Influence of the Pacific–Japan Pattern on Indian Summer Monsoon Rainfall

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ABSTRACT

This study discusses the impact of the Pacific–Japan (PJ) pattern on Indian summer monsoon (ISM) rainfall and its possible physical linkages through coupled and uncoupled pathways. Empirical orthogonal function analysis of 850-hPa relative vorticity over the western North Pacific (WNP) is used to extract the PJ pattern as the leading mode of circulation variability. The partial correlation analysis of the leading principal component reveals that the positive PJ pattern, which features anticyclonic and cyclonic low-level circulation anomalies over the tropical WNP and around Japan respectively, enhances the rainfall over the southern and northern parts of India. The northwestward propagating Rossby waves, in response to intensified convection over the Maritime Continent reinforced by low-level convergence in the southern flank of westward extended tropical WNP anticyclone, increase rainfall over southern peninsular India. Meanwhile, the anomalous moisture transport from the warm Bay of Bengal due to anomalous southerlies at the western edge of the low-level anticyclone extending from the tropical WNP helps to enhance the rainfall over northern India. The atmospheric general circulation model forced with climatological sea surface temperature confirms this atmospheric pathway through the westward propagating Rossby waves. Furthermore, the north Indian Ocean (NIO) warming induced by easterly wind anomalies along the southern periphery of the tropical WNP–NIO anticyclone enhances local convection, which in turn feeds back to the WNP convection anomalies. This coupled nature via interbasin feedback between the PJ pattern and NIO is confirmed using coupled model sensitivity experiments. These results are important in identifying new sources of ISM variability/predictability on the interannual time scale.

1. Introduction

The Indian summer monsoon (ISM) system is sensitive to both local forcing and remote influences (e.g., Goswami 1998; Webster et al. 1998; Annamalai et al. 2005; Wu and Kirtman 2003). Prediction of ISM rainfall is a challenging task for the scientific community due to its complex nature. It is therefore important to focus on different aspects that influence the ISM rainfall variability. Sea surface temperature (SST) over the tropical regions is a key parameter in providing precursor for the ISM prediction (e.g., Shukla 1984; Delsole and Shukla 2002; Krishnamurti et al. 2000; Terray et al. 2003; Gadgil et al. 2005; Krishna Kumar et al. 2005; Rajeevan et al. 2007). On the interannual time scale, SST anomalies associated with El Niño–Southern Oscillation (ENSO)
are the dominant forcing for the ISM variability in spite of the uncertainty in the stability of the monsoon–ENSO teleconnections (Krishna Kumar et al. 1999, 2006; Lau and Nath 2000; Turner and Annamalai 2012). Many studies have addressed the teleconnections between ISM rainfall and ENSO (e.g., Shukla and Paolino 1983; Webster and Yang 1992). ENSO influences ISM directly through large-scale circulation changes over the Indo-western Pacific (e.g., Webster and Yang 1992; Annamalai et al. 2005; Krishna Kumar et al. 2005; Wu et al. 2012) by modulating the tropical Indian Ocean (TIO) SST (Kawamura et al. 2001; Wu and Kirtman 2004; Achuthavarier et al. 2012; Krishnan et al. 2006) and the meridional temperature gradient over South Asia (Yang and Lau 1998; Goswami and Xavier 2005).

Apart from ENSO, the dominant modes in the Pacific such as the Pacific decadal oscillation (PDO) and interdecadal Pacific oscillation (IPO) have considerable impact on ISM rainfall variability at various time scales (Krishnan and Sugi 2003; Joshi and Kucharski 2017; Joseph et al. 2013; Malik et al. 2017). The PDO is the first leading mode of variability in the North Pacific (20°–70°N) monthly SST anomalies (Mantua et al. 1997). Krishnamurthy and Krishnamurthy (2014) suggested that the in-phase (out of phase) PDO and ENSO could enhance (weaken) the conventional ENSO–ISM relationship. This indicates that the PDO phase is important in the ENSO–ISM rainfall relationship. The IPO warm (cold) phase is characterized by warm (cold) SST anomalies in the tropical Pacific and cold (warm) SST anomalies in the central North Pacific (Trenberth and Hurrell 1994; Meehl et al. 2009). It has been reported that during the negative phase of IPO, rainfall enhances in some parts of central-northwest India and peninsular India and reduces over northeast India (e.g., Joshi and Rai 2015). These observations demonstrate that the leading modes of the Pacific SST variability display significant impact on ISM rainfall on interannual and longer time scales. However, the influence of the dominant mode in the low-level circulation, namely the Pacific–Japan (PJ) pattern (Nitta 1987), over the western North Pacific (WNP; 0°–60°N) on ISM rainfall is not explored.

Previous studies highlighted the PJ pattern as the dominant mode of climate variability in the WNP during boreal summer [June–August (JJA)] (Nitta 1987; Kosaka and Nakamura 2006; Kosaka et al. 2013; Kubota et al. 2016). The PJ pattern features a meridional dipole structure in lower-tropospheric circulation and precipitation anomalies with the tropical and midlatitude WNP lobes (Nitta 1987; Kosaka and Nakamura 2006) and provides a crucial link between the tropics and midlatitudes. The positive PJ pattern features anticyclonic (cycloidal) surface anomalies with drier and hotter (wetter and cooler) summer in the tropical (midlatitude) WNP (e.g., Huang and Sun 1992; Kubota et al. 2016). The positive PJ pattern is significantly correlated with the decay phase of El Niño (e.g., Sun et al. 2010; Xie et al. 2016). On the other hand, the negative phase of the PJ pattern causes droughts and heat waves in the mid-latitudes with enhanced tropical cyclone activity over the WNP region (e.g., Kosaka et al. 2013). The tropical cyclone activity, whether enhanced or weakened by the PJ teleconnection, affects the East Asian monsoon significantly (e.g., Choi et al. 2010).

While impacts of the PJ pattern on East and Southeast Asia are well documented, its relationship with ISM rainfall variability is yet to be explored. Anomalous diabatic heating over the tropical WNP associated with the positive PJ pattern excites a cold Rossby wave, which extends westward to the NIO and Indian peninsula (Kosaka et al. 2013) and potentially affects ISM rainfall. Furthermore, anomalous surface easterlies along the southern flank of the cold Rossby wave weaken the prevailing monsoon westerlies over the summer NIO and warm the SST by reducing the latent heat flux (e.g., Du et al. 2009), which may influence the ISM. Xie et al. (2009) and Chowdary et al. (2011) showed that the warmer NIO SST excites an equatorial warm Kelvin wave that propagates to the western Pacific and suppresses convection over the tropical WNP via surface Ekman divergence, thereby exciting the PJ pattern. This eastward influence from the NIO to the PJ pattern, combined with the westward influence from the PJ pattern, constitutes an interbasin feedback, a key mechanism for the Indo-western Pacific Ocean capacitor (IPOC) mode (Xie et al. 2016). The Indo-western Pacific interbasin coupling itself does not require ENSO forcing, but El Niño excites the IPOC mode by inducing the NIO warming (Kosaka et al. 2013) and tropical WNP cooling (Wang et al. 2013; Stuecker et al. 2015) as initial perturbations. Remote influences of the PJ pattern on the ISM, if any, may have potential implications for the IPOC mode.

The purpose of the present study is therefore to provide a detailed understanding of the impact of the PJ teleconnection pattern on ISM rainfall variability, which will deepen our understanding of ISM variability beyond the well-known influence from ENSO and Indian Ocean variability. To the best of our knowledge this study is first of its kind. In section 2, we describe the details of data, methods, and models used. In section 3, the observed influence of the PJ pattern on ISM rainfall is discussed. The atmospheric pathway of PJ impact is presented using an atmospheric general circulation...
2. Data, model, and analysis methods

Monthly atmospheric data such as wind at pressure levels and mean sea level pressure (MSLP) for the period of 1979–2015 are obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis dataset (ERA-Interim; Dee et al. 2011). Tropospheric temperature is defined as mean temperature between 850 and 200 hPa. Precipitation data from Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) and the latest version of Hadley Centre Sea Ice and SST (HadISST; Rayner et al. 2003) datasets are utilized in the present study. We extract the PJ pattern as the dominant mode of lower-tropospheric circulation variability over the WNP region by performing an empirical orthogonal function (EOF) analysis of 850 hPa vorticity over 0°–60°N, 100°–160°E during the boreal summer season (JJA), similar to Kosaka et al. (2013) but for interannual variability of seasonal-mean anomalies. Statistical techniques such as correlation/regression and partial correlation/regression methods (e.g., Ashok et al. 2007) are used to examine the impact of the PJ pattern on ISM rainfall. As discussed earlier, the PJ pattern is associated with the TIO basinwide warming and ENSO. In this context we performed partial correlation analysis to extract the pure influence of the PJ pattern on the ISM. Yet, we can expect similar influence operative even when it is coupled with SST variability.

The partial correlation coefficient $r_{sp,n}$ between two variables $x$, $p$, after removing the influence of a variable $n$ is given by

$$ r_{sp,n} = \frac{r_{sp} - r_{sn}r_{pn}}{\sqrt{(1-r_{sn}^2)(1-r_{pn}^2)}}. \quad (1) $$

In the above equation, $x$ denotes the predictive variable (rainfall, MSLP, SST, 200- and 850-hPa winds, moisture transport, and tropospheric temperature), $p$ is the PJ index (the leading principal component of the 850-hPa vorticity over the WNP), and $n$ is the Niño-3.4 index (SST anomalies averaged over 170°W–120°W and 5°S–5°N). The partial correlation coefficient $r_{sp,nt}$ between two variables $n$, $t$, is obtained by

$$ r_{sp,nt} = \frac{r_{sp,t} - r_{sn,t}r_{pn,t}}{\sqrt{(1-r_{sn,t}^2)(1-r_{pn,t}^2)}} = \frac{r_{sp,n} - r_{xt,n}r_{pt,n}}{\sqrt{(1-r_{xt,n}^2)(1-r_{pt,n}^2)}} \quad (2) $$

where $t$ is the Indian Ocean SST index (SST anomalies averaged over 40°S–100°E and 10°S–25°N).

To examine the atmospheric pathway that links the PJ pattern and ISM, we have used an AGCM experiment forced with observed climatology of SST (Reynolds et al. 2007) globally. The model used in this study is the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model version 2.1 (AM2.1; Anderson et al. 2004) with horizontal resolution of ~200 km and 24 vertical levels. This model run is carried out with a single ensemble member for 200 years. Furthermore, a fully coupled ocean–atmosphere model, the National Centers for Environmental Prediction (NCEP) Climate Forecasting System version 2 (CFSv2; Saha et al. 2014), is used in this study to examine the roles of air–sea interactions. Its atmospheric component is NCEP Global Forecast System (GFS) with 64 sigma layers vertically and T126 (~100 km) horizontal resolution, and the oceanic component is the Modular Ocean Model version 4 (MOM4P0; Griffies et al. 2004) with horizontal resolution of 0.25° in the 10°S–10°N latitude band and 0.5° elsewhere. The CFSv2 retrospective 9-month forecast (hindcast) initialized in May is prepared for a period of 30 years (1985–2014) at the Indian Institute of Tropical Meteorology (IITM), which is referred to as the control run (CTL). There are 10 atmospheric initial conditions (10 ensemble members); these are partitioned into two sets. The first set uses five atmospheric initial states of 1, 2, 3, 4, and 5 May with a common ocean initial condition for the pentad of the first through the fifth days of the same month. The second set uses the five atmospheric initial states of 6, 7, 8, 9, and 10 May with a common ocean initial condition for the pentad of the sixth though tenth days of the same month. For the analysis, we have utilized ensemble mean forecasts obtained by averaging the 10 ensemble members (Srinivas et al. 2018). The ocean initial conditions are obtained from the NCEP Global Ocean Data Assimilation System. The atmospheric initial conditions are obtained from the NCEP Reanalysis (R2) data (Saha et al. 2010). Note that AM2.1 is different from the atmospheric component of CFSv2. This would partially be responsible for the differences in the response of ISM rainfall to the PJ pattern teleconnections in the two models.

The Student’s $t$ test is used to check the statistical significance of partial correlations. Prior to it, the Shapiro–Wilk normality test (Shapiro and Wilk 1965) is
applied on rainfall, MSLP, and winds. It is confirmed that the PJ index (as well as other variables) is normally distributed at 95% confidence level. In addition to this, the normality of distribution is confirmed with a probability density function and a box plot of the PJ index (figure not shown).

3. Influence of the PJ pattern on ISM

The EOF analysis identifies the dominant mode of interannual variability over the WNP in the lower troposphere during the boreal summer season (JJA). The first EOF mode (EOF-1) of 850-hPa relative vorticity features a meridional tripole pattern with positive vorticity anomalies from equator to 10°N and in the midlatitudes (25°–45°N), and negative vorticity anomalies at 10°–25°N (Fig. 1a). This pattern represents the PJ pattern (Nitta 1987), which provides a crucial link between the tropical and midlatitude WNP and impacts rainfall over most East Asian regions in summer (Huang and Sun 1992). Figure 1b displays the correlation of the corresponding principal component of relative vorticity (RV–PC1) with SST and 850-hPa circulation anomalies. Anomalous cyclonic circulation north of the tropical WNP anticyclone and strong low-level convergence over the Maritime Continent are apparent as part of the PJ pattern. The SST shows weak negative anomalies over the equatorial central Pacific, representing a weak La Niña–like pattern. In fact, the simultaneous correlation between RV–PC1 and the Niño-3.4 index is $-0.36$, which is significant at a 95% confidence level (Fig. 1c). Furthermore, the TIO displays basinwide positive anomalies in association with RV–PC1 during summer (Fig. 1b) with a correlation coefficient of $0.47$ (significant at a 95% confidence level), suggesting a positive relationship between TIO warming and the PJ pattern. The TIO basinwide warming follows an El Niño event. Earlier studies found significant correlations of the PJ pattern with El Niño decaying and La Niña developing phases (e.g., Kosaka and Nakamura 2010). The basinwide TIO warming in post–El Niño summers exerts strong impact on South Asian, East Asian, and WNP monsoon rainfall and circulation partly through the PJ pattern (e.g., Yang et al. 2007; Xie et al. 2009; Jiang et al. 2013; Chowdary et al. 2017).
The impact of the PJ pattern on ISM rainfall is examined by correlation analyses (Fig. 2). The raw correlation between the PJ index (RV–PC1) and rainfall over the Indo-western Pacific region displays a meridional wavelike pattern with suppressed precipitation over the tropical WNP and enhanced precipitation over mid-latitude and equatorial regions (Fig. 2a). On the other hand, enhanced precipitation associated with the PJ pattern is notable over the southern Bay of Bengal, Indian subcontinent, and eastern Arabian Sea (Fig. 2a). This enhanced rainfall is due to the combined impact of the PJ pattern, TIO warming, and ENSO. The partial correlation map of the PJ pattern on precipitation anomalies, after removing the influence of ENSO (Niño-3.4) and TIO SST as in (2), shows a significant (at 95% confidence level) positive correlation over the Maritime Continent and southern and northern parts of India (Fig. 2b). The enhanced convection over southern peninsular India is a response to enhanced convection over the Maritime Continent via northwestward propagating warm Rossby waves (e.g., Jiang et al. 2004; Kemball-Cook and Wang 2001; Annamalai 2010). This atmospheric pathway is important in linking the PJ pattern with ISM rainfall.

The PJ pattern dominates on intraseasonal to interannual time scales. The associated anomalous convection over the Maritime Continent induces a warm stationary Rossby wave response with an anomalous low-level cyclonic circulation extending northwestward from the forcing (Matsuno 1966; Gill 1980). This suggests that the enhanced convection over southern peninsular India is a response to enhanced convection over the Maritime Continent via northwestward propagating warm Rossby waves. The anomalous deep convection over the Maritime Continent is strongly tied to the tropical WNP anomalous anticyclone as a part of the PJ pattern, as reflected in MSLP and low-level wind anomalies (Fig. 2c). This atmospheric pathway is important in linking the PJ pattern with ISM rainfall.

Low-level convergence in the southern flank of westward extended tropical WNP anticyclone helps to maintain the convection belt from the Maritime Continent to southern peninsular India along with warm SST anomalies over the NIO due to modulation of low-level winds (Figs. 2c,d). Anomalous convection over the NIO region including the Indian subcontinent triggered by local SST warming in turn affects the PJ related circulation over the WNP through the atmospheric Kelvin
wave–induced Ekman divergence mechanism (Fig. 2c; Xie et al. 2016; Ma et al. 2017), thereby closing an interbasin feedback suggested for the IPOC mode. The interaction between circulation associated with the PJ pattern and NIO SST warming highlights the importance of the coupled perspective in the context of the PJ impact on ISM rainfall apart from the atmospheric pathway through the Rossby waves, as examined in section 5.

The spatial pattern of partial correlations suggests that the PJ pattern features the anomalous anticyclonic circulation over the tropical WNP extending to the NIO as a cold Rossby wave response to precipitation decrease over the tropical WNP (Lau and Peng 1990; Lawrence and Webster 2002; Chowdary et al. 2013). Positive rainfall anomalies over northern India are associated with moisture transport by southerlies in the
northwestern flank of the westward extended anomalous anticyclone (Fig. 3a). Anomalous low-level convergence associated with the moisture transport is evident in the northern parts of India, although displaced slightly southwest of the enhanced rainfall. Further, this moisture transport along with the Himalayan orographic uplift helps to sustain positive rainfall anomalies in this region. Weak upper-level (200 hPa) divergence over southern and northern India is consistent with low-level convergence (Fig. 3b). Warm tropospheric temperature is in line with the convection over India and the central Indian Ocean by enhancing the moist adiabatic adjustment (Fig. 3c). On the other hand, low tropospheric temperature over the tropical WNP region is consistent with circulation and precipitation patterns.

Partial correlation of the PJ index with 850-hPa streamfunction and rotational components of wind, and 200-hPa velocity potential and divergent component of wind are illustrated in Figs. 4a and 4b respectively. Streamfunction anomalies show westward extension of tropical WNP anticyclonic anomalies and northward extending cyclonic anomalies from the Maritime Continent toward the southern tip of India in association with the PJ pattern. This is also reflected in the 200-hPa velocity potential and divergent component of wind anomalies as well as anomalous upper-level convergence over the WNP and divergence over the TIO. A zonal–vertical cross section of circulation averaged over northern latitudes (5°–25°N) exhibits ascending (descending) motion with low-level convergence (divergence) over the ISM (tropical WNP) region. This suggests the importance of suppressed convection over the tropical WNP for PJ impact on ISM rainfall. On the other hand, anomalous meridional overturning circulation averaged over 130°–160°E shows alternating rising and sinking motions over northern latitudes (Fig. 4d). Strong subsidence and descending motion over the tropical WNP region (near the Philippines) and strong ascending motion north and south of it represent the typical PJ pattern.

Overall, the PJ pattern arising from the interbasin ocean–atmosphere feedback enhances the convection over some parts of the Indian subcontinent and NIO through the westward propagating Rossby waves and the changes in zonal overturning circulation around 10°N. This suggests that the atmospheric pathway (atmospheric Rossby wave response) and coupled interaction (the PJ and NIO interaction) play vital roles for the PJ influence on ISM rainfall. Further to test our hypothesis we have carried out sensitivity experiments using AM2.1 and CFSv2.

4. The atmospheric pathway

To exclude the influence of SST variability and to verify the atmospheric pathway of the PJ impact on ISM
rainfall, we use AM2.1 climatological SST run for 200 years. It is noticed that strong negative rainfall anomalies over the tropical WNP associated with anomalous anticyclonic circulation and suppressed convection are crucial components of the PJ pattern (e.g., Kosaka et al. 2013; Kubota et al. 2016; Li et al. 2014; Xie et al. 2016). In observations, partial regressions of the Indo-western Pacific rainfall and 850-hPa wind anomalies upon the observed area averaged rainfall over the tropical WNP ($15^\circ$–$25^\circ$N, $110^\circ$–$150^\circ$E) apparently display the PJ pattern over the WNP with positive rainfall and easterly wind anomalies over the Bay of Bengal and southern peninsular India (Fig. 5a), similar to Fig. 2b. When regressed onto this tropical WNP rainfall index, somewhat similar patterns in rainfall and circulation are noted in the AGCM experiment as well (Fig. 5b). In particular, positive rainfall anomalies over the Maritime Continent and southern peninsular India, linked via a northwestward-propagating warm Rossby wave response, are well captured by the AGCM climatological SST run with some differences in magnitude. The AGCM experiment indicates that without any influence of SST, this intrinsic PJ mode apparently interacts with ISM rainfall through a westward propagating warm Rossby wave and moisture convergence.

Differences are noticeable in regression patterns between the AM2.1 run and observations, especially over the Bay of Bengal and northern India (Fig. 5). The positive rainfall band over northern India is shifted too far west in the model compared to the observations (Fig. 4). These differences could be due to the design of the climatological SST run as well as the model resolution and other model inherent properties. Still, it is noted that positive rainfall anomalies over the Maritime Continent and southern peninsular India are well captured by the AGCM climatological SST run, along with the circulation pattern over the WNP–NIO region.

5. Role of air–sea interactions

Analysis of the observations evidences the influence of NIO SST in linking the PJ pattern and ISM rainfall. Figure 6 shows the regression of 850-hPa wind and precipitation anomalies onto rainfall averaged over the tropical WNP in the observations and CFSv2 CTL run during JJA for the period 1985 to 2014. Enhanced rainfall north and south of suppressed tropical WNP convection and over most of the Indian subcontinent is well reproduced by the model as compared to the observations. The coupled model is also able to reproduce circulation anomalies associated with the PJ pattern over the Indo-western Pacific region fairly well (Fig. 6b). This suggests that the coupled model is able to represent the association of the PJ teleconnection with ISM rainfall and circulation. To examine the relation between the

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PJ pattern and ISM rainfall through air–sea interactions, we have carried out a sensitivity experiment (EXP) with 1990 conditions, which are free from both the PJ pattern and ENSO in observations (Fig. 7a). Anomalies of precipitation and circulation for JJA in the ensemble mean of CTL are calculated with respect to 1985–2014 climatology. EXP is similar to CTL but tropical WNP convection is suppressed by imposing strong negative SST anomalies (varying between \(-1^\circ\) and \(-2^\circ\)C) over the tropical WNP region to artificially excite the PJ pattern. Corresponding anomalies in the EXP run with respect to climatology display a PJ-like pattern in precipitation but with widespread negative values over the WNP (Fig. 7b). Nevertheless, positive rainfall anomalies in the northern and southern flanks of the anticyclone are evident in the EXP run. Most importantly, enhanced rainfall over most of the Indian subcontinent is evident as a result of the PJ-like anomaly pattern in the model. The rainfall difference between EXP and CTL is positive over the Indian subcontinent and negative over the tropical WNP with enhanced anticyclonic circulation (Fig. 7c). Compared to the observations, the convection anomalies shifted southward over the Maritime Continent and east equatorial Indian Ocean in EXP (Figs. 7b and 6a). This would affect the magnitude of positive precipitation anomalies over southern peninsular India.

In CTL, the spatial pattern of TIO SST displays negative anomalies over the Arabian Sea region and positive anomalies over the central southern Indian Ocean and southern parts of Bay of Bengal (Fig. 8a). The MSLP anomalies are negative over most of the region except for the northwestern parts of India and the Arabian Sea. In the EXP run, anomalously high MSLP and easterly winds from the tropical WNP extend toward the Indian subcontinent as in observations (Figs. 7b and 6a). These easterly wind anomalies over the NIO reduce the strength of prevailing southwestlies and latent heat release to the atmosphere, which in turn causes warm SST anomalies locally (e.g., Du et al. 2009). This illustrates the interbasin influence of the PJ pattern on the NIO, and the warmer NIO likely contributes to the enhanced ISM rainfall, especially over the southern parts of the Indian subcontinent. It is thus suggested that the PJ pattern impacts on ISM rainfall through the atmospheric pathway are further amplified by the air–sea interaction over the NIO. This enhanced rainfall over the NIO region results in increased tropospheric temperature and affects the WNP circulation through the Kelvin wave–induced Ekman divergence mechanism (Xie et al. 2009). This interbasin feedback between the WNP and NIO is a part of the IPOC mode. Thus rainfall patterns associated with the PJ pattern

![Fig. 7. Spatial pattern of JJA precipitation (shaded; mm day\(^{-1}\)) and 850-hPa wind (vectors; m s\(^{-1}\)) anomalies for (a) the CTL run in 1990, (b) the EXP run, and (c) their difference (EXP minus CTL).](image-url)
over the WNP and ISM region can be explained through the IPOC mode.

6. Summary and conclusions

Leading modes of the Pacific Ocean such as ENSO, the PDO, and the IPO are known to have strong influence on ISM on interannual to interdecadal time scales (e.g., Joshi and Kucharski 2017; Krishnan and Sugi 2003; Krishnamurthy and Goswami 2000). Previous studies have paid less attention to the impact of the PJ pattern, the leading mode of variability in WNP circulation, on ISM rainfall variability. The present study examines the impact of the PJ pattern on ISM rainfall variability. The partial correlation of the PJ index (the leading PC of 850-hPa relative vorticity over the WNP) with precipitation anomalies shows significantly positive correlation over the Maritime Continent and southern and northern parts of India. Enhanced convection in southern peninsular India is due to the response of deep convection over the Maritime Continent through northwestward propagation of warm Rossby waves (see Fig. 9). Enhanced deep convection over the Maritime Continent is associated with tropical WNP anomalous anticyclone as a part of the PJ pattern (e.g., Xie et al. 2016). In conjunction with this, the east–west overturning circulation cell with ascending motion corroborated by low-level convergence and upper-level divergence over the Indian subcontinent and strong subsidence over the WNP region indicates the influence of the PJ pattern on ISM rainfall. It is found that the positive rainfall band

![Fig. 8. Spatial pattern of SST (°C; shaded) and MSLP (hPa; contours) anomalies for (a) CTRL, (b) EXP, and (c) their difference.](image)

![Fig. 9. Schematic diagram that shows pathways associated with the impact of the PJ pattern on ISM rainfall. Ellipse with solid (dashed) line indicates anomalous anticyclone (cyclonic) circulation over the tropical (midlatitude) WNP and vectors represent broad low-level circulation associated with PJ pattern. Thick black (white) arrow represents westward extension of warm (cold) Rossby wave, and dark gray (light gray) shading in the Indo-western Pacific Ocean region indicates warm (cold) SST anomalies.](image)
over northern India is due to anomalous low-level moisture convergence in the northwestern edge of the westward propagating atmospheric cold Rossby wave as a response to suppressed convection over the tropical WNP region associated with the PJ pattern. This highlights the importance of the atmospheric pathway of the PJ influence on ISM rainfall (Fig. 9). Further, easterly wind anomalies over the NIO associated with westward extension of the tropical WNP anticyclone warm SST by reducing evaporation. This warming amplifies enhanced atmospheric convection over the NIO and southern parts of India, in turn influencing the PJ pattern (e.g., Kosaka et al. 2013). Thus the interbasin interaction between the NIO and WNP circulation reinforces the PJ impact on ISM rainfall through air–sea coupling.

To verify the PJ influence on ISM rainfall through the atmospheric pathway, the AM2.1 climatological SST run is used. The intrinsic PJ mode captured in this model exerts significant precipitation anomalies over the ISM region, especially over southern parts of India, through westward extension of the tropical WNP anticyclone. These results suggest that modulations of atmospheric circulation by the PJ pattern could affect the ISM rainfall even without any influence of SST change. Besides, the air–sea coupled influence connecting the PJ pattern and ISM through the NIO is examined with a coupled model (CFSv2). A PJ-like anomalous circulation pattern is generated in a coupled model sensitivity experiment by prescribing local negative SST anomalies over the tropical WNP region during summer. This experiment showed enhanced rainfall over the southern and central northern India and suppressed rainfall over the east coast of India, although with some magnitude differences compared to the observations. This result indicates that the PJ mode plays an important role in ISM rainfall variability in a coupled system. Kosaka et al. (2013) identified a coupled mode over the Indo-WNP region that arises from interaction between TIO SST and the PJ pattern during summer. This coupled nature can enhance the westward extension of the PJ pattern and contributes to the predictability of ISM rainfall. Further, the present study shows a robust interbasin influence of the PJ pattern. In particular, NIO SST warming due to easterly wind anomalies associated with the PJ pattern could contribute to increased rainfall in the former region including parts of the Indian subcontinent. This supports the idea of the interbasin interaction associated with the IPOC mode. Wang et al. (2015) pointed out that the operational forecasts during recent decades are found to have little skill in predicting ISM rainfall, basically due to lack of new predictors. Thus, identifying sources of ISM rainfall variability is a prime need. The present study suggests that the PJ pattern and IPOC mode would be useful elements to be considered for monsoon prediction.

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APPENDIX

Long-Term Stability of the PJ Teleconnections

Before looking at the long-term stability of the PJ teleconnections toward ISM rainfall, we have analyzed various precipitation products to reconfirm the robustness of signals obtained (Fig. 2a) over the ISM region. Figure A1 displays partial correlation of PC1 with the GPCC, ERA-Interim, and CRU precipitation datasets. It is found that all products display similar patterns of rainfall over the ISM region. This demonstrates robustness of the PJ influence on ISM rainfall.

Long-term stability of the PJ impact on ISM rainfall for the period of 1901 to 2014 is examined further using the Twentieth-Century Reanalysis (20CR; Compo et al. 2011) 850-hPa winds, the Extended Reconstructed Sea Surface Temperature (ERSST; Huang et al. 2015), and the CRU gridded monthly land precipitation dataset (Harris et al. 2014). A 9-yr running mean has been subtracted to suppress decadal and longer-term variations in the datasets. The leading EOF mode of 20CR 850-hPa relative vorticity during JJA for the period of 1901 to 2014 is shown in Fig. A2a. The PJ mode with a meridional tripole pattern of vorticity anomalies consisting of positive lobes in the equatorial (0°–10°N) and midlatitude (25°–45°N) WNP and a negative lobe in the tropical WNP (10°–25°N) is similar to Fig. 1a. The corresponding RV-PC1 varies interannually throughout the analysis period (Fig. A2c). Weaker variability in the earlier period is likely due to ensemble averaging of 20CR under fewer observational constraints (Kubota et al. 2016). Correlation of RV-PC1 with high-resolution CRU precipitation (land only) and 850-hPa winds is illustrated in Fig. A2b. Anomalous anticyclonic circulation...
in the tropical WNP and low-level convergence over the Maritime Continent and north of 25°N are evident as a part of the PJ pattern. Enhanced rainfall over the Maritime Continent and southern peninsular India are apparent. These results indicate robust teleconnections of the PJ pattern to ISM even for the longer period of data and with different resolution.

To examine the temporal variability of the PJ teleconnections to ISM rainfall, we have carried out partial correlation analysis with 20CR RV-PC1 similar to Fig. 2 for different epochs (Fig. A3). Figure A2c shows modulations in strength of interannual variability of 20CR RV-PC1 with lesser magnitude for the period 1910–40 (epoch 1), moderate for 1958–78 (epoch 2), and high for 1979 to 2015 (epoch 3). These epoch selections are supported by previous studies (e.g., Krishna Kumar et al. 1999; Xie et al. 2010; Chowdary et al. 2012). The modulations in the PJ pattern enable us to investigate the interdecadal modulations of the PJ impact on ISM rainfall. Note that the climate regime shift in ocean–atmospheric anomalies over the tropical Indo–Pacific Oceans in the mid-1970s modulated many of the well-known teleconnections as reported by several studies (Zhang et al. 1997; Garreaud and Battisti 1999). For example, the correlation of ISM rainfall with the ENSO developing phase has changed significantly from epoch 2 to epoch 3 (e.g., Krishna Kumar et al. 1999). Wu and Wang (2002) reported a strengthening correlation of the WNP summer monsoon with ENSO after the mid-1970s. In addition to this, Xie et al. (2009) showed the enhanced TIO warming during El Niño decay phase in the recent epoch displaying a strong impact on the atmosphere over the WNP region. Thus, the PJ teleconnections also potentially modulate from one epoch to another. Anomalous easterly winds over the ISM region and positive rainfall anomalies over the Maritime Continent and southern parts of India associated with the PJ pattern are apparent in all three epochs (Fig. A3). Some changes
in rainfall pattern over the northern parts of India from epoch 1 to epoch 3 are also found. These differences are possibly related to the strength of easterly winds associated with the PJ pattern. Nevertheless, the PJ teleconnections to ISM rainfall and in particular circulation anomalies are overall stable from one epoch to another. Note that changes of the quality of data can lead to changes in correlations epoch by epoch. But the spatial pattern is stable, which implies the robustness of the PJ influence on the ISM. This suggests that the PJ pattern has a stable impact on the ISM, unlike ENSO teleconnections.

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