Interannual Variability of the Indonesian Rainfall and Air–Sea Interaction over the Indo–Pacific Associated with Interdecadal Pacific Oscillation Phases in the Dry Season

Murni Ngestu NUR’UTAMI and Tadahiro HAYASAKA

Graduate School of Sciences, Tohoku University, Sendai, Japan

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Abstract

The interannual and interdecadal variabilities of Indonesian rainfall in dry seasons (June–November) are investigated by using rainfall data from the Climate Research Unit (CRU) from 1939 to 2016 and from the Global Precipitation Climatology Project (GPCP) from 1979 to 2016. The first principal component (PC1) of both the CRU and GPCP data shows that the canonical El Niño–Southern Oscillation (ENSO), ENSO Modoki, and Indian Ocean Dipole (IOD) are major climate modes influencing the interannual variability of rainfall in Indonesia, and the Interdecadal Pacific Oscillation (IPO) is the major decadal phenomenon affecting the decadal variability of the rainfall. Furthermore, the IPO modulates the influence of IOD on Indonesian rainfall, with a weaker influence during the positive IPO phase during 1979–1997 and a stronger influence during the negative IPO phases during 1939–1978 and 1998–2016. The dependency of Indonesian rainfall response to the canonical ENSO and ENSO Modoki on IPO phases is not significant, although the response to the ENSO Modoki (canonical ENSO) becomes significant (less significant) in the negative IPO phase during 1998–2016 when compared with earlier periods.

Keywords Indonesian rainfall; El Niño–Southern Oscillation; ENSO Modoki; Indian Ocean Dipole; Interdecadal Pacific Oscillation; interannual variability

1. Introduction

Indonesia has two seasons, a dry and a wet season, the seasonal evolution of which is influenced by the Asian–Australian monsoon system (Qian et al. 2002; Hung et al. 2004). Typically, the dry season is from July to September, and the wet season is from November to April (Alsepan and Minobe 2020). Indonesian rainfall variability is strongly linked to climate variability over the Indo–Pacific Ocean, namely, the canonical El Niño–Southern Oscillation (ENSO; Trenberth 1997; Trenberth 2019; Wang et al. 2017; Nur’utami and Hidayat 2016; Alsepan and Minobe 2020), the ENSO Modoki (Ashok et al. 2007; Alsepan and Minobe 2020), and the Indian Ocean Dipole (IOD; Saji et al. 1999; Li et al. 2003; Nur’utami and Hidayat 2016; Alsepan and Minobe 2020) on interannual timescales, and the Interdecadal Pacific Oscillation (IPO; Salinger et al. 2001; Dong and Dai 2015) on multidecadal timescales. Additionally, there may be remote influences from climate variabilities over the Atlantic Ocean on interannual and decadal timescales (Trenberth and...
Indonesian rainfall anomalies influenced by the interannual climate modes are large during the dry season and the transition period from the dry-to-wet season (June–November; JJASON) (Hendon 2003; Yanto et al. 2016; Alsepan and Minobe 2020). During canonical El Niño, positive anomalies of sea surface temperature (SST) occur over the central to eastern tropical Pacific Ocean and weaken the southeast trade winds. These conditions drive the movement of the upward branch of the Walker circulation from the western to the eastern equatorial Pacific, bringing water vapor to the area of high SST (i.e., the central to the eastern Pacific Ocean; Bjerknes 1969), and then reduce rainfall over Indonesia (Ropelewski and Halpert 1987; Hendon 2003; Yanto et al. 2016; Alsepan and Minobe 2020). El Niño Modoki, characterized by positive SST anomalies (SSTAs) over the central Pacific Ocean and negative SSTAs in the western and eastern parts, also reduces Indonesian rainfall. However, the regionality of rainfall anomalies caused by El Niño Modoki has some differences when compared with canonical El Niño. For example, El Niño Modoki reduces rainfall over southern Indonesia more substantially than canonical El Niño (Weng et al. 2007; Alsepan and Minobe 2020). During the positive phase of IOD, large negative SSTAs occur in the tropical southeastern Indian Ocean from July to September, decreasing rainfall over southwestern Indonesia (As–Syakur et al. 2014; Alsepan and Minobe 2020).

The influence of these climate modes should be examined with caution because of their interdependency (canonical ENSO–IOD, Ashok et al. 2003; Nur’utami and Hidayat 2016; canonical ENSO–El Niño Modoki, Weng et al. 2007; IOD–El Niño Modoki, Feng and Chen 2014). Generally, to analyze the relationship between rainfall and specific interannual climate variability by eliminating the influence of other climate variabilities that might affect the correlation coefficient, the partial correlation method is used (Ashok and Saji 2007; Weng et al. 2007; Taschetto and Matthew 2009; Cai et al. 2011; As–Syakur et al. 2014; Alsepan and Minobe 2020). For regional-scale Indonesian rainfall, Alsepan and Minobe (2020) described the seasonal migration of areas with rainfall anomalies associated with each of the interannual climate modes using field-based observational rainfall data. However, their analysis period was limited to 1960–2007. Data from 2007 onward should be incorporated into the analysis because ENSO Modoki, for example, has become more frequent and persistent during recent decades (Ashok et al. 2007).

On decadal and interdecadal timescales, there are several types of Pacific modes, including the Pacific Decadal Oscillation (PDO; Mantua et al. 1997; Newman et al. 2016) and the IPO (Salinger et al. 2001; Dong and Dai 2015). The spatial patterns of SSTAs are similar in both climate modes. Although the PDO pattern exhibits negative SSTAs extending from Japan into the central North Pacific and a horseshoe of positive SSTAs along the North American continent, the IPO pattern spans the entire Pacific basin and includes a southern hemispheric component of the PDO. These long-term variabilities lead to changes in the physical and biological environments in the surrounding area (Hare and Mantua 2000; Mantua 2004). Transitions in long-term variabilities from one phase to another or changes in a phase progression trend are called climatic regime shifts (Minobe 1997). Several studies have analyzed the impact of the IPO not only on changes in the magnitude and frequency of the canonical ENSO but also on rainfall (Power et al. 1999; Salinger et al. 2001; Chew and Leahy 2003; Dong and Dai 2015), temperature, sea surface pressure (Salinger et al. 2001), the Intertropical Convergence Zone (Salinger et al. 2001), the South Pacific Convergence Zone (Folland et al. 2002), and tropical cyclones (Li et al. 2015; Magee et al. 2017). The second principal component (PC2) of Indonesian rainfall at decadal timescales from 1950 to 1980 is related to the PDO, although there is also an IPO signal in the leading principal component (PC) of Indonesian rainfall post-1980 (Yanto et al. 2016). Additionally, simulation results suggest that the IPO also causes changes in the Indian Ocean SST (Dong et al. 2016). These results showed that the decadal variability in Indian Ocean SST is dominated by internal variability that is induced by the tropical Eastern Pacific Ocean SST (especially the IPO).

The IPO significantly modulates the effects of ENSO on rainfall in many regions, where IPO plays the dominant decadal variability (Dong and Dai 2015). For example, the Australian rainfall variability is more significantly correlated with canonical ENSO during the negative IPO phase than during the positive IPO phase (Power et al. 1999; Dong and Dai 2015). However, the relationship between regional-scale Indonesian rainfall and interannual climate modes during different IPO phases is not well understood. Hence, this study aims to improve our understanding of the variability in Indonesian rainfall associated with the interannual climate modes over the tropical Indo–Pacific and its modulation by interdecadal climate
mode (IPO).

This paper is organized as follows. Section 2 presents the data, indices, and method. Section 3 documents the analysis of the Indonesian rainfall variability on multiple timescales (interannual and interdecadal) in two types of rainfall data, including the analysis of the effect of IPO on Indonesian rainfall and the IPO’s modulation of the Indonesian rainfall response to canonical ENSO, ENSO Modoki, and IOD. Section 4 presents the summary and discussion.

2. Data, indices, and method

2.1 Data

We used the Climate Research Unit (CRU) Time Series v.4.01 monthly rainfall dataset from January 1939 to December 2016 with a horizontal resolution of 0.5° × 0.5° (Harris et al. 2014). This dataset is derived from the archives of global climate station records over land areas that have been subject to extensive manual and semiautomated quality control measures (Harris et al. 2014). This study covers the Indonesian region spanning 95.25–141.25°E and 10.75°S–6.25°N (Fig. 1) and focuses on the season in JJASON, which includes the dry and dry-to-wet transition periods. We also used the Global Precipitation Climatology Project (GPCP) monthly precipitation dataset (version 2.3; Adler et al. 2003) for the period from 1979 to 2016. The GPCP is part of the World Climate Research Program and the associated activity of the Global Water Cycle and Energy Experiment. The GPCP dataset covers both land and ocean with a horizontal resolution of 2.5° × 2.5° latitude–longitude and is merged analysis from the following input datasets: low–orbit satellite microwave data, geosynchronous–orbit satellite infrared data, and surface rain–gauge observations (Adler et al. 2003). We can compare these two rainfall datasets and the different characteristics of their input methods should provide a robust result.

The Extended Reanalysis SST version 5 (ERSSTv5; Huang et al. 2017) is used to represent the global SST and to calculate the indices of climate modes (further explanation is given in Section 2.2) for the period of 1939–2016. This product is the revised ERSST from version 4 to version 5. The revised results provide output data that show more realistic spatiotemporal variations and gives a better representation of high-latitude SSTs; additionally, ship SST biases are now calculated relative to more accurate buoy measurements, although the global long-term trend remains about the same (Huang et al. 2017).

To depict large-scale atmospheric circulation, we use the Japanese 55–year Reanalysis (JRA-55; Kobayashi et al. 2015) to obtain the horizontal wind field at 850 hPa. The JRA-55 is the second Japanese global atmospheric reanalysis product provided by the Japan Meteorological Agency. It covers the global domain with a resolution of 1.25° × 1.25° in 37 isobaric surfaces from January 1958 to December 2016, and the available time steps are sub-daily and monthly (Harada et al. 2016). The JRA-55 is the first comprehensive reanalysis that covers the last half-century and applies a four-dimensional variational assimilation technique (Kobayashi et al. 2015).

2.2 Indices

The climate indices used to represent the interannual and interdecadal variabilities are constructed from the ERSSTv5 dataset for the JJASON mean in each year. The indices are described below.

a. NINO3 index

The NINO3 index is defined by SSTA relative to the 1981–2010 mean averaged over 5°S–5°N and 150–90°W (Niño 3 region). El Niño (La Niña) is usually defined as a phenomenon in the equatorial Pacific Ocean characterized by five consecutive 3 month running means of SSTA in the Niño 3 region that are above (below) the threshold of +0.5°C (−0.5°C) (Trenberth 1997). Nevertheless, in this study, SSTA averaged over the JJASON season is used to calculate NINO3. El Niño (La Niña) identified in this study has the NINO3 index above (below) the threshold of +1σ (−1σ), where σ is the seasonal standard deviation for JJASON (Hanley et al. 2003).

b. El Niño Modoki index

The El Niño Modoki index (EMI) is defined as the difference in SSTA in the central equatorial Pacific (SSTA:C: 165°E–140°W, 10°S–5°N), the eastern equatorial Pacific (SSTA:E: 110–70°W, 15°S–5°N), and

![Fig. 1. Map of the study area in the Indonesian archipelago.](image-url)
the western equatorial Pacific ($SSTA_w$: 125–145°E, 10°S–20°N).

$$EMI = SSTA_c - (0.5 \times SSTA_w) - (0.5 \times SSTA_e).$$ (1)

The threshold is +0.7σ (−0.7σ) for El Niño Modoki (La Niña Modoki) (Ashok et al. 2007), where σ is the seasonal standard deviation for JJASON.

c. IOD mode index

The IOD mode index (DMI) is the difference between $SSTA$ in the western Indian Ocean (50–70°E and 10°S–10°N) and that in the southeastern Indian Ocean (90–110°E and 10°S–0°N) (Saji et al. 1999). The threshold is +0.75σ (−0.75σ) for the positive IOD (negative IOD) (Cai et al. 2013).

d. Tripole Index–IPO

The Tripole Index (TPI)–IPO is based on the difference between $SSTA$ in the eastern equatorial Pacific ($SSTA_{A1}$: 10°S–10°N, 170°W–90°W), and the $SSTA$ averaged over the central–west North Pacific ($SSTA_{A2}$: 25–45°N, 140°E–145°W) and the central–west South Pacific ($SSTA_{A3}$: 50–15°S, 150°E–160°W) (Henley et al. 2015).

$$TPI = SSTA_{A1} - (0.5 \times SSTA_{A2}) - (0.5 \times SSTA_{A3}).$$ (2)

We performed a 3 year moving averaging on the original time series index to eliminate high-frequency variations and then applied a 9 year moving averaging to remove interannual variation and most decadal (<20 years) variation from interdecadal changes (Dong and Dai 2015). The threshold is the zero-line from the smoothed result (Dong and Dai 2015).

2.3 Method

The time period of data used to analyze interannual and decadal variability of Indonesian rainfall is 1939–2016. The period for which the climatological mean is defined, which is used to calculate anomalies of each variable, is 1981–2010. In this study, we do not apply a filtering process to remove the global warming trend because there is no evidence of ENSO trends in the seasonal and temporal behavior through a period of strong global warming (1958–2007) (Nicholls 2008). Furthermore, there is a trend toward more El Niño-like behavior in March–September and not in November–February, the peak season of canonical El Niño and La Niña events (Nicholls 2008).

We use empirical orthogonal function (EOF) analysis to obtain a measure of the importance of each dominant pattern and to find a new set of variables that capture most of the observed variance from the data through linear combinations of the original variables. The EOF method is commonly used in most meteorological centers to compare observations and reanalysis with climate model simulations. The EOF equations that are used in this analysis follow Hannachi et al. (2007). Furthermore, Mantua (2004) provided a deep review of methods for detecting regime shift, one of which uses EOF analysis.

The standard and partial correlation techniques are used to compare and quantify the linear relationship between two or more time series datasets. The partial correlation is used to determine the specific contributions of each interannual climate mode. Equations (3)–(5) show Pearson’s correlation and two types of partial correlation formula based on the influence of several variables.

$$r_{12} = \frac{\sum_{i=1}^{n}(x_i - \bar{x})(y_i - \bar{y})}{\sqrt{\sum_{i=1}^{n}(x_i - \bar{x})^2 \sum_{i=1}^{n}(y_i - \bar{y})^2}},$$ (3)

$$r_{12,3} = \frac{r_{12} - r_{13}r_{23}}{\sqrt{(1 - r_{13}^2)(1 - r_{23}^2)}},$$ (4)

$$r_{12,34} = \frac{r_{12,4} - r_{13,4}r_{23,4}}{\sqrt{(1 - r_{13,4}^2)(1 - r_{23,4}^2)}} = \frac{r_{12,3} - r_{14,3}r_{24,3}}{\sqrt{(1 - r_{14,3}^2)(1 - r_{24,3}^2)}}.$$ (5)

Here, $x$ and $y$ are variables 1 and 2, respectively; the bar sign indicates the mean of the data, and $r_{12}$ is the correlation coefficient between the two variables ($x$ and $y$). $r_{12,3}$ is the correlation coefficient between variables 1 and 2 after removing the influence of variable 3, and $r_{12,34}$ is the correlation coefficient after removing the influence of variables 3 and 4.

The partial correlation method is implemented to obtain the independent contributions of each interannual climate mode, namely, canonical ENSO, ENSO Modoki, and IOD, to Indonesian rainfall. Furthermore, the relation between IPOs and Indonesian rainfall is analyzed in two ways. The first involves the direct application of linear correlation analysis between the TPI index and a 9 year running average of a 3 year running average Indonesian rainfall. The second involves the analysis of the IPO modulating effect on Indonesian rainfall response to the interannual climate modes.

3. Results

3.1 Comparison of rainfall data on interannual and interdecadal timescales

Figure 2 shows the spatial pattern of the first two
EOF modes (EOF1 and EOF2) of CRU and GPCP data for the 1979–2016 period. The rainfall variabilities in the two datasets are consistent with each other. The first EOF mode of the rainfall anomalies captures half of the total variance, highlighting reduced rainfall over the Indonesian region, except the northern part of Sumatra (Figs. 2a, b for EOF1). In the EOF2 mode, rainfall anomalies in both datasets are not prominent and are cluttered, especially over land (Figs. 2a, b for EOF2).

To understand the main features of EOF1 for both the CRU and GPCP data further, we show the composites of rainfall, SST, and horizontal wind anomalies over the tropical Indo–Pacific Ocean based on the positive and negative phase years of the time series of the EOF1, or first principal component (PC1), for each rainfall dataset (Fig. 3). The positive and negative rainfall anomalies in the CRU and GPCP data (Figs. 3a, b) are generally consistent with their corresponding EOF1 spatial patterns (Figs. 2a, b for EOF1). During the positive phase years of EOF1, the increase in rainfall seems to be caused by canonical La Niña, La Niña Modoki, and negative IOD (Fig. 3c for positive years). Large negative SSTA occur in the tropical central to the eastern Pacific Ocean corresponding to the canonical La Niña and La Niña Modoki patterns, and these SSTA are accompanied by significant positive SSTA in the tropical western Pacific. This SST gradient between the Indonesian seas and central to the eastern Pacific Ocean induces anomalous easterly winds, which bring the mass of water vapor from the Pacific Ocean to the Indonesian region. In the tropical Indian Ocean region, the SST gradient shows a negative IOD, which is coupled with strong anomalous northwesterly winds from the western tropical Indian Ocean and brings a mass of water vapor to the Indonesian region. These phenomena generate a convergence zone over Indonesia. The opposite conditions occur during the negative phase years of EOF1, in which the rainfall reduction is caused by the canonical El Niño, positive IOD, and El Niño Modoki patterns (Fig. 3c for negative years). We also performed a correlation analysis between the time series PCs of both rainfall datasets and SST–wind anomalies. Figure 4 shows that the spatial distribution of the correlation coefficients of SST and horizontal wind with sign-reversed PC1 of the CRU data (Fig. 4a), which resembled that with PC1 of the GPCP data (Fig. 4b), and it is similar to the composites of SSTA and horizontal wind anomalies for positive and nega-
Composite of anomalous years in the Indonesian rainfall PC1 on rainfall, SST, and wind

Fig. 3. Composite of Indonesian rainfall of (a) CRU and (b) GPCP data in JJASON based on the positive and negative years for PC1 in the period 1979–2016, and (c) composite of SST and wind anomalies over the Indian–Pacific Ocean based on the same positive and negative years for PC1 of the CRU and GPCP data. The threshold is $+1\sigma$ ($-1\sigma$) for the positive (negative) years. The diagonal cross-hatched areas and wind vectors show that the composite means are significant at the 90% confidence level.

Consequently, this shows that the two types of rainfall data consistently show the climate variability in the Indian–Pacific Ocean, which affects Indonesian rainfall.

For the correlation coefficients of SST and horizontal wind with PC2 of the CRU and GPCP data, the spatial patterns are different from each other (Fig. S1). The correlation with PC2 of the CRU data generally shows no significant anomalies of SST and wind over the whole of the tropics (Fig. S1a), but the PC2 of the GPCP data shows a Ningaloo Niño-like pattern (Doi et al. 2013; Zhang et al. 2018), in which positive SSTA and the associated wind anomalies occur over the northwest of Australia (Fig. S1b).

We calculate the partial correlation of the CRU and GPCP data with each climate mode index during...
1979–2016 to investigate the independent influences of each climate mode on Indonesian rainfall during the JJASON season in more detail (Fig. 5). Generally, the spatial patterns of partial correlation between the CRU and GPCP data are similar for each climate mode. For canonical ENSO, negative partial correlations of rainfall occur over central to eastern Indonesia (Borneo, Sulawesi, and western Papua) for both datasets (Figs. 5a, b for NINO3) with incoherent partial correlations over western Indonesia (Sumatra) and over southern Indonesia (Java, Bali, and Nusa Tenggara). For ENSO Modoki, the negative partial correlations also occur over central–eastern Indonesia (Figs. 5a, b for EMI) similar to those for canonical ENSO (Figs. 5a, b for NINO3). However, an interesting difference is that the influence of ENSO Modoki over southern Indonesia is more robust compared with that of canonical ENSO. For IOD, the negative partial correlations of precipitation for both datasets occur over southwestern Indonesia, especially over southern Sumatra and Java (Figs. 5a, b for DMI).

Figure 6 compares the PC1 time series for CRU and GPCP rainfall and the partial correlation between rainfall PCs and climate mode indices during 1979–2016. The distribution of PC1 is similar for both rainfall datasets (Fig. 6a), despite differences in their input and characteristic data. Compared with the climate mode indices, PC1 of both rainfall datasets is significantly correlated with each climate mode index (Figs. 6b, d–f), except for the correlation between PC1 of the GPCP data and IPO, with a significant correlation between IPO and PC2 of the GPCP data instead (Fig. 6c). On the interannual timescale, the correlations between both the rainfall PCs and the three climate mode indices are significant and high (Figs. 6d–f). On the interdecadal timescale, Indonesian rainfall is related to the IPO, with correlation coefficients of −0.76 and −0.86 for the 9 year running mean of PC1 of the CRU data and PC2 of the GPCP data with the TPI–IPO (Figs. 6b, c), respectively, which may explain why the Indonesian rainfall decreases during the positive IPO phase and increases during the negative IPO phase.

Although the PC1 time series of both the CRU and GPCP data have high correlations (Fig. 6a) with the interannual climate mode indices, the 9 year running mean of PC1 of the GPCP data has no significant correlations with the TPI–IPO, in contrast to the significant correlations for the 9-year running mean of PC1 of the CRU data and PC2 of the GPCP data. To understand why PC2 of the GPCP data is closely related to decadal variation rather than PC1, we perform spectrum analysis on the PC1 and PC2 time series and the climate mode indices (Fig. S2). The spectrum analysis indicates that the peak signal for PC2 of
the GPCP data is at the 5 year oscillation (Fig. S2d), which is likely related to the Ningaloo Niño phenomenon, as discussed earlier (Fig. S1b). Tanuma and Tozuka (2020) suggested that the occurrence of the Ningaloo Niño is intense and more frequent during the IPO phase associated with the positive interdecadal SSTA to the northwest of Australia that promotes deep atmospheric convection and cyclonic anomalies off the west coast of Australia through a Matsuno–Gill response. The strong link between the Ningaloo Niño, which is captured by PC2 of the GPCP data, and IPO may be a reason why PC2 of the GPCP data has a significant correlation with the IPO.

Comparing the CRU and GPCP data in the Indo-nesian region shows similar results both spatially and temporally for the interannual timescale using EOF analysis (Figs. 2, 3), correlation analysis (Fig. 6a), and spectrum analysis (Figs. S2a, c). Additionally, the spatial pattern of the correlation coefficient of the SST and wind anomalies with PC1 of the CRU and GPCP data shows a similar pattern of interannual climate modes in the Indian Ocean and the Pacific Ocean (Fig. 4). The area affected by each interannual climate mode is the same for both rainfall datasets, as indicated by the spatial patterns of the partial correlations between rainfall and each interannual climate index (Fig. 5). However, the interdecadal rainfall variability of the CRU and GPCP data is shown in the different PC mode, where the interdecadal variability corresponding to the IPO are found in PC1 of the CRU data and PC2 of the GPCP data.
3.2 Identification of interannual and interdecadal climate modes

Figure 7 compares the PC1 time series of the CRU and GPCP datasets with the climate mode indices. On the interdecadal timescale (Fig. 7a), the variability of Indonesian rainfall is strongly correlated with the IPO index. Between 1939 and 2016, there are two negative phases of IPO during 1939–1978 and 1998–2016 and a positive phase in 1979–1997. The correlation coefficient between Indonesian rainfall and the IPO index is insignificant before the 1940s ($r = 0.04$; figure not shown); hence, this study excludes the period before 1939. The correlations between the PC1 time series of Indonesian rainfall for the CRU and GPCP data with each interannual climate mode index are shown in Figs. 7b–d. The PC time series have high and significant correlations with the climate indices.

Table 1 shows the years of canonical ENSO, ENSO Modoki, and IOD (Figs. 7b–d). From 1939 to 2016, 20 combinations of interannual climate mode phases occurred. Because there are cases in which canonical ENSO and ENSO Modoki occurred simultaneously, there is still uncertainty in the effects of the combination of these climate modes on Indonesian rainfall. These results also show that canonical ENSO events do not always occur together with ENSO Modoki events. Thus, the combination of interannual climate modes on Indonesian rainfall variability should be considered.

We use the composite analysis of CRU rainfall, SST, and horizontal wind anomalies for five combination events to examine the effect of simultaneous interannual climate mode phases (Fig. 8). The five combination events are El Niño Modoki (Fig. 8a), canonical El Niño coinciding with El Niño Modoki and positive IOD (Fig. 8b), canonical La Niña (Fig. 8c), canonical La Niña coinciding with La Niña Modoki (Fig. 8d), and canonical La Niña coinciding with negative IOD (Fig. 8e). The analysis of these five climate events compares the effects of a phenomenon alone with those of multiple climate modes. Decreases and increases in rainfall caused by the interannual climate phases are clearly visible and are consistent with the area shown in the partial correlation in Fig. 5.
Fig. 7. (a) Time series of normalized PC1 of the CRU data (light blue bars), PC2 of the GPCP data (dark blue bars), and normalized TPI–IPO for the 9 year running mean in JJASON, and (b–d) comparison of the time series of PC1 of the CRU data (light blue bars) and PC1 of the GPCP data (dark blue bars) and (b) NINO3, (c) EMI, (d) DMI in JJASON. The horizontal solid lines in (b–d) show the thresholds used to define the positive or negative phase of each climate mode, and $r$ is the correlation coefficient between the PCs and climate mode indices.

Table 1. Years of the three interannual climate modes in the period 1939–2016.

|                | El Niño Modoki | Normal | La Niña Modoki |
|----------------|---------------|--------|----------------|
|                | Pos. IOD | Normal | Neg. IOD | Pos. IOD | Normal | Neg. IOD | Pos. IOD | Normal | Neg. IOD |
| El Niño        | 1991, 1997, 2006 | 1965, 1987, 2002, 2009 | – | 1946, 1972, 1982, 2015 | 1957, 2014 | – | – | 1976 | – |
| Normal         | 1963, 1986, 1994 | 1940, 1941, 1966, 1977, 2003, 2004 | 1958, 1990, 1992, 2001 | 2012 | 1944, 1951, 1952, 1953, 1968, 1969, 1979, 1980, 1993, 2000, 2013 | 1959, 1960, 1981, 1989, 2005 | 1945, 1983, 2008 | 1939, 1943 | 2016 |
| La Niña        | 1967 | – | – | 1949, 1961, 2007 | 1947, 1948, 1950, 1955, 1962, 1970, 1971, 1978, 1984, 1985 | 1942, 1954, 1956, 1995, 1996 | – | 1964, 1973, 1974, 1988, 1999, 2011 | 1975, 1998, 2010 |
for each single event, especially for El Niño Modoki and canonical La Niña (left-hand side of Figs. 8a, c). Based on this comparison of the composite analysis and the partial correlation analysis, we expect that the responses of Indonesian rainfall to each climate mode are linear during the positive and negative phases. The SSTA composites can also show the pattern of the single events (right-hand side of Figs. 8a, c). Additionally, if we compare the rainfall anomaly for a single event with that of two or three coinciding events, the rainfall anomaly is larger when the interannual climate modes occur simultaneously. For example, the reduced rainfall in Fig. 8b, when canonical El Niño coincides with El Niño Modoki and positive IOD, is more robust than the single El Niño Modoki in Fig. 8a. Another interesting example is that when canonical La Niña occurs along with negative IOD, the increase in rainfall is larger over western and southwestern
Indonesia (a large region of Sumatra; Fig. 8e) than for single canonical La Niña (Fig. 8c) and canonical La Niña coinciding with La Niña Modoki (Fig. 8d). These results confirm that increased rainfall over Sumatra may be attributed to the large contribution of the negative IOD.

3.3 Modulation by IPO of the effects of interannual climate modes

In this subsection, we analyze the impact of the IPO on Indonesian rainfall and the modulation of the Indonesian rainfall response to three interannual climate modes due to IPO. Table 2 shows the occurrence frequency of interannual climate variability for each IPO phase, as indicated in Fig. 7 and Table 1. During the first negative IPO phase, the total cases of canonical La Niña and La Niña Modoki are higher than the total cases of canonical El Niño and El Niño Modoki (26 and 13 cases, respectively). In this case, the negative IPO triggers a cooling trend over the tropical Pacific Ocean and a higher magnitude and frequency of canonical La Niña and La Niña Modoki. Conversely, the total cases of canonical El Niño and El Niño Modoki are higher than for the total cases of canonical La Niña and La Niña Modoki during the positive IPO phase (11 and 7 cases, respectively). However, in another negative IPO phase (1998–2016), the positive and negative periods of canonical ENSO and ENSO Modoki have similar total cases.

Table 2. Frequency of occurrence of the three interannual climate modes in each IPO phase during 1939–2016.

| IPO Phase     | El Niño | La Niña | Pos. IOD | Neg. IOD | El Niño Modoki | La Niña Modoki |
|---------------|---------|---------|----------|----------|----------------|----------------|
| 1939–1978 (Neg. IPO) | 5       | 18      | 6        | 7        | 8              | 8              |
| 1979–1997 (Pos. IPO)   | 4       | 5       | 6        | 6        | 7              | 2              |
| 1998–2016 (Neg. IPO)   | 5       | 5       | 5        | 6        | 6              | 6              |

To understand the Indonesian rainfall variability related to IPO, we perform a composite analysis of the rainfall, SST, and wind anomalies averaged over the JJASON season for the periods of the three IPO phases (Fig. 9). During the negative IPO phases in 1939–1978 and 1998–2016, positive rainfall anomalies occur over almost all the Indonesian region for both CRU and GPCP datasets (Figs. 9a, c). This positive anomalous rainfall is associated with anomalous easterly winds from the Pacific Ocean to the Indonesian region and with an anomalous westerly wind over the Indian Ocean across Indonesia. These anomalous easterly and westerly winds are induced by negative SSTAs in the eastern Pacific and western Indian Ocean, respectively (Fig. 9f). This process is similar to canonical La Niña (Hendon 2003; Dong and Dai 2015). Nevertheless, there is a striking difference in global SSTAs between these two negative IPO phases. The global SSTAs is dominated by negative values in the first negative IPO phase (1939–1978) and by positive values in the second negative IPO phase (1998–2016). This difference in global SSTAs may be attributed to the component of global warming that is not eliminated in this study, and we must also consider other decadal modes, such as the Atlantic Multidecadal Oscillation (Liu 2012; Dong and Dai 2015). During the positive IPO phase in 1979–1997, negative rainfall anomalies occur over the whole Indonesian region (Fig. 9b), associated with westerly wind anomalies from eastern Indonesia to the Pacific Ocean and with easterly wind anomalies from western Indonesia to the Indian Ocean. These anomalous westerly and easterly winds are induced by positive SSTAs in the eastern Pacific and western Indian Ocean, respectively. This process is similar to canonical El Niño when drought conditions in Indonesia during the dry season are accompanied by surface south-easterlies across and to the west of Indonesia toward the west Indian Ocean (e.g., in Figs. 4b, c of Hendon 2003; Dong and Dai 2015).

To identify the modulation impact of IPO, the spatial distributions of the partial correlations between Indonesian rainfall and the interannual climate indices for each of the three IPO phases are also analyzed (Fig. 10). Generally, the partial correlation patterns in each IPO phase are consistent with Fig. 4, in which the analysis period is 1979–2016. However, there are slight differences in the affected area for each climate mode among the IPO phases. In the positive and second negative phases of IPO (Figs. 10b, c for the CRU data and Figs. 10d, e for the GPCP data), there are some differences in the influence of the interannual climate modes. For canonical ENSO, the affected area over Indonesia is decreased from the positive phase to the second negative phase, especially over central to eastern Indonesia. Additionally, the influences of ENSO Modoki and IOD are stronger in the second
negative phase than in the positive phase, especially over the southern part of Indonesia for ENSO Modoki and over the southwestern part for IOD.

To understand the influence of large-scale variability on the regional-scale Indonesian rainfall variability during each climate mode for each IPO phase, we calculate the partial correlations between climate indices and SSTA and horizontal wind anomalies at 850 hPa (Fig. 11). The SSTA and wind correlation patterns among the IPO phases are generally consistent, except for canonical ENSO. For canonical El Niño, the negative correlation of the SSTA patterns over Indonesia
Partial correlation between rainfall and climate indices on each IPO phase

Fig. 10. (a–c) The same as in Fig. 5a, except for (a) the first negative IPO phase, (b) the positive IPO phase, and (c) the second negative IPO phase. (d, e) The same as in Fig. 5b except for (d) the positive IPO phase, and (e) the second negative IPO phase.

Changes from the positive IPO phase (Fig. 11b for NINO3) to the second negative IPO phase (Fig. 11c for NINO3), where the negative SSTA correlation moves slightly southeastward. In the second negative IPO phase (Fig. 11c), the negative SSTA correlations over Indonesia are more substantial in El Niño Modoki and positive IOD than in canonical El Niño and are also more robust compared with the first negative and the positive IPO phase (Figs. 11a, b).

We investigate the correlation between the interannual climate indices and Indonesian rainfall averaged over two representative areas to show the modulation by the IPO of the influence of the interannual climate modes. The areas are central to eastern Indonesia...
Figure 12 compares the correlation between interannual climate modes and area-averaged rainfall in IPO phases from 1939 to 2016. The correlation coefficient between the rainfall over central to eastern Indonesia and EMI (Fig. 12b) is higher in the second negative IPO phase than in the positive phase. The results for the correlation coefficients between DMI and the rainfall over southwest Indonesia are similar (Fig. 12c). Different results are found for canonical ENSO, where the correlation between NINO3 and the rainfall over central to eastern Indonesia is slightly smaller in the second negative IPO phase (Fig. 12a).

The modulation by the IPO of the effects of the interannual climate response to Indonesian rainfall is also analyzed by using 13 year sliding partial correlation between PC1 of Indonesian rainfall and the climate indices (Fig. 13). Generally, the responses in PC1 of the CRU data and PC1 of the GPCP data to all three interannual climate modes are consistent. The modulation by the IPO phases of the responses of Indonesian rainfall to canonical ENSO and ENSO Modoki is not significant, where the correlation coefficient between the IPO index and the time series of the 13 year sliding partial correlation between PC1 of the CRU data with NINO3 is −0.30 and that with EMI is 0.09. However, the response of Indonesian rainfall to canonical ENSO is significant until the 2000s and weakens thereafter (Fig. 13a). This is consistent with the result that the impacted area of canonical ENSO is decreased in the second negative IPO phase (Fig. 10c for NINO3) compared with the previous IPO phases. By contrast, the Indonesian rainfall response to ENSO Modoki is continuously significant after the 1990s (Fig. 13b). For IOD, the correlation is weakened when
the IPO is in the positive phase and strengthened when
the IPO is in the negative phase (Fig. 13c). Figure
10 for DMI on both rainfall datasets shows that the
responses of Indonesian rainfall to IOD are significant
during the negative IPO phase, where the correlation
coefficient between the IPO index and the time series
of the 13 year sliding partial correlation between PC1
of the CRU data with DMI is 0.66.

4. Summary and discussion

We have investigated the interannual and inter-
decadal variability of Indonesian rainfall associated
with climate modes over the Indo–Pacific Ocean. Monthly rainfall data from the CRU from 1901 to
2016 and from the GPCP from 1979 to 2016 for the
JJASON season were used. Despite the differences
in source input data, the EOF analysis revealed that
PC1 of the CRU and GPCP data for 1979 to 2016
show similarities in the temporal and spatial patterns
(Figs. 2a, b for EOF1, Figs. 3a, b, Fig. 6a). The PC1
time series for the CRU and GPCP data are related
to the canonical ENSO, ENSO Modoki, and IOD,

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Fig. 12. Time series of normalized sign-reversed rainfall averaged over the area in Fig. 4, and interannual climate
mode indices. (a) Rainfall over the rectangular areas in eastern Indonesia (118–135°E, 7°S–4°N) and normalized
NINO3 in JJASON, (b) rainfall over eastern Indonesia and normalized EMI, and (c) rainfall over the rectangular
areas in southwest Indonesia (100–110°E, 7.5–2.5°S) and normalized DMI. The sign of the rainfall is reversed
for easier comparison with the climate mode indices, and $r$ denotes the correlation coefficients between the sign-
reserved rainfall and the indices.
suggesting that these are the major interannual climate modes influencing Indonesian rainfall (Figs. 6d–f). A correlation technique shows that PC1 of the CRU data and PC1 of the GPCP data produce similar patterns for SSTA and wind circulation anomalies at 850 hPa over the Indo–Pacific (Fig. 4). Moreover, the results show significant correlation coefficients over the central Pacific, highlighting the ENSO Modoki pattern.

On the interdecadal timescale, Indonesian rainfall is related to the IPO, with significant correlations of the PC1 of the CRU data and PC2 of the GPCP data with the IPO index.

All the interannual climate modes examined show a negative correlation with Indonesian rainfall (Fig. 5). The effects of canonical ENSO and ENSO Modoki on rainfall prevail over almost all of Indonesia, which is larger than the area influenced by IOD. The regionality of the canonical ENSO and ENSO Modoki influences is different, especially over western and southern Indonesia. The ENSO Modoki influence is stronger over southern Indonesia and weaker over western Indonesia compared with the canonical ENSO influence. The responses of rainfall to the IOD are confined in southwestern Indonesia. These results are consistent with those of Alsepan and Minobe (2020), which suggest that the differences in the affected area

Fig. 13. Time series of 9 year running mean TPI–IPO (light blue bars) and the 13 year sliding partial correlation coefficient of PC1 of the CRU data (solid orange line) and PC1 of the GPCP data (orange dashed line) with (a) NINO3, (b) EMI, and (c) DMI. The horizontal solid red lines denote significant correlations at the 95% confidence level.
among the climate modes are related to moisture flux convergence anomalies.

Previous studies on Indonesian rainfall variability focus only on canonical ENSO and IOD events. As-syakur et al. (2014) analyzed the influences of ENSO and IOD for a shorter period using a high-resolution (0.25° × 0.25°) satellite-derived rainfall dataset over 13 years from 1998 to 2010. Their results for canonical ENSO are consistent with the present study, but for IOD, we found much larger responses of rainfall over southwestern Indonesia (Figs. 5a, b for DMI) than their study, in which the responses of rainfall are not statistically significant over land (their Fig. 3 for JJA and SON). This difference may be attributed to the longer period of the datasets in this study compared with As-syakur et al. (2014). Yanto et al. (2016) also identified variability in Indonesian rainfall on an interannual timescale. Based on their EOF analysis of CRU rainfall data for the period of 1901–2012, the driver of the leading modes of rainfall variability is the canonical ENSO. They only considered the variability over the Pacific Ocean without considering the role of Indian Ocean variability. Additionally, the spatial correlation between their PC1 of the CRU data and global SST (Fig. 6a in Yanto et al. 2016) showed a pattern over the Indian Ocean that is similar to the IPO phenomenon, but their analysis did not focus on that area.

Regarding the interdecadal climate variability, the transition from the positive to negative IPO phase affects Indonesian rainfall. During the study period, there are three phases of IPO: two negative phases during 1939–1978 and 1998–2016 and a positive phase during 1979–1997 (Fig. 7a). There is a clear decrease in Indonesian rainfall in the positive phase of the IPO using the CRU and GPCP dataset (Fig. 9b), and there is an increase in rainfall in the negative phase (Figs. 9a, c). On an interannual timescale, the composite of Indonesian rainfall anomalies for five combinations of interannual climate modes (Fig. 8) revealed that the areas with large rainfall anomalies are consistent with the partial correlation pattern and that the co-occurrence of multiple modes has a stronger impact on rainfall. Nur’utami and Hidayat (2016) also reported rainfall composite results for co-occurrences but only for the interannual climate modes canonical ENSO and IOD from 1960 to 2011. By considering the ENSO Modoki, we show a more diverse set of climate modes affecting Indonesian rainfall variability for a longer period.

To examine the interaction between the interannual climate mode and IPO, the influences of canonical ENSO, ENSO Modoki, and IOD on Indonesian rainfall are analyzed for each IPO phase (Fig. 10). For canonical ENSO, the affected area pattern during the second negative IPO phase (1998–2016) is less significant than during the other IPO phases. Conversely, for ENSO Modoki, the affected area is significant during the second negative IPO phase compared with during the other IPO phases. These results indicate that the Indonesian rainfall responses to canonical ENSO and ENSO Modoki are not closely related to IPO phases. Additionally, the time series of the partial correlation of NINO3 and EMI with Indonesian rainfall averaged over two representative areas do not show a consistent pattern with the IPO index (Figs. 13a, b). The different response between canonical ENSO and ENSO Modoki during the last IPO phase might be due to changes in the tropical Pacific because the appearance of ENSO Modoki has been more frequent and persistent in recent decades (Ashok et al. 2007). In northeastern Australia, the rainfall responses to canonical ENSO are stronger during the negative IPO phases, although the mechanism by which the IPO modulates the effect of ENSO on northeastern Australia is still elusive (Dong and Dai 2015). By contrast, the IPO clearly modulates the Indonesian rainfall response to IOD. During the positive IPO phase (second IPO phase), the area with significant rainfall response to IOD (Figs. 10b, d for DMI) is smaller than during the negative IPO phases (Figs. 10a, c, e for DMI). The strong response of Indonesian rainfall to IOD during the negative IPO phase is also seen in the time series of the sliding partial correlation between DMI and Indonesian rainfall (Fig. 13c). Dong et al. (2016) explained that the IPO-induced atmospheric adjustment modulates the decadal variability in Indian Ocean SSTs. Warming over the tropical eastern Pacific Ocean (positive IPO phase) induces positive sea level pressure anomalies over the tropical western Pacific and the Indian Ocean. Additionally, this warming weakens the tropical Walker circulation, reduces cloudiness, and increases the sea surface height and thus, increases the SSTs in the Indian Ocean. The weakening or strengthening of the Walker circulation triggers a change in Indonesian rainfall.

Yanto et al. (2016) also analyzed Indonesian rainfall on the multidecadal timescale. Their results showed that there was a strong coherency of PC1 of the CRU data in the decadal band of an 8–16-year period after 1980. These conditions are similar to the ENSO pattern, but at a multidecadal timescale, which is associated with IPO. However, their study focused on PC2, which was related to the PDO. Additionally, based on
their results, the strong coherency of the modulation of Indonesian rainfall by the ENSO at the interannual and interdecadal timescales is a reason for the significant strengthening of the influence of NINO3.4 on Indonesian rainfall after 1980. Our analysis shows that in recent decades, the response of Indonesian rainfall to canonical ENSO has weakened, and the response of Indonesian rainfall to the ENSO Modoki and IOD has strengthened.

In conclusion, our study compared the results of the three interannual climate modes over the Indo–Pacific to examine the effects of the modes on Indonesian rainfall in two types of rainfall data. By contrast, previous studies have focused on the interactions of only one or two interannual climate modes on one data type. We showed that the three interannual climate modes associated with the IPO, which is an interdecadal phenomenon, also affect the interdecadal variability of Indonesian rainfall. To increase the capability of rainfall forecasts in Indonesia, further analysis of these climate modes on interannual and interdecadal timescales should be conducted. Model data, such as the Coupled Model Intercomparison Project, can be used to investigate the interactions between the effect of the climate modes and anthropogenic global warming that might occur if these conditions continue according to the emissions scenarios in the model data. Furthermore, the possible interactions between oceans (Pacific–Atlantic–Indian Ocean) and the climate variability in tropical Atlantic oceans should be considered.

Supplements

Figure S1. Spatial distribution of correlation coefficients of SST and 850 hPa wind anomalies with the JJASON PC2 time series of (a) CRU rainfall and (b) GPCP rainfall data for the period 1979–2016 period. The correlation coefficients that are significant at the 95 % confidence level are shown by color shading and vectors.

Figure S2. Power spectrum of (a) Indonesian rainfall PC1 CRU, (b) PC2 CRU, (c) PC1 GPCP, (d) PC2 GPCP, (e) IPO, (f) NINO3, (g) EMI, (h) DMI for JJASON during 1979 to 2016. The dashed curve is the 90 % confidence level based on a “Chi-Squared” test of the theoretical red noise spectrum.

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