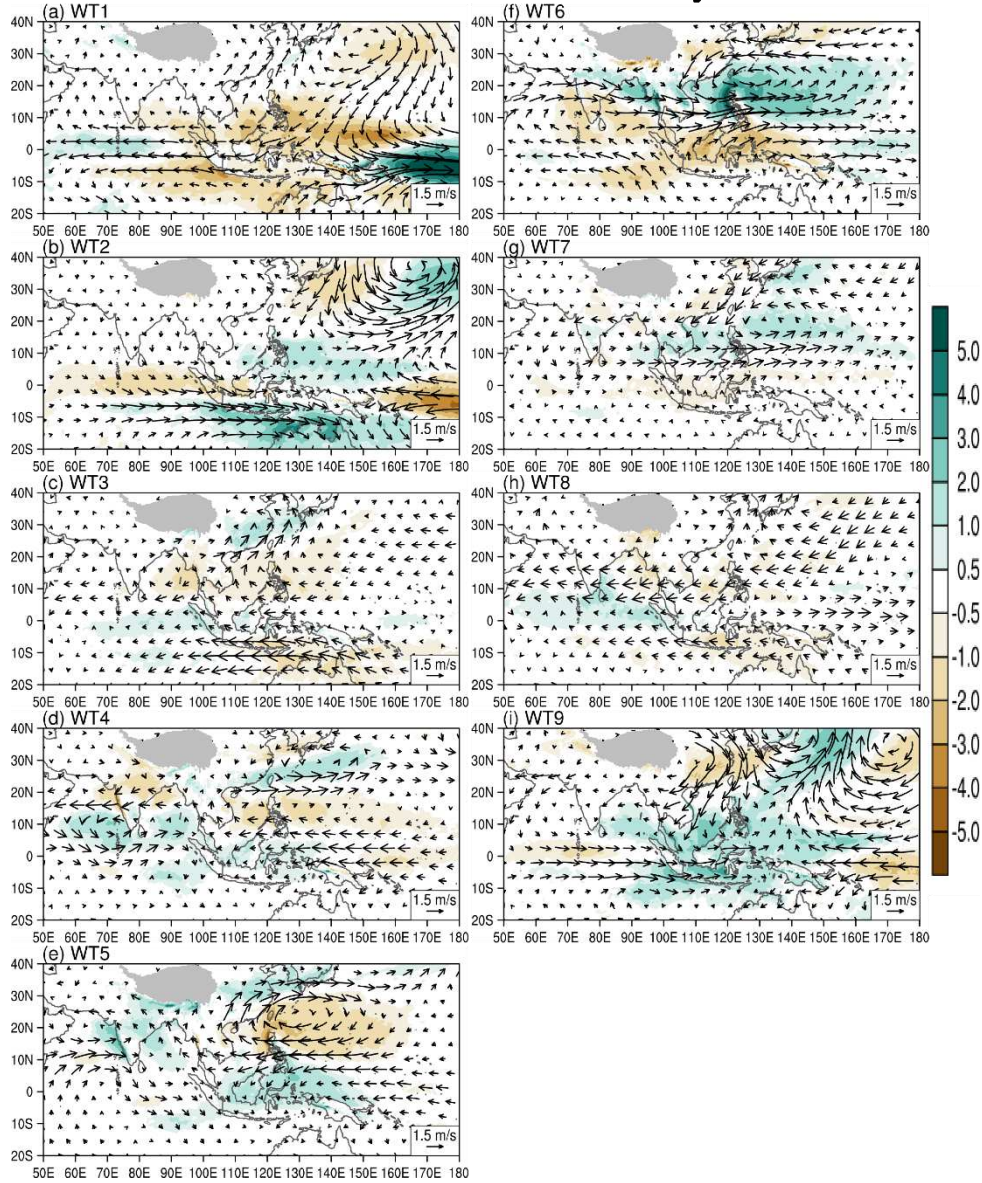


Fig. 6 Composite anomalies of 850-hPa horizontal winds (vectors, m s^{-1}) and precipitation (shading, mm day^{-1}) for (a) WT1, (b) WT2, (c) WT3, (d) WT4, (e) WT5, (f) WT6, (g) WT7, (h) WT8, and (i) WT9. Gray shading indicates topography above 4000 m.

4. East Asia-Western Pacific monsoon annual cycle revealed by weather types

The monsoon annual cycle may be viewed as a planetary-scale overturning circulation, with ascending motion in the summer hemisphere and descending motion in the winter hemisphere. The ascending branch is driven by enhanced solar insolation and persistent tropical deep convection

(Webster et al. 1998). As demonstrated in Section 3, 46 years of daily 850-hPa wind fields over the Asian–Australian–western Pacific region can be concisely represented by nine weather types (WTs). The composite circulation and precipitation anomalies associated with WT occurrence successfully capture key features of the monsoon annual cycle. A notable feature revealed by the WT framework is that boreal winter circulation during January–February cannot be represented by a single dominant WT. Instead, three WTs occur with comparable frequency (Fig. 5), indicating the presence of multiple winter climate states. In this section, we first characterize these boreal winter climate states using WT frequencies and then examine their relationships with preceding and subsequent summer climate states. The results show that the winter climate state is strongly linked to the preceding summer but exhibits only a weak connection with the following summer. This asymmetry reflects two distinct types of seasonal progression in the East Asia–western Pacific (EA–WP) monsoon annual cycle.

4.1 Boreal winter climate states, strong and weak monsoon years

To characterize boreal winter climate states, we examine the relationships among the three dominant winter weather types—WT1, WT2, and WT9—using scatterplots of their standardized winter (DJF) frequency anomalies (Fig. 7). WT1 is strongly anticorrelated with WT2 ($r = -0.9$) and WT9 ($r = -0.7$), whereas WT2 and WT9 are not significantly correlated ($r = 0.4$). As shown in Figure 5, WT9 occurs most frequently in early winter (December) and decreases rapidly after mid-January, while WT2 peaks in mid-January and persists through early March. Composite circulation and precipitation anomalies (Fig. 6) indicate that both WT2 and WT9 represent strong East Asian winter monsoon (EAWM) conditions.

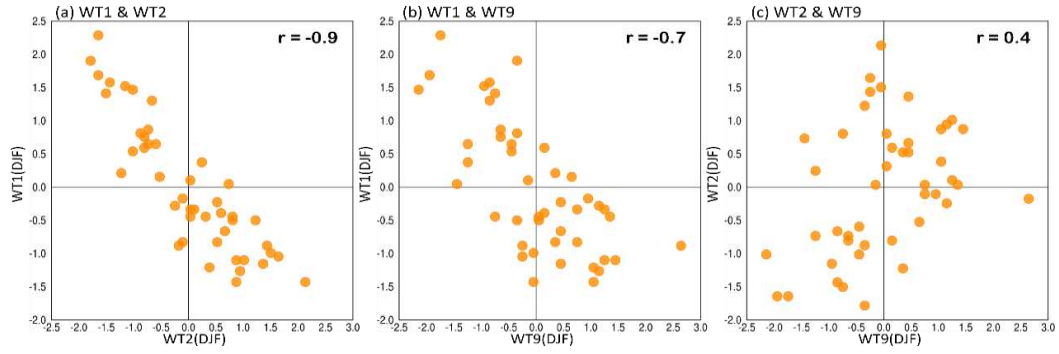


Fig. 7. Scatterplots of the standardized winter frequency anomalies for (a) WT1 and WT2, (b) WT1 and WT9, and (c) WT2 and WT9. The correlation coefficient for each pair is shown in the upper-right corner of each panel.

Based on these results, we define a winter WT index as the sum of WT2 and WT9 days during December–February. Regression of seasonal-mean ERA5 925-hPa winds onto this index reveals enhanced northeasterly flow over the East China Sea, South China Sea (SCS), and Philippine Sea (Fig. 8a). These northeasterlies intrude from northeastern Asia into the SCS and converge with westerly flow from the tropical Indian and Pacific Oceans, forming a cyclonic circulation centered near the Borneo–Philippine region. This circulation pattern is consistent with the composite anomalies shown in Fig. 6. Fig. 8b compares the WT-based winter index with the conventional EAWM index of Yu et al. (2021). The two indices are highly correlated ($r = 0.7$), exceeding the 99% confidence level. Accordingly, the WT-based EAWM index used here is defined as the sum of WT2 and WT9 days during DJF and is hereafter referred to as the WT9+2 index.

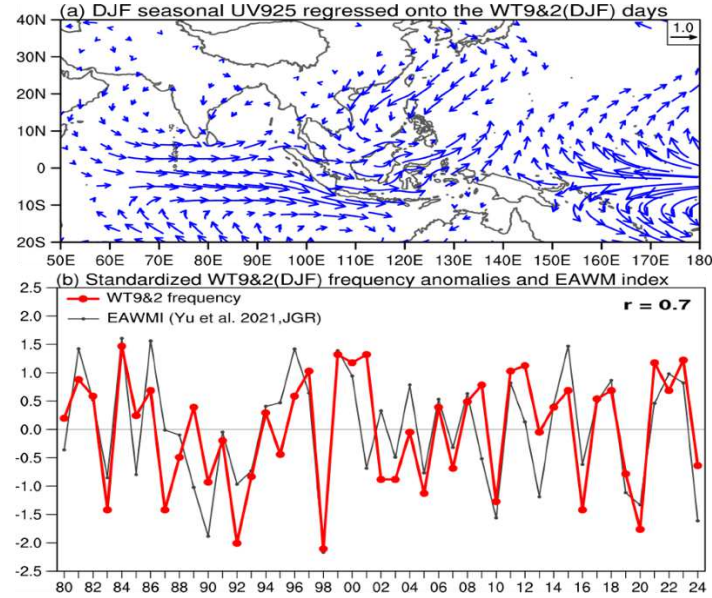


Fig. 8. (a) December-February (DJF) seasonal ERA5 925-hPa horizontal winds regressed onto the sum of WT9 and WT2 days during DJF. Only vectors significant at the 95% confidence level are shown. (b) Time series of the normalized sum of the WT9 and WT2 days (red line) and the normalized EAWM index (Yu et al. 2021) (black line). The correlation coefficient is shown in the upper-right corner of the panel.

Using standardized anomalies of the WT9+2 index, EAWM years are classified as strong (weak) when the index exceeds +0.5 (falls below -0.5); remaining years are classified as normal. The upper portion of Table 1 lists strong and weak EAWM years, where the year corresponds to January. To examine the evolution of WT frequencies across seasons, daily WT occurrences are aggregated into pentads. Fig. 9 shows pentad differences in WT frequency between strong and weak EAWM years over a two-year window spanning the preceding winter through the subsequent summer. During boreal winter (pentads 67–12), strong EAWM years exhibit significantly higher frequencies of WT2 and WT9 and lower frequencies of WT1 than weak EAWM years, consistent with their circulation characteristics (Fig. 6). The fine-scale evolution evident in Fig. 9 highlights the utility of the WT framework for linking daily weather variability with seasonal climate states.

Table 1. Years of the strong/weak East Asian Winter Monsoon (EAWM) and East Asian Summer Monsoon (EASM). Superscript "E" represents El Niño winter, "L" represents La Niña winter, "E0" ("L0") represents the summer before El Niño (La Niña), and "E1" ("L1") represents the summer after El Niño (La Niña) winter.

EAWM	Strong	1981,1982,1984 ^L ,1986,1996 ^L ,1997,1999 ^L ,2000 ^L ,2001 ^L ,2009 ^L ,2011 ^L ,2012 ^L ,2015 ^E ,2017 ^L ,2018 ^L ,2021 ^L ,2022 ^L ,2023 ^L
	Weak	1983 ^E ,1987 ^E ,1990,1992 ^E ,1993,1998 ^E ,2002,2003 ^E ,2005 ^E ,2007 ^E ,2010 ^E ,2016 ^E ,2019 ^E ,2020,2024 ^E
EASM	Strong	1980 ^{E0} ,1983 ^{E0/L1} ,1988 ^{E0/L1} ,1995 ^{E0/L1} ,1998 ^{E0/L1} ,1999 ^{L0/L1} ,2000 ^{L0/L1} ,2003 ^{E0} ,2008 ^{L0/L1} ,2010 ^{E0/L1} ,2013,2017 ^{L0/L1} ,2020 ^{L1} ,2021 ^{L0/L1} ,2022 ^{L0/L1}
	Weak	1981,1982 ^{E1} ,1985 ^{L0} ,1986 ^{E1} ,1994 ^{E1} ,1997 ^{E1} ,2002 ^{E1} ,2004 ^{E1} ,2006 ^{L0/E1} ,2009 ^{L0/E1} ,2015 ^{E0/E1} ,2018 ^{L0/E1} ,2019 ^{E0} ,2023 ^{L0/E1}

In addition to wintertime contrasts, Figure 9 reveals pronounced differences during the preceding summer. Strong EAWM years are preceded by summers with more frequent WT5 occurrences than WT6, with the sharpest contrast during pentads 41–48 (20 July–28 August). As discussed in Section 3, WT5 is associated with a strengthened western North Pacific subtropical high (Fig. 6e), which favors a persistent Meiyu–Baiu front and enhanced precipitation, characteristic of a strong East Asian summer monsoon (EASM) (Wang and Fan 1999; Yu et al. 2021). Based on this relationship, we define a WT-based EASM index as the total number of WT5 days during June–August, referred to as the WT5 index. Regression of JJA mean 850-hPa winds onto the WT5 index (Fig. 10a) shows enhanced easterly and southeasterly flow transporting warm, moist air into East Asia, consistent with strong EASM conditions. The WT5 index is highly correlated ($r = 0.8$; >99% confidence level) with the conventional EASMI (Wang and Fan 1999; Yu et al. 2021).

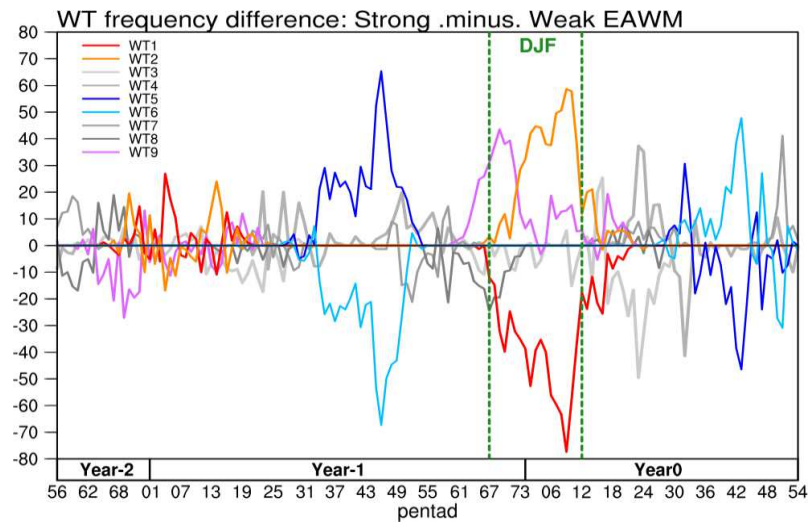


Fig. 9 Time evolution of pentad differences in each weather type (WT) between strong and weak EAWM. The green dashed lines mark the time interval of the winter (DJF) season.

Following the same procedure as for winter, EASM years are classified as strong or weak based on standardized WT5 anomalies exceeding ± 0.5 . The lower portion of Table 1 lists strong and weak EASM years. Comparison of EASM and EAWM classifications reveals that most strong (weak) EASM years are followed by strong (weak) EAWM conditions in the subsequent winter. Fig. 11 further illustrates this relationship by showing pentad WT frequency differences between strong and weak EASM years over a two-year window. During boreal summer (pentads 31–49), strong EASM years are characterized by substantially higher WT5 and lower WT6 frequencies, particularly during pentads 41–48. This period coincides with a secondary climatological peak in WT5 frequency (Figure 5). In contrast, WT differences during the preceding winter are relatively weak, while contrasts during the following winter—characterized by increased WT2 and WT9 and reduced WT1—are clearly evident.

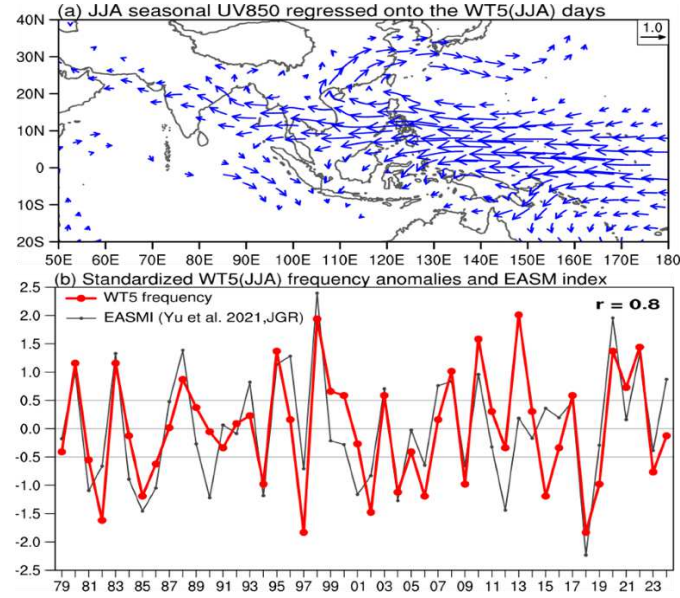


Fig. 10 As in Fig. 8, but showing (a) June-August (JJA) ERA5 850-hPa horizontal winds regressed onto the JJA WT5 annual frequency, and (b) the normalized time series of the JJA WT5 annual frequency (red) and the summer (JJA) mean EASM index (black).

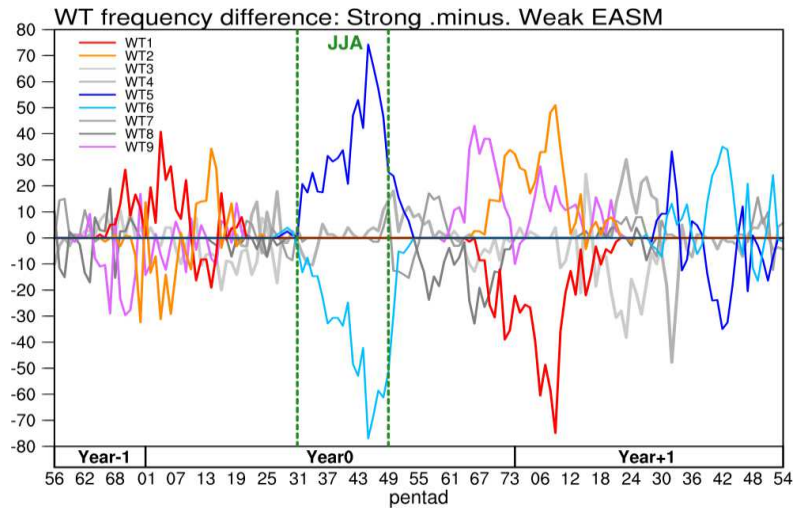


Fig. 11. Similar to Fig. 9, but showing the differences between strong and weak EASM. The green dashed lines denote the time interval corresponding to the summer (JJA) season.

Overall, the WT analysis demonstrates that the EA–WP monsoon annual cycle exhibits two distinct modes. In the strong mode, a strong EASM precedes a strong EAWM, whereas in the weak mode, a weak EASM precedes a weak EAWM. No systematic linkage is identified between the EAWM and the

following EASM. These results provide a WT description of the EA–WP monsoon annual cycle in terms of a monsoon year, defined as the period from 1 May to 30 April of the following year (Yasunari, 1991). A strong (weak) monsoon year is defined as one in which both the EASM and EAWM are anomalously strong (weak).

4.2 Two types EA-WP monsoon annual cycle

As shown in Table 1, among 18 strong EAWM years, 13 occur during La Niña winters, one during an El Niño winter, and four during neutral ENSO conditions. Conversely, among 15 weak EAWM years, 11 occur during El Niño winters, four during neutral winters, and none during La Niña winters. The relationship between monsoon years and ENSO phase is summarized in the scatterplot of standardized WT5 and WT9+2 indices (Figure 12), with El Niño (La Niña) winters marked in red (blue). Strong monsoon years correspond to cases where both indices exceed +0.5, while weak monsoon years correspond to values below −0.5. Among 17 strong monsoon years, four (24%) are not associated with La Niña winters, whereas among 14 weak monsoon years, only two (14%) are not associated with El Niño winters. These results indicate that strong and weak monsoon years arise from climate states in which the progression of the EA–WP monsoon annual cycle is tightly coupled with ENSO evolution. Years that do not meet the strong or weak criteria are classified as neutral monsoon years, during which EASM and EAWM variations are less coherent. Of the 14 neutral years, eight (57%) occur during ENSO normal winters. As this study focuses on monsoon years with coherent seasonal progression, neutral monsoon years are not discussed further.

We also checked the scatterplot of the standardized conventional EASMI and EAWMI (Yu et al. 2021) (Fig. S1) and found that the proposed WT-based monsoon indices (Fig. 12) capture better monsoon years and ENSO phase relationship.

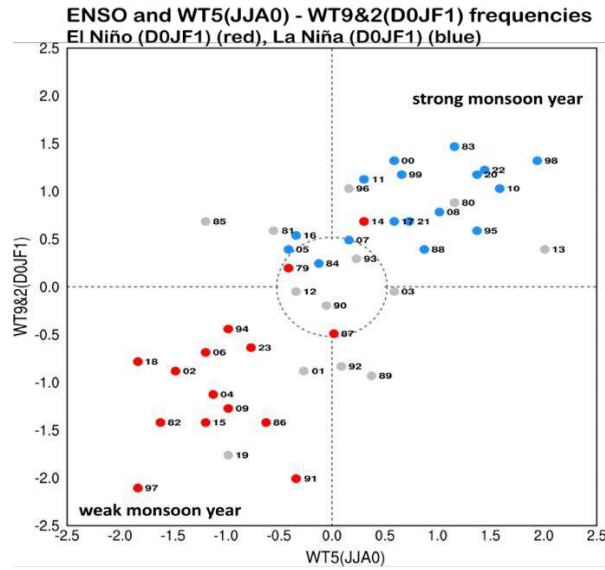


Fig. 12 Scatterplot of the standardized WT5 and WT9+2 indexes. The years with El Niño winter are marked in red and the years with La Niña winter are marked in blue. The radius of the dashed circle is 0.5, hence the years within the circle correspond to weak monsoons. The years to the upper right side of the circle are strong monsoon years of which both WT5 and WT9+2 indices are larger than 0.5. The years to the lower left side of the circle are weak monsoon years of which both WT5 and WT9+2 indices are less than -0.5. The correlation of two indices is 0.7.

5. Coupling of EA-WP monsoon annual cycle and ENSO SSTs

The contrast in ENSO evolution between strong and weak monsoon years is evident in the two-year evolution of the Oceanic Niño Index (ONI; Niño-3.4 SST anomalies), spanning from the summer preceding a monsoon year to the summer following it (Fig. 13). For strong monsoon years (red symbols in Fig. 13), the median ONI falls below -0.5°C during August–September–October, reaches its coldest values in October–November–December, and then gradually weakens, returning above -0.5°C by April–May–June. In contrast, weak monsoon years (blue symbols) exhibit median ONI values exceeding $+0.5^{\circ}\text{C}$ by July–August–September, peaking in October–November–December, and then rapidly declining after March to below $+0.5^{\circ}\text{C}$ by April–May–June. Notably, differences in ONI between strong and weak monsoon years are not discernible prior to the winter of the preceding

monsoon year. A gradual separation of ONI medians emerges during the spring before monsoon onset, remains relatively steady during early summer (May–July), and then amplifies rapidly from late summer into early winter. This timing coincides with the period of strongest contrast in summer weather types, identified during the 40-day interval from pentads 41 to 48 (20 July–28 August; Figs. 9 and 11). This correspondence suggests that this interval represents a critical window during which the EA–WP monsoon low-level circulation and ENSO SST anomalies may become dynamically coupled and evolve coherently.

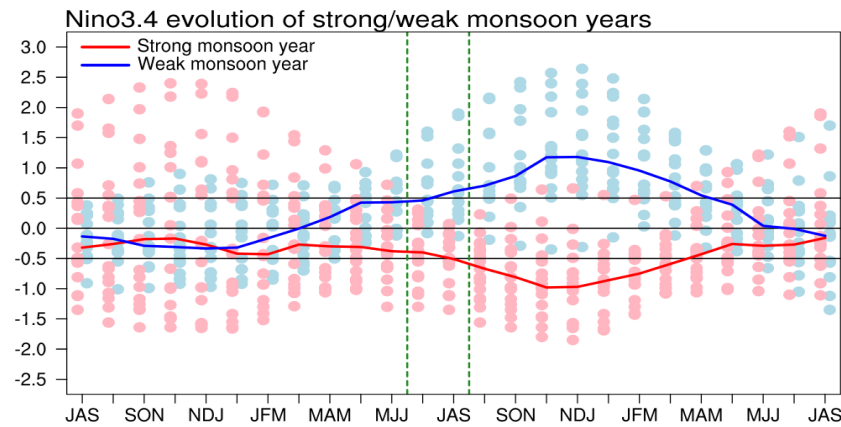


Fig. 13 Time evolution of the 3-month running mean Niño 3.4 index for strong (red) and weak (blue) monsoon years. Dots denote individual years, and solid curves indicate the median values for each category.

To examine this hypothesis, we analyze composite large-scale dynamic and thermodynamic fields regressed onto the total number of WT5 days during pentads 41–48. The regression coefficients of the bi-monthly mean fields (SST, precipitation, OLR, 850-hPa and 200-hPa winds, 850-hPa and 200-hPa velocity potential, 850-hPa and 200-hPa stream function) are presented in Figs. S3–S11. Fig. 14 presents results for strong monsoon years; corresponding composites for weak monsoon years show similar spatial structures with opposite polarity (Fig. S12). During April–May (Fig. 14a), anomalous anticyclonic circulation appears over the Philippine Sea and western North Pacific, accompanied by anomalous westerlies over the eastern Indian Ocean south of Sumatra. Associated SST anomalies

include negative values over the North Pacific and positive anomalies over the eastern tropical South Indian Ocean and eastern equatorial Pacific. By July–August (Fig. 14b), SST anomaly patterns intensify and expand spatially, and low-level wind anomalies strengthen markedly. The circulation closely resembles the JJA 850-hPa wind anomalies regressed onto the WT5 index (Fig. 10), indicating a coherent summer monsoon signal. Comparison of April–May and July–August precipitation anomalies (Figs. 14e–f) reveals a northward shift of enhanced rainfall from the eastern South Indian Ocean to the equatorial eastern Indian Ocean and the marginal seas of South Asia. Positive precipitation anomalies also extend into East Asia between 30° and 40°N and around the Tibetan Plateau. From summer to winter, negative SST anomalies over the Niño-3.4 region intensify (Figs. 14c–d), accompanied by anomalous low-level easterlies (Figs. 14c–d) and enhanced divergence over the equatorial central Pacific (Figs. 14g–h). These anomalies are associated with increased precipitation over the Philippine Sea, the Maritime Continent, and adjacent oceans. During December–January, northerly wind anomalies emerge along the East Asian coast (Figs. 14d and 14h), indicative of a strengthened EA–WP winter monsoon.

Together, these results demonstrate that strong EA–WP monsoon years are characterized by a coupled evolution of monsoon circulation and ENSO SST anomalies, extending from the pre-monsoon spring through the subsequent winter. While precursory signals appear during spring, the persistence of WT5 during the critical late-summer window (20 July–28 August) plays a key role in determining whether the monsoon circulation and ENSO SST anomalies become phase-locked, leading to a coherent strong monsoon year.

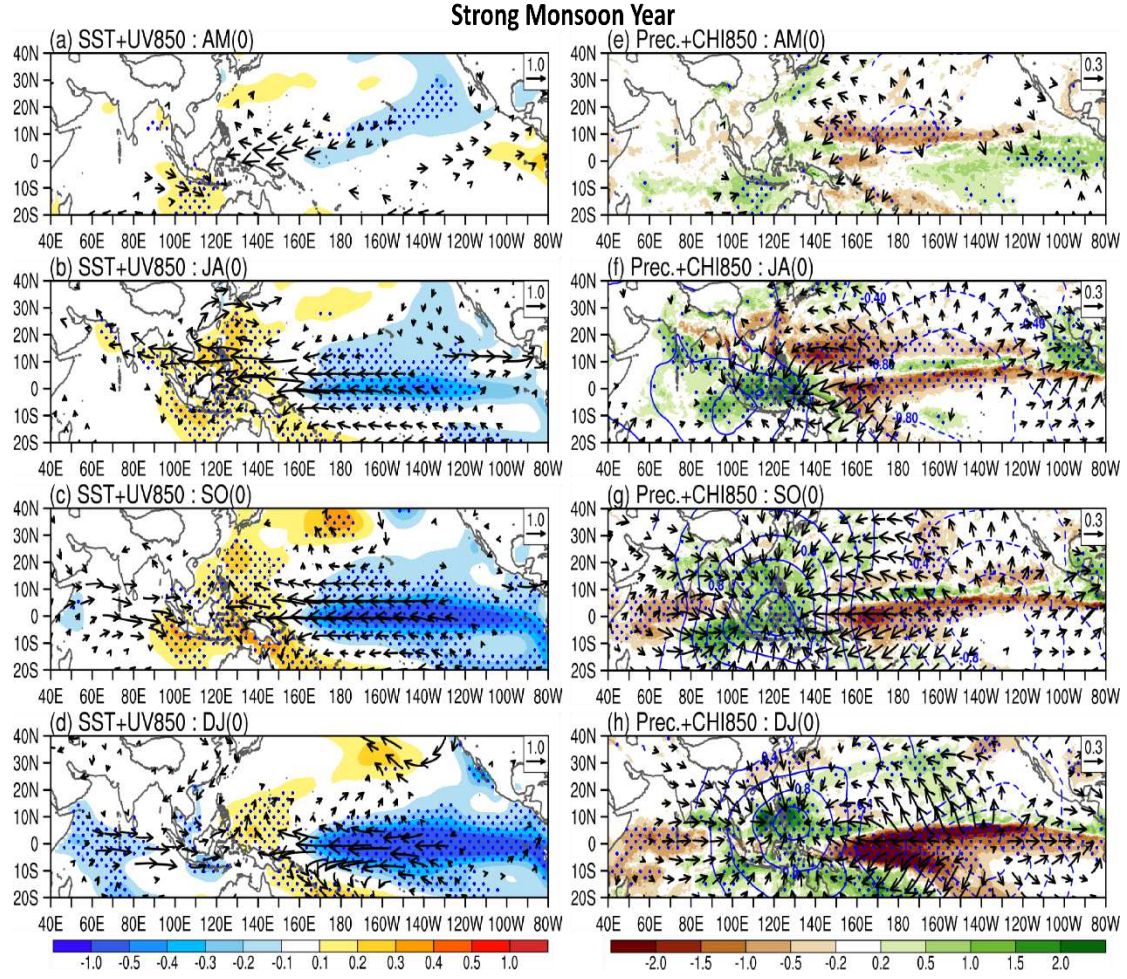


Fig. 14 Composite fields regressed onto the frequency of WT5 during pentads 41–48 for strong monsoon years: (a–d) SST (shading, °C) and 850-hPa winds (vectors, m s^{-1}), and (e–h) precipitation (shading, mm day^{-1}), 850-hPa velocity potential (contour, $\text{m}^2 \text{s}^{-1}$; $0.2 \times 10^6 \text{ m}^2 \text{s}^{-1}$ interval), and divergent winds (vectors; m s^{-1}). Panels correspond to (a, e) April–May, (b, f) July–August, (c, g) September–October, and (d, h) December–January.

6. Summary and conclusions

In this study, we constructed a 46-year (1979–2024) low-level circulation weather-type (WT) calendar using daily ERA5 850-hPa wind fields over the Asian–Australian–western Pacific monsoon region. Nine distinct WTs were identified, whose circulation and precipitation characteristics and seasonal occurrence frequencies collectively capture the annual cycle and seasonal progression of the

monsoon system. Two WT-based monsoon indices were developed: the WT9+2 index to represent East Asian winter monsoon (EAWM) intensity and the WT5 index to represent East Asian summer monsoon (EASM) intensity. Classification based on these indices reveals a pronounced asymmetry in cross-seasonal linkage: strong (weak) EAWM conditions tend to be preceded by strong (weak) EASM conditions, whereas no robust relationship is found between EAWM intensity and the following EASM. These results are consistent with earlier studies (e.g., Meehl 1987; Yasunari 1991) and highlight the importance of analyzing the EA–WP monsoon system within the framework of a monsoon year extending from 1 May to 30 April of the following year. Strong (weak) monsoon years—defined by concurrent strong (weak) EASM and EAWM—are closely associated with ENSO phase. Approximately 86% of strong monsoon years coincide with La Niña winters, while 76% of weak monsoon years coincide with El Niño winters. This strong correspondence underscores the role of air–sea coupling in shaping the EA–WP monsoon annual cycle.

A key advantage of the WT framework is its ability to identify a critical 40-day window (pentads 41–48; 20 July–28 August) during which the coupling between monsoon circulation and ENSO development is established. Regression analyses of large-scale dynamic and thermodynamic fields onto WT5 frequency during this period reveal a coherent air–sea coupling process. In strong monsoon years, enhanced convection over the eastern Indian Ocean and South Asian coastal regions during July–August shifts eastward with seasonal progression toward the Philippine Sea and western Pacific warm pool. This enhanced convection supports equatorial Pacific easterlies, promotes cooling in the Niño-3.4 region, and strengthens winter northerlies along the East Asian coast, reinforcing convection over the Maritime Continent. In contrast, weak monsoon years are characterized by suppressed convection and anomalous westerlies during July–August over the eastern Indian Ocean and South Asian coastal regions. These conditions favor eastward-propagating oceanic Kelvin waves, warm SST anomalies in the central and eastern equatorial Pacific, and the development of El Niño conditions.

Overall, the observational evidence presented here suggests that strong and weak air–sea coupling states should be considered separately in studies of the EA–WP monsoon system. The strong coupling

state is marked by coherent evolution of monsoon circulation and ENSO, leading to more persistent weather patterns and amplified monsoon variability. In contrast, the weak coupling state—represented by ENSO-neutral or unclassified monsoon years—is likely dominated by internal atmospheric variability and high-frequency disturbances. These distinctions have important implications for extended-range weather forecasting and subseasonal-to-seasonal prediction of the EA–WP monsoon.

Declarations

Competing Interests: The authors declare no competing interests.

Data Availability

ERA5 reanalysis data is downloaded from ECMWF <https://cds.climate.copernicus.eu/datasets/reanalysis-era5-pressure-levels?tab=download> (Hersbach et al., 2023). The OLR data is downloaded from NOAA <https://www.ncei.noaa.gov/products/climate-data-records/outgoing-longwave-radiation-daily> (Lee, 2011). The daily SST data is obtained from the NOAA <https://psl.noaa.gov/data/gridded/data.noaa.ersst.v5.html> (Huang et al., 2017).

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