Two Physical Mechanisms Controlling the Interannual Variability of Baiu Precipitation

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Abstract

This work examines two physical mechanisms that control the interannual variability of Baiu precipitation from late May to mid-July, using objective analysis data of atmospheric parameters including precipitation from 1979 to 2010 and of sea surface temperature (SST) from 1982 to 2010. In late June, of this period, the mean atmospheric circulation that produces Baiu precipitation is altered. In the early Baiu season (May 26–June 24) after a warm El Niño/Southern Oscillation (ENSO) event, an anomalous anticyclone appears in the western North Pacific (WNP) through the Pacific–East Asian teleconnection, which enhances Baiu precipitation. The anomalous anticyclone is maintained by the Rossby wave response to negative SST anomalies (SSTAs) in the WNP, which persist from the preceding winter until this period. On the other hand, the effect of the Indian Ocean capacitor is essential for Baiu precipitation anomalies in the late Baiu season (June 25–July 19) after a warm ENSO event. Positive SSTAs in the tropical Indian Ocean (TIO) induce a specific atmospheric response with the Matsuno–Gill pattern, forcing an anomalous anticyclone in the WNP that increases Baiu precipitation. This effect occurs only in the late Baiu season, because it is necessary for the anomalous tropospheric heating to extend northeastward from the western TIO with intraseasonal development of the Indian summer monsoon.

Thus, the physical processes producing the interannual variability of Baiu precipitation differ between the early and late Baiu seasons. Based on these differences, more detailed consideration can be given for the indicators of the interannual variability of Baiu precipitation, i.e., the SSTAs in the tropical Pacific and Indian Oceans. The detailed monitoring of SSTAs in these oceans from the preceding winter has the potential to improve predictions of the interannual variability of Baiu precipitation.

Keywords Baiu precipitation; interannual variability; ENSO; predictability

1. Introduction

In the East Asia (EA)–western North Pacific (WNP) sector, the boreal early summer is termed the Baiu (in Japanese, Meiyu in Chinese, and Changma in Korean) season. Atmospheric features in this season are characterized by a large-scale quasi-stationary front and the monsoon rainfall that are referred to as the Baiu front and Baiu precipitation, respectively. The Baiu front is located along the northwestern periphery of the North Pacific subtropical high (NPSH) in the EA–WNP sector and moves northward...
from May to August (Ninomiya and Mizuno 1987; Tanaka 1992; Murakami and Matsumoto 1994; Wang and Xu 1997; Wang and LinHo 2002). Yoshino (1965, 1966) divided the Baiu season into four stages based on the position of the Baiu front and westerly jet at 500 hPa, and he indicated how atmospheric circulation was modified around the Baiu front with the progress of each stage.

In the modification of atmospheric circulation, the Pacific–Japan (PJ) pattern (Nitta 1987; Kosaka and Nakamura 2006) has an important role inducing some aspects of the monsoonal activity. For example, this pattern is related to the end of the Baiu season and beginning of the WNP summer monsoon (WNPSM) (Nakazawa 1992; Ueda et al. 1995, 2009). Wu and Wang (2001) and Wu (2002) reported that the monsoon in the WNP progressed stepwise, which begins in the South China Sea in mid-May, extends into the southwestern Philippine Sea in mid-June, and finally moves northeastward to the northeastern Philippine Sea in late July where it represents the onset of the WNPSM. Such a stepwise progression may generate peculiar interannual variability during each stage of the WNP development.

Because Baiu precipitation is crucial for summer-time water resources and sometimes causes severe natural disasters in East Asian countries (Ogura et al. 1985; Ninomiya 2000; Ninomiya and Shibagaki 2003; Yoshizaki et al. 2000; Kato and Aranami 2005; Kato 2006), its accurate prediction is important in the region, and many researchers have investigated its interannual variability (Shen and Lau 1995; Tanaka 1997; Kawamura 1998; Chang et al. 2000a, b; Krishnan and Sugi 2001; Tomita et al. 2004, 2010, 2011; Kosaka et al. 2011; Yamaura and Tomita 2011, 2012). In particular, its relationship with the El Niño/Southern Oscillation (ENSO) has been discussed actively (Tanaka 1997; Tomita et al. 2004; Yamaura and Tomita 2012). Yamaura and Tomita (2011) reported that the interannual variability of Baiu precipitation in June could be linked to the preceding wintertime ENSO through the Pacific–East Asian (PEA) teleconnection (Wang et al. 2000; Wang and Zhang 2002). The enhancement of anticyclones in the WNP is induced through this teleconnection, and it remains until early summer owing to local air–sea feedback, which increases Baiu precipitation. Alternatively, Xie et al. (2009) suggested that the summer-time enhancement of the anticyclone in the WNP was caused by the effects of the Indian Ocean capacitor, rather than the PEA teleconnection after a warm ENSO event. It is therefore necessary to determine how these two mechanisms control the interannual variability of Baiu precipitation.

The interannual variability of Baiu precipitation is not well understood during each stage from May to July, because the Baiu front gradually moves northward in the EA–WNP sector. It is unclear whether the interannual variability holds an interannual tendency during the entire Baiu season. Furthermore, many studies considering such an interannual tendency have used monthly or seasonal mean data (Tomita et al. 2004, 2010; Yamaura and Tomita 2011, 2012). It is questionable whether this time resolution is suitable for investigating the detailed interannual tendency.

This work considers the persistence and transition of the interannual variability of Baiu precipitation during the Baiu season. We first determine if the Baiu season can be divided into several stages, each having a similar interannual tendency. We then examine how the PEA teleconnection or Indian Ocean capacitor controls the tendency in these stages. If a stage overlaps into June, the interannual tendency must be similar to that shown by Yamaura and Tomita (2011). The results of this work will help improve prediction of the interannual variability of Baiu precipitation and to mitigate the associated natural disasters.

The remainder of this paper is organized as follows. Section 2 describes the data and methodology used in this work. Section 3 presents the interannual tendencies for the stages of the Baiu season, and Section 4 examines the physical processes that form them. The predictability of the interannual variability of Baiu precipitation is discussed in Section 5. Finally, a summary of this work is presented in Section 6.

2. Data and methodology

The following three datasets were used in this work: (1) the Global Precipitation Climatology Project (GPCP) 5-day (pentad) mean data for precipitation rates, which are on global grids with 2.5° intervals (Adler et al. 2003); (2) 6-hourly data for atmospheric parameters of the Japanese 25-year Reanalysis and Japan Meteorological Agency Climate Data Assimilation System, which covers the globe at a spatial resolution of 1.25° (Onogi et al. 2007); and (3) the daily optimum interpolation data of sea surface temperature (SST) compiled by the National Oceanic and Atmospheric Administration, which are on global grids with 0.25° intervals (Reynolds et al. 2007). To remove small-scale variations, we further revised the SST data to produce a lower spatial resolution with a 1.0° interval. To unify the temporal resolution, we changed all data such that it had a pentad mean
interval, similar to the GPCP data. Hereafter, the order of pentads in a year is expressed by “P” plus a number from 1 to 73, e.g., P40. Note that the period of the first two datasets was from 1979 to 2010 (32 years), while that of the third was from 1982 to 2010 (29 years) because the SST data with a high temporal resolution such as daily interval are available only after 1982.

To divide the numerous statistical samples into a smaller number of groups with similar characteristics, we employed cluster analysis using the Ward method (Ward 1963). An application of this methodology for meteorological data has been documented in Yamaura and Tomita (2011). The data used in this cluster analysis is described in detail in the next section. To identify anomalous patterns in the oceanic and atmospheric fields associated with the interannual variations estimated through the cluster analysis, we used linear correlation and regression techniques. Anomaly indices were then deduced from the regression coefficients, and the statistical significance of the correlation coefficients was evaluated using the two-tailed Student’s t-test.

3. Interannual variation of Baiu precipitation in the early and late Baiu seasons

In climatology, Baiu precipitation and the related atmospheric circulation display different spatial patterns between the early (P30–35; May 26–June 24) and late (P36–40; June 25–July 19) Baiu seasons (Fig. 1). The details of this division will be discussed later. In Fig. 1, the vertically (1,000–300 hPa) integrated water vapor flux can be used to diagnose the linkages between precipitation and lower tropospheric circulation (Tomita et al. 2004; Yamaura and Tomita 2011).

In P30–35 (Fig. 1a), the regions with large precipitation rates extend from the southeastern part of China to the east of Japan with strong southwesternlies, i.e., Baiu precipitation. In P36–40 (Fig. 1b), the area of Baiu precipitation is elongated zonally along 36°N with slight meandering. Northward migration is observed in the western part of the Baiu front, which is part of the change in meridional circulation, concurrent with the enhancement of convective activity to the east of the Philippines (Fig. 1c). This pattern, extending from the east of the Philippines to near Japan, is similar to the PJ pattern. The enhanced convective activity to the east of the Philippines corresponds to the onset of the summer monsoon over the southwestern Philippine Sea (Wu and Wang 2001; Wu 2002).

With the change in the East Asian summer monsoon, the mean precipitation rate increases near (25°N, 80°E) over the Indian subcontinent (Fig. 1c). The intraseasonal development of the Indian summer monsoon, which is characterized by the northward shift of a large-scale monsoon rainband, creates a precipitation contrast between the Indian subcontinent and Arabian Sea. The change in the regions with latent heating of precipitation may force the northward extension of the NPSH in the WNP by modifying the zonal circulation (Enomoto et al. 2003).

It is important to consider why the Baiu season can be divided into two distinct stages, i.e., P30–35 and P36–40 (Fig. 1). To answer this question, we first estimate the Baiu season through evolution of the climatological mean precipitation rate near Japan (Fig. 2). To the north of 20°N along 110°–170°E (Fig. 2a), regions with a large precipitation rate and standard deviation slowly migrate northward from P30 to P40. The standard deviation tends to be large when the mean precipitation rate is large in the Baiu front (Tomita et al. 2004; Yamaura and Tomita 2011). Before P30, these values along 30°N probably reflected the activity of mid-latitude cyclones. After P40, the precipitation rate decreases in mid-latitudes, while the value increases along 20°N with the onset of the WNPSM (Nakazawa 1992; Ueda et al. 1995, 2009; Suzuki and Hoskins 2009). Ueda et al. (1995) confirmed that the onset of this monsoon coincides with the end of the Baiu season. Thus, the Baiu season near Japan is defined as the period P30–40, which is consistent with the estimations of many other studies (Tanaka 1992; Murakami and Matsumoto 1994; Wang and Xu 1997; Tomita et al. 2011).

The velocity of the climatological northward shift of the Baiu front is different between its western and eastern parts (Figs. 2b, c). Before P35, the center of western Baiu precipitation is located around 25°N, but after P35, it shifts northward to around 32°N (Fig. 2b). Therefore, along 25°N, the precipitation rate abruptly decreases at P35. This northward movement occurs with a specific meridional circulation change in the EA–WNP sector (Fig. 1c), which further reflects the onset of the summer monsoon over the southwestern Philippine Sea with an increase in precipitation (Figs. 1c, 2b). Kawamura and Murakami (1998) reported that an atmospheric circulation change with a PJ-like pattern appeared in mid-June and yielded a stepwise northward shift of the Baiu front. The eastern part of the Baiu front remains around 35°N almost the entire Baiu season (Fig. 2c). At P40, the precipitation rate rapidly increases around 23°N, which indicates the onset of the WNPSM and the end of the Baiu season (Ueda et al. 1995).
Fig. 1. (a) Climatological mean precipitation rate (shading; mm day$^{-1}$), interannual standard deviation (contours; mm day$^{-1}$), and vertically (1,000–300 hPa) integrated water vapor flux (vector; kg m$^{-1}$ s$^{-1}$) averaged during P30–35 (May 26–June 24). The shading scale is shown on the right and the vector scale at the bottom. The contour interval is 1 mm day$^{-1}$ from an initial value of 2 mm day$^{-1}$. The rectangle near Japan indicates the Baiu frontal region in this period. (b) Same as (a) for the period P36–40 (June 25–July 19). (c) The differences between (a) and (b) [(b) − (a)] for the precipitation rate (contours; mm day$^{-1}$) and water vapor flux (vector; kg m$^{-1}$ s$^{-1}$). The contour interval in (c) is 1 mm day$^{-1}$, and negative values are depicted by dotted contours. The zero contours have been omitted for clarity.
Fig. 2. (a) Latitude-time cross section of climatological precipitation rate (shading; mm day$^{-1}$) and interannual standard deviation (contours; mm day$^{-1}$) in 110°–170°E. The abscissa indicates time with the number of pentads. The shading scale is shown on the right. (b) Same as (a) except for the longitudinal mean in the western region (110°–140°E) and (c) in the eastern region (140°–170°E).
To diagnose whether the interannual variability of Baiu precipitation changes before and after P35 with the change in meridional circulation in the EA–WNP sector (Fig. 1), we examined the persistence of the interannual variability of Baiu precipitation during P30–40 using cluster analysis. In this analysis, the similarity was evaluated among the interannual variations of the precipitation rate in the 11 pentads from P30 to P40. The respective interannual time series were then prepared by the following three steps: (1) the areal mean of the Baiu frontal region (20°–45°N, 110°–170°E) was calculated, and it corresponded to the region where the mean precipitation rate was greater than 4 mm day$^{-1}$ with a standard deviation greater than 2 mm day$^{-1}$, (2) a three-pentad running mean was applied to the time series data from P29 to P41 in each year to eliminate the influence of synoptic or smaller scale disturbances, and (3) an interannual time series was prepared for each of the 11 pentads (P30–40). The cluster analysis was then applied to the 11 interannual time series to identify the clusters with a similar interannual tendency. It was expected that pentads located close together would have a similar interannual tendency. The results showed that the Baiu season from P30 to P40 can be divided into two clusters, i.e., P30–35 and P36–40. It is interesting that this division is consistent with the climatological stepwise northward shift of the Baiu front (Figs. 1, 2b), suggesting that the interannual tendencies are altered before and after this rapid northward shift with the change in meridional circulations in the WNP (Fig. 1c).

To diagnose the respective interannual variations of Baiu precipitation in P30–35 and P36–40, two representative interannual time series were prepared: The first was an interannual time series for the Baiu frontal region (20°–40°N, 110°–170°E; the rectangle in Fig. 1a) during P30–35, and the second was the region of (25°–45°N, 110°–170°E; the rectangle in Fig. 1b) during P36–40. When the two areal means were estimated, averaging was based only on grid points that satisfied the following two conditions: (1) the precipitation rate was greater than 4 mm day$^{-1}$, and (2) the standard deviation was greater than 2 mm day$^{-1}$.

Figure 3 shows the normalized representative interannual time series of P30–35 and P36–40. Hereafter, we refer to these two time series as the early Baiu (EB) and late Baiu (LB) indices, respectively. The correlation coefficient between the two indices was +0.15, which was statistically insignificant, and implies that the physical mechanisms controlling these two interannual variations are different.

It is known that the interannual variability of Baiu precipitation is correlated with the preceding winter-time ENSO (Yamaura and Tomita 2011). To evaluate how the two interannual variations follow the
wintertime ENSO, we first examined the correlation coefficients between five ENSO indices and EB and LB indices (Table 1). The five ENSO indices are as follows: SST in the regions of (1) NINO West, (2) NINO 4, (3) NINO 3.4, (4) NINO 3, and (5) NINO 1+2. See Table 1 for details of each region. The EB index was significantly and positively correlated with the preceding wintertime SSTs in NINO 3 (+0.37) and NINO 1+2 (+0.46). These correlation coefficients indicate that the interannual variation of the EB index is related to that of the earlier wintertime ENSO, particularly in the eastern tropical Pacific. The LB index was also significantly correlated with all five ENSO indices, i.e., NINO West (−0.46), NINO 4 (+0.42), NINO 3.4 (+0.43), NINO 3 (+0.40), and NINO 1+2 (+0.35). The absolute values of the correlation coefficients were larger with the western indices (NINO West, NINO 4, NINO 3.4) than with the eastern ones (NINO3, NINO 1+2). The interannual variations of the EB and LB indices were related to the preceding wintertime ENSO, despite the relationship between the two indices having an insignificant correlation coefficient (+0.15; Fig. 3). We will discuss this discrepancy in Sections 4 and 5.

4. Physical processes controlling the interannual variations in the early and late Baiu seasons

To examine the detailed physical processes controlling the interannual variations of the EB and LB indices (Fig. 3), correlation and regression techniques based on these two time series were applied to data for the precipitation rate and vertically integrated water vapor flux (Fig. 4). As expected, the Baiu precipitation anomalies were significantly positive near Japan in both P30–35 and P36–40. Anomalous southwesterlies to the south of Japan, which reflect the enhanced southwesterly summer monsoon, contribute to the maintenance of the positive Baiu precipitation anomalies with large water vapor transport. These features are consistent with the findings of Tomita et al. (2004) and Yamaura and Tomita (2011). In addition, there is a difference between the two panels of Fig. 4, with positive precipitation anomalies shifted northward in the western part of the Baiu front by P36–40 (Fig. 4b). This northward shift is consistent with that shown in the climatological mean precipitation rate field (Fig. 1). There are also large differences in tropical regions lower than 20°N. Negative precipitation anomalies appear in the western equatorial Pacific in P30–35 (Fig. 4a), while in P36–40 (Fig. 4b) weak but significantly positive precipitation anomalies extend zonally in the intertropical convergence zone between the equator and 10°N in the western Pacific. To the north, negative precipitation anomalies extend zonally along 20°N. These spatial patterns indicate that the spatial phase of anomalous meridional circulation is changed in the EA–WNP sector between P30–35 and P36–40.

To diagnose the differences in the anomalous meridional circulation in the EA–WNP sector, we examined latitude–height cross sections of meridional wind and vertical p-velocity at 120°–160°E for P30–35 and P36–40 (Fig. 5). In the two panels, anomalous ascent is significantly large near the latitude of the Baiu front (~30°N) across the entire troposphere and anomalous descent appears to the south. The anomalous descent corresponds to a strengthening of the anticyclone between 10°N and 20°N in P30–35 (Figs. 4a, 5a) and along 20°N in P36–40 (Figs. 4b, 5b). In P30–35 (Fig. 5a), the significant descent anomalies range over the entire troposphere near 20°N across the entire troposphere and anomalous descent appears to the south. The anomalous descent corresponds to a strengthening of the anticyclone between 20°N and 20°N in P30–35 and P36–40 (Fig. 5b). The significant descent anomalies range over the entire troposphere near 20°N in P36–40 (Fig. 5b), which suggests that the physical mechanisms producing anomalous atmospheric circulation to the south of the Baiu front are different between P30–35 and P36–40. During P36–40, another anomalous

<table>
<thead>
<tr>
<th>NINO West</th>
<th>NINO 4</th>
<th>NINO 3.4</th>
<th>NINO 3</th>
<th>NINO 1+2</th>
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<tbody>
<tr>
<td>EB index</td>
<td>−0.16</td>
<td>+0.14</td>
<td>+0.30</td>
<td>+0.37</td>
</tr>
<tr>
<td>LB index</td>
<td>−0.46</td>
<td>+0.42</td>
<td>+0.43</td>
<td>+0.40</td>
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Table 1. Correlation coefficients of the EB and LB indices with the wintertime (DJF) SSTs in the regions of NINO West (0°–15°N, 130°–150°E), NINO 4 (5°S–5°N, 160°E–150°W), NINO 3.4 (5°S–5°N, 170°–120°W), NINO 3 (5°S–5°N, 150°–90°W), and NINO 1+2 (10°S–0°, 90°–80°W). An absolute value of correlation coefficient larger than 0.32 (0.38) indicates statistical significance at the 10 % (5 %) level, where the degree of freedom is 27.
ascent appears near 5°N, which forms an anomalous meridional circulation between the equator and 20°N. In the EA–WNP sector, the anomalous meridional circulations are obviously different between P30–35 and P36–40.

These anomalous horizontal and meridional circulations yield the respective Baiu precipitation anomalies in P30–35 and P36–40 (Figs. 4, 5), which may further be associated with different processes from the preceding wintertime ENSO (Table 1). To examine the differences in air–sea interaction and its persistence, we compared SST anomalies (SSTAs) and stream function anomalies at 850 hPa between P30–35 and P36–40 (Figs. 6, 7).

Figure 6 shows SSTAs and stream function anomalies at 850 hPa respectively in P30–35 and P36–40 (Figs. 4, 5), which may further be associated with different processes from the preceding wintertime ENSO (Table 1). To examine the differences in air–sea interaction and its persistence, we compared SST anomalies (SSTAs) and stream function anomalies at 850 hPa between P30–35 and P36–40 (Figs. 6, 7).

The anomalous anticyclone in the WNP and negative SSTAs around (15°N, 150°E) are weakened by P36–40 (Fig. 6b), with the onset of the summer monsoon over the southwestern Philippine Sea. Convective activity is largely strengthened from this time onward, weakening the anticyclone to the east of the Philippines. This further switches off the positive air–sea feedback in the PEA teleconnection, increasing the climatological monsoon westerlies and cyclonic circulation (Fig. 1c). The persistence of the anomalous anticyclone in the WNP is crucial for producing the interannual variability of Baiu precipitation in P30–35.

We then examined the anomalies associated with the interannual variation of the LB index in P30–35 and P36–40 (Fig. 7). An anomalous anticyclone is weak to the east of the Philippines in P30–35 (Fig. 7a), but strong in South Asia or in the tropical Indian Ocean (TIO). Significant positive SSTAs appear in the TIO, which indicate the delayed warming of the TIO due to the effects of a tropical atmospheric bridge after a warm ENSO event (Klein et al. 1999). The positive SSTAs in the TIO remain until P36–40.
(Fig. 7b), especially in western areas including the Arabian Sea, while positive SSTAs are weakened in the central tropical Pacific. In P36–40 (Fig. 7b), an anomalous anticyclone develops along 20°N with decreasing negative SSTAs in the WNP, which is concurrent with the anomalous descent in that area (Fig. 5b). Xie et al. (2009) proposed that a specific atmospheric circulation with the Matsuno–Gill (MG) pattern (Matsuno 1966; Gill 1980) could form an anomalous anticyclone over the Philippine Sea in the summer following a warm ENSO event as a response to positive SSTAs appearing in the TIO, which is an effect of the Indian Ocean capacitor. We found that the development of the anomalous anticyclone through this effect was particularly large in P36–40. In this period, local air–sea interaction in the WNP is weak and has less influence on the formation of the anticyclone, because negative SSTAs under the anticyclone are small (Fig. 7b). The significant negative SSTAs around Japan seem to be caused by weakened radiation anomalies due to enhanced Baiu frontal activity (Tomita et al. 2010).

To determine how the effects of the Indian Ocean capacitor appear in the WNP, we again applied the correlation and regression techniques based on the EB and LB indices to precipitation rate, tropospheric (850–250 hPa) temperature, and vertically (1,000–300 hPa) integrated water vapor flux (Figs. 8, 9). Because tropical convection with precipitation adjusts the tropospheric temperature to a moist adiabatic profile, positive SSTAs could heat the tropospheric column with active convection occurring in the tropics (Emanuel et al. 1997; Neelin and Su 2005). Through this coupling between the SST and tropospheric temperature, the warming of the tropical upper ocean forces an atmospheric response with the MG pattern. Thus, the tropospheric temperature can be used to diagnose the atmospheric response forced by the effects of the Indian Ocean capacitor (Xie et al. 2009).

In the anomaly fields in P30–35 associated with the interannual variation of the EB index (Fig. 8a), positive precipitation anomalies appear in and around the Baiu front. Significant positive anomalies in the tropospheric temperature extend in parallel to the south of this active convective region, which may be induced by the combined effect of adiabatic heating with anomalous descent (Fig. 5a), latent heating of precipitation, and temperature advection. The temperature anomalies are small at latitudes lower than 20°N in the tropical Pacific and Indian Oceans, even when we consider the regions with insignificant correlations. An anomalous anticyclone appears around (15°N, 150°E) with negative precipitation anomalies in P30–35, while these anomalies

**Fig. 5.** (a) Latitude-height cross section of correlations (shading) and regressions (contours; hPa h−1) of the vertical p-velocity in 120°–160°E in P30–35 based on the EB index. Vectors are depicted by the regressions of meridional wind (m s−1) and vertical p-velocity (hPa h−1). Vertical p-velocity is multiplied by −1 to express upward motion as positive, and the unit was changed from Pa s−1 to hPa h−1 to adjust the size of the vector in the vertical direction. Shading indicates the regions with statistical significance at the 5 % level for the correlations. The shading and vector scales are shown at the bottom. (b) Same as (a) except for P36–40 based on the LB index.
are weakened in the subsequent P36–40 (Fig. 8b). There are no well-organized circulation anomalies in P36–40, except for an anomalous anticyclone around (40°N, 180°). The SSTAs and temperature anomalies are small in the TIO throughout the entire Baiu season from P30 to P40 (Figs. 6, 8), which indicates that the effects of the Indian Ocean capacitor are inactive in the anomaly fields reflected by the EB index.

The precipitation anomalies following the interannual variation of the LB index are different from those associated with the EB index (Fig. 9). In P30–35 (Fig. 9a), precipitation is anomalously suppressed in the Bay of Bengal, the South China Sea, and the region to the east of the Philippines, while it is enhanced in the western TIO. Anomalous easterlies prevail in the TIO and induce divergence off the equator owing to surface friction, leading to suppression (Xie et al. 2009). Because positive SSTAs in the western TIO (Fig. 7a) increase the tropospheric temperature through active convection, positive temperature anomalies appear in the equatorial region of the TIO and are particularly high in the western half of the region (Fig. 9a). The anomalous atmospheric heating forms a specific tropical circulation with the MG pattern in and around the Indian Ocean, which induces the anomalous anticyclone in South Asia (Fig. 7a). Note that regions with positive temperature anomalies appear in the tropical Pacific if we consider

Fig. 6. Regressions of SST (shading; K) and stream function at 850 hPa (contours; $10^5 \text{ m}^2 \text{ s}^{-1}$) based on the EB index in (a) P30–35 and (b) P36–40. Dots indicate grids with statistical significance at the 10 % level for the correlations of SST. The contour interval of the stream function is $3.0 \times 10^5 \text{ m}^2 \text{ s}^{-1}$. The shading scale is on the right.
a lower significance level in the correlations.

In the following P36–40 (Fig. 9b), negative precipitation anomalies and anomalous easterlies appear along 20°N from Southeast Asia to the international date line. Positively large temperature anomalies extend locally to the northwest of this region around (22°N, 110°E), in the western TIO and central-eastern tropical Pacific. It is known that the region with anomalously high tropospheric temperature largely extends from the TIO to the eastern tropical Pacific in the summer following a warm ENSO event (Yulaeva and Wallace 1994; Sobel et al. 2002), which is formed together with warming of the TIO (Kumar and Hoerling 1998; Lau et al. 2005; Xie et al. 2009). The local positive temperature anomalies around (22°N, 110°E) seem to be linked to the activity of the Baiu front extending northeastward.

In this period (Fig. 9b), anomalous easterlies still appear in the South Indian Ocean, while anomalous westerlies prevail over the North Indian Ocean, which implies that the Indian summer monsoon has been enhanced. The positive temperature anomalies extend simultaneously northeastward in the western TIO and reach the Indian subcontinent. These temperature and circulation anomalies are similar to those with the modified MG pattern forced by asymmetric heating for the equator (Gill 1980). This change in anomalous tropospheric heating distribution leads the eastward extension of the anomalous anticyclone along 20°N in the WNP (Fig. 7b). In addition, enhanced precipitation in the western equatorial Pacific, probably induced by outward wind from the anticyclone near the surface (Xie et al. 2009), can recursively enhance the anticyclone through the establishment of local Hadley
circulation (Fig. 5b). It is conceivable that the eastward shift of the modified MG pattern first forms a weak anomalous anticyclone along 20°N in the WNP, then the outflow from the anomalous anticyclone at the surface induces convection near the equator, and finally the locally established meridional circulation recursively enhances the anomalous anticyclone.

The northeastward shift of the regions with tropospheric heating in the TIO is key in the interannual variability of the LB index (Fig. 9). To examine the physical processes in more detail, we magnified the square region bounded by a dashed-line in Fig. 9 and considered this further (Fig. 10). In P30–35 (Fig. 10a), positive precipitation anomalies were well developed and also corresponded to the latent heating in the western TIO. The regions with anomalously high tropospheric temperature extend symmetrically to the equator. In the following P36–40 (Fig. 10b), the positive precipitation anomalies are weakened, and the center moves to the west of the Indian subconti-
Although the regions with positive precipitation anomalies are distributed locally in this domain, regions with anomalously high tropospheric temperature extend widely. This is because of the difference in the typical spatial scales of precipitation and temperature and of the large temperature advection (cf. Su et al. 2003). The positive precipitation anomalies correspond to the positive SSTAs in the western TIO and Arabian Sea (Fig. 7) and the enhanced Indian summer monsoon (Fig. 9b). Tomita and Yasunari (1996) reported that the Indian summer monsoon tended to be strong after a warm ENSO event in the biennial oscillation tendency. A northeastward shift of precipitation anomalies occurred with the intraseasonal development of the large-scale rainband of the enhanced Indian summer monsoon (Fig. 1c). The asymmetric northeastward extension of regions with anomalously high tropospheric temperatures then causes deformation of the MG pattern in the TIO and moves it eastward (Gill 1980).

5. Predictability of the two interannual variations in Baiu precipitation

The previous section discussed the two physical processes that control the interannual variations of Baiu precipitation in P30–35 and P36–40, which are both related to the preceding wintertime ENSO (Table 1). This section examines the processes leading to the
interannual variations associated with the EB and LB indices (Fig. 3) from the preceding winter to the Baiu season. That is, we attempted to extend the predictability to the preceding winter.

During December, January, and February (DJF; Fig. 11a), the spatial pattern of SSTAs based on the EB index is similar to the pattern that appears in a warm ENSO event. Positive SSTAs appear in the eastern tropical Pacific, surrounded by negative SSTAs in a horseshoe pattern, and positive SSTAs again surrounding the horseshoe pattern to the west, from the South China Sea to the Timor Sea. As an atmospheric response to these SSTAs, anomalous cyclones (anticyclones) straddle the equator in the central-eastern tropical Pacific (around the Indonesian maritime continent). This pattern is typical in the PEA teleconnection (Wang et al. 2000; Wang and Zhang 2002), which tends to remain until the following March, April, and May (MAM; Fig. 11b) with specific air–sea interaction. The positive SSTAs in the eastern tropical Pacific are weakened and retreat eastward in MAM. There appear to be no significant anomalies in the TIO from DJF to MAM.

The spatial pattern of SSTAs leading the interannual variation of the LB index is again similar to the pattern appearing in a warm ENSO event in DJF (Fig. 12a). The positive SSTAs extend to the west of the international date line from the central-eastern tropical Pacific. Significant positive SSTAs also appear in the equatorial Indian Ocean and are maintained from DJF to MAM (Fig. 12b; Klein et al. 1999; Huang and Kinter 2002; Xie et al. 2002; Tokinaga and Tanimoto 2004). In DJF (Fig. 12a), the anomalous twin cyclones and anticyclones straddling the equator again appear around 100°E and 170°W, and they are located 10°–20° west of these locations in Fig. 11a. In the following MAM (Fig. 12b), positive SSTAs are located in the central equatorial Pacific, but are weakened in the eastern tropical Pacific. The northern circulation of the anomalous twin cyclones in the tropical Pacific moves westward to around 160°E by MAM. The anomalous twin anticyclones in the TIO also persist until MAM and contribute to the maintenance of positive SSTAs in the area (Tokinaga and Tanimoto 2004). The positive SSTAs in the TIO contribute to formation of the MG pattern through tropospheric heating in the following Baiu season (Fig. 9; Xie et al. 2009).

In Fig. 12a, negative SSTAs appear in the western tropical Pacific, which can form anomalous twin anticyclones around the Indonesian maritime continent through the Rossby wave response (Lau and Nath 2000). This effect is related to the PEA teleconnection (Wang et al. 2000) and was important for the variation associated with the EB index (Figs. 6, 11). In the variation of the LB index (Fig. 12), anomalous twin anticyclones appear from the South China Sea to the Timor Sea with negative SSTAs in the western tropical Pacific. However, by the following P30–35 (Fig. 7a), twin anticyclonic anomalies extend zonally in the tropical Indian Ocean, while the negative SSTAs...
are shrinking. By the following P36–40 (Fig. 7b), anticyclonic anomalies are again strengthened along 20°N in the WNP, whereas the negative SSTAs have disappeared. This is caused by the deformation of the MG pattern following the northeastward extension of regions with an anomalously high tropospheric temperature in the TIO and by the strengthened meridional circulation with anomalous active convection along the equator –10°N band in the western tropical Pacific (Fig. 9b). In the variation of the LB index, the negative SSTAs in the western tropical Pacific partially contribute to the anomalous twin anticyclones until MAM (Fig. 12b).

The decay processes of the ENSO cause differences in the interannual variations of Baiu precipitation between P30–35 and P36–40. The appearance of positive SSTAs in the central-eastern tropical Pacific in the preceding winter, which manifests as the mature phase of a warm ENSO event, is similar between the two processes (Figs. 11a, 12a). However, teleconnection to the TIO and detailed distribution of SSTAs in the tropical Pacific in the following MAM differed between the two processes (Figs. 11b, 12b), leading to a weak and insignificant correlation between the EB and LB indices (Fig. 3). These features have the potential to provide more accurate predictions of the interannual variability in Baiu precipitation. For this prediction, three questions arise as follows: (1) How westward do the positive SSTAs extend from the central-eastern tropical Pacific in the mature phase

Fig. 11. Regressions of SST (shading) and stream function at 850 hPa (contours; $10^5$ m$^2$ s$^{-1}$) based on the EB index in the preceding (a) DJF and (b) MAM. The contour interval is $2 \times 10^5$ m$^2$ s$^{-1}$. Dots indicate grids with statistical significance at the 10 % level for the correlations with SST. A common shading scale is shown on the right.

of a warm ENSO event? (2) How long do the positive SSTAs in the central tropical Pacific remain after the warm ENSO event? (3) How long do the positive SSTAs last in the TIO, particularly in the western part? The answers to these questions would improve the prediction of interannual variability of Baiu precipitation following a warm ENSO event, as discussed in the previous section. Note that in the case of a cold ENSO event, the signs of the anomalies are reversed because the analyses are based on linear correlation and regression techniques. With the answers to these three questions, it should be possible to improve the prediction of the interannual Baiu precipitation anomalies.

Finally, the spatial pattern of SSTAs shown in Fig. 12b is reminiscent of the so-called Dateline ENSO (Larkin and Harrison 2005), El Niño Modoki (Ashok et al. 2007), or a central type of ENSO (Kao and Yu 2009). However, the correlation coefficient between the LB and El Niño Modoki indices (Ashok et al. 2007) earlier in MAM was insignificant (+0.31). Further studies are needed to determine the relationship between these ENSO classifications and the predictability of the interannual variability in Baiu precipitation proposed in this work.

6. Summary

This work investigated how the interannual variability of Baiu precipitation remained throughout the Baiu season, from late May to mid-July, using the globally gridded pentad mean data. We first identified that the interannual tendencies changed around late June near Japan, and the Baiu season was divided in half by this change. The early period was from
Fig. 13. Schematic diagram of the physical processes controlling the interannual variability of Baiu precipitation in P30–35 (May 26–June 24) for (a) DJF, (b) MAM, (c) P30–35, and (d) P36–40. Red (blue) shading indicates the region with positive (negative) SSTAs. The circular vector labeled L (H) denotes a specific anomalous cyclone (anticyclone). The green vector depicts the enhanced flow of the vertically integrated water vapor flux near the Baiu front. The bold thick line near Japan indicates the Baiu front.
Fig. 14. Same as Figure 13 for the physical processes controlling the interannual variability of Baiu precipitation in P36–40 (June 25–July 19). The thin lines in the TIO and near the Indonesian maritime continent denote regions near the eastern edge of the MG pattern from the TIO. Black straight vectors depict anomalous surface wind. The region with red hatching indicates anomalous tropospheric heating. White vectors denote anomalous meridional circulations.
May to 24 June (P30–35), and the late period was from 25 June to 19 July (P36–40). This division is closely related to the onset of the summer monsoon over the southwestern Philippine Sea with a specific meridional circulation referred to as the PJ pattern (Nitta 1987; Kosaka and Nakamura 2006). The western part of the Baiu front climatologically shifts northward with the seasonal establishment of the PJ pattern (Fig. 1). The northward shift of the large-scale rainband of the Indian summer monsoon correspondingly occurs in the TIO (Fig. 1c) and may contribute to the northward migration of the Baiu front through the Rossby wave response (Enomoto et al. 2003).

The physical processes controlling the interannual variability of Baiu precipitation are different before and after this shift. Figure 13 schematically shows the processes generating interannual variation in the early Baiu season, P30–35. In the preceding winter (Fig. 13a), positive SSTAs occur in the central-eastern tropical Pacific, while negative SSTAs extend to the west or northwest, and positive SSTAs are again amplified to the west in the South China Sea. An anomalous cyclone appears to the northwest of the eastern positive SSTAs, while the counterpart anomalous anticyclone is forced between the negative SSTAs and western positive SSTAs. This pattern is just part of the PEA teleconnection with a warm ENSO event (Wang et al. 2000; Wang and Zhang 2002). In the following spring (Fig. 13b), the anomalous anticyclone persists in the WNP as the Rossby wave response for the negative SSTAs, while the eastern positive SSTAs are diminished. The anomalous anticyclone intensifies the NPSH in the WNP and maintains the negative SSTAs through a local air–sea feedback until P30–35 (Fig. 13c). In the western part of this anomalous anticyclone, westerlies are strengthened in the lower troposphere and increase the Baiu precipitation. With the onset of the monsoon over the southwestern Philippine Sea, the positive air–sea feedback maintaining the anomalous anticyclone in the WNP is switched off (Fig. 13d).

Figure 14 shows the physical processes producing interannual variation in the late Baiu season, P36–40. In the preceding winter (Fig. 14a), positive (negative) SSTAs develop in the central-eastern tropical Pacific (WNP). In comparison with the distributions shown in Fig. 13a, the region with positive SSTAs extends more westward, and an anomalous cyclone is correspondingly developed in the northwestern part of the region. Significant positive SSTAs do not occur in or around the South China Sea, but they do develop in the TIO through the effects of the tropical atmospheric bridge (Klein et al. 1999). The anomalous anticyclone extends zonally from the northeastern TIO to the South China Sea, and it is partially forced by the negative SSTAs in the WNP. In the following spring (Fig. 14b), positive SSTAs persist in the central tropical Pacific with an anomalous cyclone in the northwestern part of the region, which contributes to the maintenance of negative SSTAs in the WNP. These anomalies diminish by the end of spring, with a seasonal development of the NPSH (Fig. 14c). Conversely, the positive SSTAs in the TIO persist until summer (Xie et al. 2009), particularly in the western part with an anomalous anticyclone in the northeast. With the intraseasonal evolution of the Indian summer monsoon, the anomalous tropospheric heating moves northeastward from the western TIO by P36–40 (Fig. 14d). The MG pattern is then modified and moves eastward, and an anomalous anticyclone appears in the WNP. The outflow near the surface from the anomalous anticyclone to the equator is able to enhance the local meridional circulation in the WNP (Xie et al. 2009), which recursively reinforces the anomalous anticyclone. The anomalous anticyclone leads southwesterlies in the northwestern part and activates the Baiu precipitation.

Understanding these two processes could improve the predictability of the interannual variability of Baiu precipitation after warm ENSO events. The following three questions are crucial: (1) How westward do the positive SSTAs extend from the central-eastern tropical Pacific in the warm ENSO winter? (2) How long do the positive SSTAs remain there after the winter? (3) How do the positive SSTAs stay in the TIO? In the case of a cold ENSO event, the anomalies have opposite signs. It is proposed that more detailed monitoring of SSTAs from winter to early summer in the tropical Pacific and Indian Oceans would improve the prediction of the interannual Baiu precipitation anomalies. This work should initiate further discussions about the more detailed and accurate prediction of the interannual variability of Baiu precipitation.

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