How Robust is the Asian Precipitation–ENSO Relationship during the Industrial Warming Period (1901–2017)?

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ABSTRACT

Instrumental observations (1901–2017) are used to uncover the seasonality, regionality, spatial–temporal coherency, and secular change of the relationship between El Niño–Southern Oscillation (ENSO) and Asian precipitation (AP). We find an abrupt seasonal reversal of the AP–ENSO relationship occurring from October to November in a large area of Asia north of 20°N due to a rapid northward shift of the ENSO-induced subsidence from Indonesia to the Philippines. We identified six subregions that have significant correlations with ENSO over the past 116 years with $|r| > 0.5$ ($p < 0.001$). Regardless of the prominent subregional differences, the total amount of AP during a monsoon year (from May to the next April) shows a robust response to ENSO with $r = -0.86$ (1901–2017), implying a 4.5% decrease in the total Asian precipitation for 1° of SST increase in the equatorial central Pacific. Rainfall in tropical Asia (Maritime Continent, Southeast Asia, and India) shows a stable relationship with ENSO with significant 31-yr running correlation coefficients (CCs). However, precipitation in North China, the East Asian winter monsoon front zone, and arid central Asia exhibit unstable relationships with ENSO. Since the 1950s, the AP–ENSO relationships have been enhanced in all subregions except over India. A major factor that determines the increasing trends of the AP–ENSO relationship is the increasing ENSO amplitude. Notably, the AP response is asymmetric with respect to El Niño and La Niña and markedly different between the major and minor ENSO events. The results provide guidance for seasonal prediction and a metric for assessment of climate models’ capability to reproduce the Asian hydroclimate response to ENSO and projected future change.

1. Introduction

Investigation of the interannual variability of Asian precipitation (AP) and El Niño–Southern Oscillation (ENSO), driven by demands of seasonal prediction, has nearly a century-long history. Walker (1924, 1928) first tied Indian monsoon variability to a global-scale east–west seesaw oscillation in sea level pressure that was coined the “Southern Oscillation” by Walker and Bliss (1932). This statistical linkage was not understood physically until Bjerknes (1966, 1969) developed a coupled

atmosphere–ocean theory linking the Southern Oscillation to El Niño. Since the 1980s the term “ENSO” has been widely adopted and the impact of ENSO on regional and global precipitation has been studied extensively, mainly because the variability of AP has an immense impact on the Asian economy and water and food security, and because the seasonal prediction of AP has attracted enormous attention in the climate research community (Webster et al. 1998).

Figure 1 shows a canonical picture of the impact of El Niño on global precipitation published by the International Research Institute for Climate and Society (IRI) website (<http://iri.columbia.edu/our-expertise/climate/enso>), which is based on the analysis of historical El Niño events carried out by Ropelewski and Halpert (1987) and Mason and Goddard (2001). Over Asia, an El Niño

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FIG. 1. Canonical picture of the impact of El Niño on global precipitation (adopted from the International Research Institute for Climate and Society website, <http://iri.columbia.edu/our-expertise/climate/enso>).

event induces dry anomalies in the Maritime Continent (MC) from June to the following January (Haylock and McBride 2001) due to the suppressed convection over the MC caused by an eastward shift of the Walker circulation (Webster and Yang 1992; Palmer et al. 1992). El Niño also induces deficient Indian rainfall from June to September (Sikka 1980; Shukla and Paolino 1983; Rasmusson and Carpenter 1982) due to the descending Rossby wave response to the suppressed MC heating (Wang et al. 2003), which is further modulated by air–sea interaction in the northern Indian Ocean (Lau and Nath 2000). Notably, the signal of El Niño impacts over East Asia (EA) is absent in Fig. 1. However, mounting evidence has been shown on the impact of ENSO on East Asian monsoon (e.g., Tao and Chen 1987; Kang and Jeong 1996; Zhang et al. 1996), and the details can be found in a recent review paper by Yang et al. (2018). It remains an issue regarding whether or not El Niño has a significant impact on EA rainfall.

It should be noted that Asia has an immense land area of over 44 million square km and ranges from 10°S to 75°N and from 40° to 180°E; the AP–ENSO relationship has been studied mostly on a regional scale, not as a whole, and the spatial and temporal coherency of the AP–ENSO relationship and the response of the total amount of Asian precipitation to ENSO remain a gap in our knowledge.

The nonstationarity of the AP–ENSO relationship has been described as a specter for seasonal prediction (Webster et al. 1998). The Indian summer monsoon–ENSO relationship has undergone significant periods of waxing and waning (Troup 1965; Mooley and Parthasarathy 1984; Wang et al. 2015). The interdecadal variability of the Indian summer monsoon (ISM)–ENSO relationship has been attributed to changes in the equatorial Walker

circulation and local meridional circulation over Indian Ocean, which are associated with interdecadal Pacific oscillation variability (Goswami 2006). Since the late 1970s, the ISM–ENSO relationship has weakened (Kumar et al. 1999; Torrence and Webster 1999). Meanwhile, the relationships between ENSO and EA summer and winter precipitation have also experienced significant changes (Wu and Wang 2002; Yu et al. 2004). On a large scale, the overall coupling between the Asian–Australian monsoon system and ENSO has strengthened because the relationships between ENSO and the western North Pacific, East Asian, and Indonesian monsoons are enhanced, overriding the weakening ISM–ENSO relationship (Wang et al. 2008). Analysis of 544-yr reconstructed Asian precipitation (RAP) proxy data (Shi et al. 2018) suggests that during the ENSO developing summer, the AP–ENSO relationship has been largely steady since 1620; however, during the ENSO decaying summer, the summer rainfall–ENSO relationship over the Yangtze River valley–southern East China has gone through large-amplitude multidecadal-to-centennial changes over the past five centuries (Shi and Wang 2019). The causes of such long-term changes have not been well understood, although they are linked to the Pacific decadal oscillation (Mantua et al. 1997) as reflected by proxy data.

Since 1900, the global mean temperature has risen by about 1.0°C largely due to increased greenhouse gases concentration (Trenberth and Fasullo 2013). Anthropogenic aerosols have also rapidly increased after 1980 over Asia. Given such significant environmental changes, the climate prediction community faces a critical issue—has the predictability of Asian rainfall changed? ENSO has been known as the strongest large-scale forcing for interannual variability of precipitation worldwide, especially over the global land monsoon regions (Wang et al. 2012, 2017). Therefore, it is

important to find out whether the relationship between the AP and ENSO has changed during the industrial warming period. Understanding the causes of this change is instrumental to understanding future changes under increasing anthropogenic forcing.

This study will specifically address the following questions: (i) How does the AP–ENSO correlation vary with season? (ii) What are the linkages and differences in the AP–ENSO relationship among different subregions? (iii) How does the total amount of Asian precipitation respond to ENSO? (iv) How has the AP–ENSO relationship changed with the 1°C global warming since 1900? (v) What caused the changes of AP–ENSO relationship? We will use instrumental observations over Asia during the period from 1901 to 2017 (section 2) to detect robust signals of the impact of ENSO on AP. In section 3 we examine the seasonal dependence and identify different temporal and spatial regimes of the AP–ENSO relationship. An abrupt change in the AP–ENSO relationship from October to November is found. In section 4 we study the spatial coherency and regionality of AP–ENSO relationship and propose a set of regional and integrated AP indices. Section 5 discusses nonlinear aspects of the AP–ENSO relationship. In section 6 we examine secular changes in the AP–ENSO relationship over Asia and its subregions and explore the causes of nonstationary AP–ENSO relationship. Section 7 presents a summary and discussion.

2. Data and methods

The monthly mean land precipitation from the Global Precipitation Climatology Centre (GPCC) dataset (Schneider et al. 2014) and monthly precipitation data from the Climatic Research Unit (CRU) at the University of East Anglia version 4 (Harris et al. 2014) were merged by taking an arithmetic mean of the two sets of precipitation data for the period from 1901 to 2017 with a spatial resolution of $1^\circ \times 1^\circ$. The Met Office Hadley Centre's sea surface temperature (SST) (Rayner et al. 2003) and the National Oceanic and Atmospheric Administration (NOAA) Extended Reconstructed SST, version 5 (Huang et al. 2017), were used in this study. The monthly mean SST data were obtained by averaging the two SST datasets for the period from 1901 to 2018. The circulation data used in this study is made by merging three ECMWF reanalysis datasets from the ERA-20C (1901–57) (Poli et al. 2016), ERA-40 (1958–2001) (Uppala et al. 2005), and ERA-Interim (2002–18) (Dee et al. 2011). To ensure temporal consistency, the climatological monthly means of the three datasets during the common period from 1979 to 2001 were computed, respectively, and then the ERA-40 and ERA-Interim

FIG. 2. Time series of DJF(0/1) oceanic Niño index (ONI), namely, the Niño-3.4 SST anomalies normalized by the standard deviation ($SD = 1.0^\circ\text{C}$). Years with absolute value of ONI greater than 1.0 (between 0.5 and 1.0) are considered major (minor) ENSO events. The threshold of ± 1.0 and ± 0.5 are marked by blue dashed lines. Red solid dots (empty circles) mark the 19 major (20 minor) El Niño events, and blue solid dots (empty circles) mark the 18 major (18 minor) La Niña events.

reanalysis data were calibrated by removing their climatological differences from ERA-20C dataset.

The oceanic Niño index (ONI) defined by three-month running mean SST anomalies in the Niño-3.4 region (5°N – 5°S , 120° – 170°W), which is also known as the Niño-3.4 index, is used to measure the intensity of ENSO. Because ENSO normally matures toward the end of the calendar year (Rasmusson and Carpenter 1982), we designate the ENSO developing year as year (0), the following year as year (1), and use the December (0)–January (1)–February (1) mean (denoted as DJF) ONI to identify ENSO events. Figure 2 shows the time series of DJF ONI. Events with DJF ONI greater than 0.5°C (less than -0.5°C) are classified as an El Niño (La Niña). When we compare these with the ENSO events defined by the Climate Prediction Center (CPC), the 1951 and 1953 El Niño events and 1974, 1983, and 2016 La Niña events are missing. The missing events are all marginal events and/or do not have typical phase-locking features (e.g., maturity during DJF). The differences arise from slightly different definition: the CPC definition used the SST anomaly as the departure from a 30-yr running climatology and the El Niño (La Niña) year is defined when ONI is above (below) 0.5°C (-0.5°C) in any five consecutive overlapping months.

The Student's t test is used to make statistically significant tests for correlation coefficients and composite anomalies. A 31-yr running correlation coefficient is calculated to examine the stationarity of AP–ENSO correlations. An F test (Snedecor and Cochran 1989) is used to test if the variances between two populations are significantly different. In this study, the F test is used to test the significance of the root-mean-square errors of the regressed API between El Niño and La Niña events.

FIG. 3. Seasonal evolution of the AP-ENSO relationship. Color shading indicates values of correlation coefficients (CCs) between the DJF ONI and bimonthly precipitation anomalies from MJ(0) to MA(1) for the period 1901–2017. Dots indicate the areas where CCs are statistically significant above 95% confidence level. The thick black solid lines represent the Tibetan Plateau above 3000 m.

3. An abrupt change of the AP-ENSO relationship from October to November

The monsoon annual cycle plays a critical role in modulating atmospheric response to remote ENSO forcing and the warm pool atmosphere–ocean interaction (Wang et al. 2003). To detect detailed seasonal and phase dependences of the AP response to ENSO, we examine the correlation maps of bimonthly mean AP anomalies and DJF ONI from May–June (0) [hereafter MJ(0)] to March–April (1) [hereafter MA(1)] of the following year (Fig. 3). The bimonthly mean removes potential impacts from the tropical intraseasonal oscillation while preserving the anomalous signals associated with ENSO.

During northern summer from MJ(0) to September–October (0) [SO(0)], the rainfall in most regions of Asia (east of 70°E) are negatively correlated with DJF ONI. A negative correlation means dry anomalies during an El Niño and/or wet anomalies during an La Niña. The negative correlations strengthen from MJ(0) to July–August (0) [JA(0)] and reach their highest correlations in SO(0), especially over central India, northern China, the Philippines, and northern Australia–Papua New Guinea. The increasing negative correlation is arguably a result of the increasing intensity of ENSO forcing (Fig. 4). As shown in Fig. 4, from MJ(0) to SO(0), the El Niño induces suppressed convection in the MC, and the divergent easterly anomalies over the Indian Ocean and westerly

FIG. 4. Bimonthly fields regressed on the DJF ONI (1901–2017). Regressed fields are precipitation anomalies over land (mm month^{-1}), SST anomalies over ocean (in units of $^{\circ}\text{C}$), and 850 hPa wind anomalies (arrows) in units of m s^{-1} . Red markers in (c) and (d) indicate the suppressed convection centers determined by the descending centers in the 500 hPa vertical velocity field. Dots denote the area where the regression coefficients are significant above 95% confidence level. The thick black solid lines represent the Tibetan Plateau above 3000 m.

anomalies over the equatorial western Pacific are clearly seen in SO(0) (Fig. 4c). The strengths of the zonal wind anomalies increase with increasing amplitude of the warming from MJ(0) to SO(0). The easterly anomalies over the Indian Ocean tend to weaken the southwesterly Indian monsoon. The anomalous westerlies over the western Pacific and associated meridional wind shear tend to generate cyclonic vorticity in off-equatorial regions, which weaken the western Pacific subtropical high over the Philippine Sea, and the moisture transport to EA monsoon region.

An abrupt reversal in correlation signs occurs from SO(0) to November–December (0) [ND(0)] over Asia except for tropical Southeast Asia (Fig. 3). Since ND(0), a strong positive correlation region emerges along the winter subtropical frontal zone extending from southern China to western Japan, which persists until the following

April. This visually discernible sudden change in the AP–ENSO relationship is further confirmed by the pattern correlation coefficients (PCCs) between any two consecutive panels in Fig. 3 (Table 1). The significantly positive correlation coefficients (CCs) indicate that each of the two consecutive bimonthly means have similar correlation patterns, except for the one between SO(0) and ND(0), which is negative, indicating a reversal of AP–ENSO relationship from October to November. The reversal of the correlation pattern occurs over a large-scale Asian continent north of 20°N and east of 70°E . There is also a footprint of sudden reduction of the negative correlation over the MC (Fig. 3). From SO(0) to ND(0), the negative correlation in the western MC (Sumatra and southwest Borneo) turns into a weak positive, and the negative correlation between the total rainfall over the entire MC and ENSO drops notably after ND(0), as

TABLE 1. Pattern correlation coefficients (PCCs) between each of the two consecutive panels in Fig. 2. The PCCs are calculated over the Asian region north of 15°N and east of 70°E . All PCCs shown in this table are statistically significant at the 95% confidence level.

	MJ(0)–JA(0)	JA(0)–SO(0)	SO(0)–ND(0)	ND(0)–JF(1)	JF(1)–MA(1)	MA(1)–MJ(1)
$15^{\circ}\text{N}, 70^{\circ}\text{E}$	0.22	0.38	−0.31	0.60	0.51	0.44

FIG. 5. Differences in precipitation anomalies between the composite major El Niño and La Niña events (color shading) during (a) northern summer MJJASO(0) and (b) northern winter NDJFMA(0/1). Blue boxes in (a) outline the subregions of MC, ISM, and NCH, and those in (b) outline the subregions of SEA, EAFZ, and CenA, respectively. Red lines outline the monsoon precipitation domains. Dots denote the area where the composite differences are significant above 95% confidence level.

previously found by Haylock and McBride (2001) and Chang et al. (2004).

What caused the October-to-November abrupt change of AP–ENSO correlation pattern? We note that the anomalous precipitation pattern change is a result of the sudden establishment of the anomalous Philippine anticyclone and associated northward shift of the ENSO-induced subsidence from Indonesia to the Philippines archipelago (Fig. 4d). During SO(0), the descending motion is centered over Indonesia, but in ND(0) it weakens and shifts to the Philippines. This anomalous Philippines anticyclone increases precipitation over EA and southern India, especially along the subtropical front zone from southern China to western Japan after November (0). The sudden establishment of the Philippine Sea anomalous anticyclone and northward shift of the suppressed convection weaken the ENSO impact over the western MC. This provides an alternative explanation to the decreased MC precipitation–ENSO relationship before the peak phase of ENSO.

What causes the sudden establishment of the anomalous anticyclone over the Philippine Sea (hereafter Philippine Sea anticyclone)? During the late fall of major El Niño developing years, the onset of the EA winter monsoon generates a strong anomalous anticyclonic flow intruding into the Philippine Sea, triggering the inception of the Philippine Sea anticyclone (Wang and Zhang 2002). To the east of the anticyclone the sea surface cooling (Fig. 4) suppresses convective heating, exciting westward-propagating, descending Rossby waves to strengthen the anticyclone; meanwhile the anticyclone enhances surface cooling by increased evaporation and turbulent mixing via increasing the total wind speed; the sea surface cooling and the dry advection caused by the

northerly wind further suppress the convection to the east of the anticyclone (Wang et al. 2000). This positive air–sea thermodynamic feedback process stabilizes and maintains the Philippine Sea anticyclonic anomaly from winter to the ensuing summer against decaying ENSO forcing (Wang et al. 2000; Lau and Nath 2006; Xiang et al. 2013). The Philippine Sea anticyclone, after its establishment, dominates the Asian anomalous circulation during El Niño winter till the following spring (Figs. 4d–f), increasing precipitation over EA to the northwest of the anomalous anticyclone and decreasing precipitation over the Philippines and Indochina Peninsula (Figs. 3, 4d–f).

4. Regional and integrated AP responses to ENSO

The sudden change of the AP–ENSO relationship from October to November provides a natural division between the northern summer and winter regimes with regard to the AP–ENSO relationship (Fig. 5). During the northern summer from May(0) to October(0), both the tropical and extratropical Asia are dominated by dry anomalies except in western central Asia and northeast tip of China (Fig. 5a). Among all the subregions in Asia, the MC has the strongest dry signal (Fig. 3). The western-central India and northern China (NCH) also exhibit prominent rainfall reduction. In contrast, during the northern winter [November(0) to April(1)], the subtropical and midlatitude continental Asia between 20° and 50°N is predominated by wet anomalies (Fig. 5b). The most significant wet anomalies are found over the EA subtropical front zone stretching from southeast China to the entire Korean Peninsula and western Japan.

FIG. 6. Bimonthly correlation coefficients between DJF(0/1) ONI and six subregional precipitation indices from MJ(0) to JA(1). The subregional indices are the area-weighted mean precipitation rate. The black dashed lines indicate CCs that are significant at 95% confidence level ($CC = \pm 0.19$ for 1901–2017 period).

From Fig. 5 we can identify six subregions with the most significant rainfall response to ENSO. The domains of each subregion are determined based on the areas with significant land precipitation anomaly signals in composite difference field between major El Niño and La Niña events during summer (Fig. 5a) and winter (Fig. 5b). The six subregions include (i) MC (10°S – 5°N , 100° – 150°E); (ii) Indian summer monsoon (ISM: 12° – 33°N , 70° – 90°E); (iii) northern China (NCH: 30° – 40°N , 100° – 115°E); (iv) EA winter front zone (EAFZ: 20° – 30°N , 110° – 120°E ; and 30° – 35°N , 120° – 140°E , including southern China and western Japan); (v) Southeast Asia (SEA: 5° – 20°N , 95° – 130°E , including Indochina, the Philippines, and northern Borneo); and (vi) central Asia (CenA: 35° – 50°N , 55° – 85°E). The precipitation–ENSO connection in SEA and CenA regions and the seasonality in each subregion are rarely discussed.

Precipitation anomalies in each subregion have their own phase dependence on the ENSO life cycle (Fig. 6). To find out the specific months during which the rainfall anomalies respond to ENSO most robustly for each subregion, Fig. 6 shows that the most persistent negative correlations occur over the MC and SEA. Over the MC, significant negative correlation persists for 8 months from MJ(0) to ND(0) with $r = -0.81$ (Table 2). Over the SEA, the most significant negative correlation also persists for 8 months but from SO(0) to MA(1) with $r = -0.79$ (Table 2). Thus, the El Niño-related dry anomaly center occurs over the MC first and then shifts

northward to the SEA four months later. Another notable feature is the in-phase anomalous rainfall responses to ENSO over the ISM and NCH with July–October (0) [JASO(0)] as a peak negative correlation season during which the ISM and NCH summer rainfall correlated with DJF ONI with $r = -0.59$ and -0.50 , respectively (Table 2). This in-phase relationship during summer (Kripalani and Kulkarni 2001) is partially related to the Silk Road teleconnection (Enomoto et al. 2003), which is part of the circumglobal teleconnection (Ding and Wang 2005). The ENSO-induced anomalous rainfall and condensational heating over India can modulate the Silk Road teleconnection, which further affects central North China rainfall through downstream propagation of the Rossby wave train. The detailed mechanisms were discussed in Wang et al. (2017). The EAFZ and CenA winter precipitation is positively correlated with the DJF ONI with $r = 0.60$ and $r = 0.50$, respectively (Table 2). The CenA precipitation exhibits the longest persistent positive correlation with DJF ONI from SO(0) to JA(1), lasting for 12 months (Fig. 6).

The coherent variations among regional precipitation imply that we can further examine the integrated Asian precipitation response to ENSO. During El Niño developing summer from May(0) to October(0), it is meaningful to propose an Asia-wide summer [May–October (MJJASO)] precipitation index (ASPI), which is defined by the area-weighted average precipitation over the vast land area of Asia from 10°S – 50°N and from 70° to 150°E . During winter [November–April (NDJFMA)], a positive correlation dominates continental Asia north of 20°N (20° – 50°N , 50° – 150°E), so an area-weighted mean precipitation index over this region can be proposed to represent the extratropical continental Asian winter precipitation, which, for simplicity, is named the extratropical Asian winter precipitation index (AWPI). The ASPI is negatively correlated with DJF ONI with $r = -0.82$ ($p < 0.001$) (Fig. 7a), while the AWPI is positively correlated with ONI with a lower $r = 0.62$ ($p < 0.001$) (Fig. 7b).

How does the total amount of Asian precipitation respond to ENSO? Instead of following the conventional calendar year, we examine the yearly mean precipitation accumulated within a “monsoon year” (Yasunari 1991), which starts from May and ends the following April. The

TABLE 2. Correlation coefficients (CCs) between the DJF ONI and each of subregional precipitation indices for the period from 1901 to 2017. Bold numbers denote the correlation coefficients significant at the 99% confidence level. The first row indicates the period during which the CCs are calculated. The second row shows the six subregions, which are outlined in Fig. 5.

MJJASO(0)		NDJFMA(0/1)		MJJASOND(0)		SONDJFMA(0/1)	
ISM	NCH	EAFZ	CenA	MC	SEA		
-0.59	-0.50	0.60	0.50	-0.81	-0.79

FIG. 7. Scattering diagrams showing the relationships between (a) boreal summer MJJASO(0) ONI and ASPI, (b) boreal winter NDJFMA(0/1) ONI and AWPI, and (c) the DJF ONI and the monsoon-year API. (d) Differences in monsoon-year precipitation anomalies between composite major El Niño and major La Niña events. The middle tilted solid line in (a)–(c) is the regression line between the rainfall index and ONI, respectively, along with the lines of the regressed value ± 1 root-mean-square error on the two sides. The red (blue) dots denote years before (after) 1950. The red line in (d) outlines the monsoon precipitation domain over Asia–Australia.

monsoon year is suitable for study of the impacts of ENSO on global climate, including tropical cyclones and global monsoons (Wang et al. 2010, 2012). The dry AP anomaly prevails during a monsoon year except in the southeast China and central Asia (Fig. 7d). Given this nearly uniform spatial structure, we define a monsoon-year AP index (API) by area-averaged monsoon yearly rainfall over Asia in the domain of (10°S – 50°N , 70° – 150°E). The monsoon-year API is highly correlated with the DJF ONI ($r = -0.86$, $p < 0.001$) (Fig. 7c). On average, the amount of monsoon-year Asian precipitation decreases by 4.25% for every degree increase of

ONI. The results shown in Fig. 7 suggest that the Asian precipitation–ENSO relationship is more robust when larger spatial scales and longer time scales are considered. The Asian precipitation–ENSO relationship is largely determined by the tropical rainfall.

5. Nonlinearity of the AP response to ENSO

Figure 7c shows that the regressed API has 13 outliers [the absolute value of error exceeds the root-mean-square error (RMSE)] for 35 La Niña events, but 8 outliers for the 37 El Niño events, which imply

FIG. 8. Comparison of the impact between major and minor ENSO events. Color shading shows composite precipitation anomalies induced by (a) major El Niño, (b) major La Niña, (c) minor El Niño, and (d) minor La Niña events. The upper and lower panels are for JASO(0) and NDJFM(0/1) seasons, respectively. The major and minor El Niño/La Niña events are shown in Fig. 2.

potentially larger errors in forecasting AP during La Niña events than in El Niño events. The RMSE of the regressed API is $1.9 \text{ mm month}^{-1}$ for El Niño years, while it is $3.1 \text{ mm month}^{-1}$ for La Niña years. The F -test results indicate that the difference in RMSEs between El Niño and La Niña is significant at the 99% confidence level, indicating a significant asymmetry in API response to El Niño and La Niña (Fig. 7c).

Figures 8a and 8b further compare the composite AP anomalies associated with 19 major El Niño and 18 major La Niña events. During JASO (0), major El Niño events induce significant dry anomalies over northern China but the major La Niña events do not induce significant wet anomalies there. During northern winter (NDJFM), the wet anomalies induced by major El Niño events occur over the EA winter front zone, but the dry anomalies induced by major La Niña events tend to shift northward. These results indicate that the EA summer and winter precipitation responds to major El Niño and La Niña asymmetrically. Precipitation over the rest of the Asian regions, on the other hand, shows nearly symmetric response to El Niño and La Niña, except for northern India.

Figure 8 also shows the AP responds to major and minor ENSO forcing differently. For major El Niño events (Fig. 8a), the spatial pattern of precipitation anomalies is basically the same as those depicted by the correlation analysis (Fig. 3). However, for minor El Niño events (Fig. 8c), significant anomalies are only seen over the MC/Philippines and some scattered regions in the Indian and EA monsoon regions. This conclusion is also generally true for the differences between the major and minor La Niña events except over northern China.

This asymmetry between the major and minor ENSO events suggests that during weak ENSO events the monsoon prediction is less reliable if the AP–ENSO relationship derived from the regression analysis is used.

6. Secular changes of the AP–ENSO relationship

To detect the secular change of the AP–ENSO relationship, we examine 31-yr sliding correlation coefficients. We describe the stability of the AP–ENSO relationship as robust (stable) if the 31-yr running mean CCs are persistently significant above 99% (95%) confidence level across the entire period of 1901–2017; otherwise, the relationship will be described as unstable.

On regional scales, the rainfall over three tropical subregions of Asia shows a robust relationship with ENSO, although the correlation coefficients show multi-decadal (ISM) or centennial (MC and SEA) fluctuations (Fig. 9a). The MC–ENSO and SEA–ENSO relationships have similar centennial variation with a common epoch of waning relationship centered around 1950s. After the late 1970s, the MC–ENSO and SEA–ENSO relationships are notably enhanced ($r < -0.8$) while the ISM–ENSO relationship has been weakened. The ISM has significant correlation with ENSO throughout the entire period, which varies between -0.45 and -0.7 without breakdowns. In contrast to tropical precipitation, the precipitation in subtropical and midlatitude Asia show unstable relationships with ENSO (Fig. 9b). The absolute values of the 31-yr running mean correlation coefficients range from 0.2 to 0.8. During summer, the negative NCH–ENSO relationship shows a breakdown in the 1930s and was gradually enhanced afterward. In the winter regime, both the

FIG. 9. The 31-yr running correlation coefficients (CCs) between DJF ONI and (a) precipitation indices in three tropical subregions, (b) precipitation indices in three subtropical and extratropical subregions, and (c) three integrated AP indices, API (green), ASPI (red), and AWPI (blue), respectively. For comparison, the 31-yr running ONI standard deviation (black) is also shown in (c). Thick (thin) dashed lines indicate the 99% (95%) significance level for a 31-yr running CC.

positive EAFZ–ENSO and CenA–ENSO relationships also show a sharp breakdown period in the 1930s. Therefore, the precipitation–ENSO relationships in the three subtropical and midlatitude Asian regions exhibit a coordinated (in phase) centennial variation with a common breakdown period in the 1930s. The breakdown in 1930s corresponds to an epoch during which ENSO has the smallest amplitude during the past 117 years (Figs. 2 and 9c).

Figure 9c shows that the Asian summer precipitation (ASPI) and Asian monsoon-year precipitation (API) have robust relationships with ENSO, while the subtropical–extratropical Asian winter precipitation (AWPI) has a stable relationship with ENSO. The ASPI–ENSO and

AWPI–ENSO correlations tend to vary coherently, and both show the weakest epoch from 1920 to 1950. After 1950, the three relationships have been enhanced, with $r = -0.85$, -0.86 , and 0.70 for the API, ASPI, and AWPI, respectively. Therefore, over the past six decades, ENSO has become an exquisite source of predictability for the large-scale Asian continental precipitation.

The centennial variations of the ASPI–ENSO and AWPI–ENSO relationships are arguably related to the ENSO amplitude variation, as evidenced by the time series of the ONI standard deviation (SD) shown in Fig. 9c. The correlation coefficient between the ONI running SD and the running CC of ASPI and ENSO is -0.92 ($p < 0.05$), which is statistically significant at 95% confidence level considering the reduced degree of freedom due to the 31-yr running mean. Similarly, the correlation coefficient between the ONI running SD and the corresponding running CC of AWPI and ENSO is 0.92 ($p < 0.05$). These results suggest that ENSO amplitude is likely a major contributor to the secular change of the AP–ENSO relationship.

7. Conclusions and discussion

a. Major findings

- (i) A reversal in the sign of AP–ENSO correlation from October to November is revealed in a large region of Asia (north of 20°N and 70° – 140°E) (Fig. 3 and Table 1) that separates the AP–ENSO teleconnection into distinct boreal summer and winter regimes. The October-to-November abrupt change is caused by a sudden northward shift of the ENSO-induced subsidence from Indonesia to the Philippines and associated establishment of the Philippine Sea anticyclonic anomalies (Fig. 4).
- (ii) Six subregions that have long-term significant ($p < 0.01$) teleconnections with ENSO are identified (Figs. 5 and 6) and summarized in Fig. 10. The rainfall in the MC and SEA both have a 10-month persistent negative correlation with DJF ONI ($r = -0.81$ and -0.84 , respectively). India, northern China, subtropical EA, and central Asia are correlated to DJF ONI with the absolute values of correlation coefficients ranging from 0.50 to 0.60 during the period of 1901–2017 (Table 2). Figure 10 provides a forecast guide for seasonal prediction.
- (iii) The AP–ENSO relationship is stronger on a larger spatial scale and longer seasonal time scale due to the spatial coherent variations among subregions. During the northern summer (MJJASO), nearly all of Asia is dominated by negative correlations (Fig. 5a) and thus the total area-weighted Asian summer precipitation index (ASPI) has a high correlation with

FIG. 10. Schematic picture showing the impact of El Niño on Asian precipitation; r denotes the correlation coefficient between DJF ONI and each regionally averaged precipitation anomalies during the corresponding marked period; 0 (1) indicates ENSO developing (decaying) year.

DJF ONI ($r = -0.82$) during 1901–2017 (Fig. 7a). In contrast, during winter [from November (0) to April (1)], the rainfall response in extratropical continental Asia poleward of 20°N is replaced by dominant positive correlations (Figs. 5b and 7b). During a monsoon year from May to the following April, the total area-weighted average precipitation over Asia (10°S–50°N, 70°–150°E) is highly correlated with the DJF ONI ($r = -0.86$, $p < 0.001$) (Fig. 7c). On average, during a monsoon year, the total amount of precipitation over Asia decreases by about 4.3% for every degree of ONI increase.

- (iv) It is important to distinguish major and minor ENSO events for application of the AP prediction guide shown in Figs. 5 and 10. Minor ENSO events show little significant signal over continental monsoon Asia (Figs. 8c,d). The AP response to ENSO also exhibits an asymmetry between El Niño and La Niña events, mainly over the EA region (Figs. 8a,b). The AP–La Niña relationship is more variable than the AP–El Niño relationship, especially for minor La Niña events (Fig. 7).
- (v) The rainfall variations in three tropical subregions of Asia (MC, SEA, and ISM) show a robust and stable relationship with ENSO with the 31-yr running mean CCs being persistently significant above the 99% (95%) confidence level across the entire period of 1901–2017. On the other hand, the precipitation in subtropical and midlatitude Asia (NCH, EAFZ, and CenA) show unstable relationships with ENSO (Figs. 9a,b), which exhibit a coordinated

centennial variation with a common breakdown period around the 1930s when the ENSO signals are the weakest. The Asia-wide precipitation anomalies measured by the ASPI and API have robust relationship with ENSO and the centennial changes of their relationship are likely modulated by the ENSO amplitude (Fig. 9c).

b. Discussion

The East Asian summer and winter precipitation, namely, the North China (NCH) and EA winter front zone (EAFZ) precipitations have significant correlations with ENSO during 1901–2017, but these significant correlations were absent in the teleconnection maps shown in Fig. 1. There are several plausible reasons. First, many previous studies used relatively short records and the multidecadal variation of the EA–ENSO relationship may obscure the correlation derived from the short records. Second, the simultaneous correlation between EA summer rainfall and SST anomalies in the eastern equatorial Pacific is not significant (Chen et al. 1992) because the EA monsoon–ENSO relationship is critically dependent on the phase of ENSO. During the El Niño developing summer, the northern EA summer monsoon tends to weaken (Guo 1987; Wang and Li 1990; Yatagai and Yasunari 1995); but during the El Niño decaying summer, the mei-yu-baiu frontal rainfall tends to increase (Lau and Sheu 1988; Fu and Teng 1988; Huang and Wu 1989; Chang et al. 2000; Yim et al. 2016; Xing et al. 2016) owing to the enhanced western North Pacific (WNP) subtropical high and its interaction with the underlying Indo-Pacific warm pool (Wang et al. 2000, 2013; Xie et al. 2009, 2016). For a more detailed review of the EA monsoon–ENSO relationship, readers are referred to Li et al. (2017). Third, the EA rainfall anomaly patterns are quite variable due to its sensitivity to the large fluctuation in ENSO amplitude and its asymmetry between El Niño and La Niña (Fig. 8). Fourth, the seasonal migration of the rainfall anomalies can smear out the conventional seasonal mean precipitation anomalies, thus its relationship with ENSO becomes nonsignificant (Wang et al. 2017).

The large-scale October-to-November abrupt change of the AP–ENSO relationship shows multidecadal variations. Further analysis suggests that the abrupt reversal relationship between SO(0) and ND(0) disappears around 1930, which is likely caused by the weakest ENSO amplitude during this period.

The ISM has significant correlation with ENSO throughout the entire period, with a 31-yr running mean correlation coefficient varying between -0.45 and -0.7

without breakdowns. However, Kumar et al. (1999) found a breakdown of the ISM–ENSO relationship since the late 1970s. This disagreement may be caused by two reasons. First, Kumar et al. (1999) examined JJA rainfall using a 21-yr running mean while we examined JASO rainfall using a 31-yr running mean. Note that the JJA mean Indian rainfall has a weaker relationship with ENSO ($r = -0.42$) than the JASO mean rainfall ($r = -0.56$), because the anticyclonic anomaly over the Indian peninsula, which brings dryness to India, is not fully established in June but it lasts well into October (Fig. 4). Therefore, the Indian rainfall has a weaker response to ENSO in June than in October (Fig. 3), and the JASO is the peak season of the ISM response to ENSO (Fig. 5). Second, this study shows significant correlations in SO(0) dominate west of 81°E in India (Fig. 3), while Kumar et al. (1999) used rainfall over all Indian stations. The western Indian (west of 81°E) rainfall is significantly better correlated with DJF ONI ($r = -0.51$) than the eastern Indian (east of 81°E) rainfall ($r = -0.30$).

While previous studies recognized the positive correlation between the central Asian precipitation and ENSO from January (1) to April (1) (Mason and Goddard 2001), results here indicate that the wet central Asia (CenA) occurs not only in the peak El Niño phase, but also lasts for one year from September (0) to August (1). The reason for the increased precipitation over CenA during an El Niño event deserves further investigation. Besides, it is of interest to find out why the precipitation–ENSO relationship in SEA and CenA tend to be out of phase and both last for ten months from SO(0) to MJ(1). Although the three tropical monsoon regions (MC, SEA, and India) have a common epoch of weakening relationship centered around 1950s, it remains unknown why during the recent three to four decades, the ISM–ENSO relationship is relatively weak, but the MC–ENSO and SEA–ENSO relationships are very strong ($r < -0.8$). The dependence of AP response on different types of El Niño events (Ashok et al. 2007; Yuan and Yang 2012) also deserves further examination using the long-term observed record.

The robust relationship between Asian yearly precipitation and ONI reflects large-scale coherent variability among different subregions. This newly proposed API turns out to be a good measure of AP–ENSO relationship from an annual mean (or monsoon year) perspective. We are interested in the variability of the total amount of Asian precipitation mainly because the anomalous atmospheric circulation is driven by the total amount of latent heating released in the atmosphere, and we want to estimate how much of the total Asian precipitation would reduce for each degree of Niño-3.4 SST warming. This rate of change is attributed to dynamic effects, which is

comparable to the impacts of 1° of global warming induced by anthropogenic change (Wang et al. 2014).

Both GPCP and CRU data are constructed using rain gauge data that are inhomogeneous in the first half of the twentieth century. The number of observations (excluding India) steadily increased from about 30 in the early twentieth century to about 400 in 1950, and after 1950 the number of observations tend to be steady. We find that the data inhomogeneity does not have a significant impact on study of the AP–ENSO relationship. As shown in Fig. 7 the AP–ENSO relationship as measured by the three integrated indices are basically the same for the pre-1950 and post-1950 epochs. The abrupt seasonal reversal of the AP–ENSO relationship is also clearly seen from the pre-1950 data (figure not shown). However, the scarcity of the data in early twentieth century may affect the regional-scale monsoon–ENSO relationship, excluding the Indian monsoon.

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