Central Asia Subtropical Westerly Jet Influences Interannual Variability of Northwest Indian Summer Monsoon rainfall

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Central Asia Subtropical Westerly Jet Influences Interannual Variability of Northwest Indian Summer Monsoon rainfall

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1 Abstract

2 In addition to various factors over the tropics, the interannual variability of northwest
3 Indian summer monsoon (NWISM) rainfall is also regulated by extratropical signals.
4 We defined a subtropical westerly jet index (SWJI) based on the meridional position
5 and intensity of 200-hPa zonal wind within [25-55°N, 40-90°E]. It is found that SWJI
6 exhibits a significant positive correlation with summer rainfall over the NWISM
7 region during 1951-2015. During positive (negative) SWJI years, an upper-level
8 anticyclonic (cyclonic) anomaly over Central Asia along with positive (negative)
9 rainfall anomaly and low-level easterly (westerly) anomalies were observed over the
10 NWISM region. The upper-level anticyclonic (cyclonic) anomaly was accompanied
11 by the descending (ascending) motion and warm (cold) tropospheric temperature
12 anomalies. The anticyclonic (cyclonic) anomaly increased (decreased) the land-ocean
13 thermal contrast by warm (cold) air advection and modified the local meridional
14 circulation. Interannual variability of rainfall over the NWISM region is associated
15 with the meridional position and intensity of the jet that manifest in both upper- and
16 low-level circulation anomalies. Further analysis showed that the interannual
17 variability of SWJI is correlated with Arctic Oscillation (AO). During the positive
18 phase of AO, an upper-level anticyclonic anomaly appeared over Central Asia and
19 favored convection over the NWISM region.

20

21 **Key words:** westerly jet, rainfall, interannual variability

1. Introduction

Interannual variability of the northwest Indian summer monsoon (NWISM) rainfall has received much attention due to its profound influence on agriculture, economy, and the livelihood of billions of people in South Asia (Prasanna 2014). However, the prediction of NWISM interannual variability remains a challenging scientific question (Shukla and Huang 2016). The standard deviation of the interannual variability of the NWISM summer rainfall amounts to approximately 10% of its long-term mean (Gadgil 2003). The interannual variability of NWISM is influenced by various factors. Intraseasonal oscillations (Ajayamohan 2007; Goswami and Mohan 2003), El Nino Southern Oscillation (ENSO; Fasullo and Webster 2002), and Indian Ocean Dipole (Saji et al. 1999) have been found to play important roles in regulating the interannual variability of NWISM. Some studies have focused on land processes such as the impact of snow covers over the Tibetan plateau and Eurasia on the interannual variability of NWISM (Blanford 1884; Hahn and Shukla 1976; Meehl 1994; Wu and Kritman 2003). Of all these factors, the strongest link was between ENSO and summer monsoon rainfall. However, ENSO variations account for less than half of the total variation in summer monsoon (Webster et al. 1998) and this relationship appears to have weakened in recent decades (Kumar et al. 1996, 2006). Another important factor to affect the interannual variations of NWISM is the extratropical large-scale circulation anomalies. These extratropical systems can influence the NWISM via Rossby wave breaking along the upper-level westerly jet (Samanta et al. 2016). The upper-level westerly jet is a planetary scale atmospheric circulation system affecting the weather and climate in East Asia (Zhang et al. 2006; Dong et al. 2010).

The subtropical westerly jet stream is one of the critical systems influencing the summertime weather and climate in China and the Asian-Pacific region with three active centers over the Caspian Sea, the Tibetan Plateau, and the Asian-Pacific, respectively (Yang and Zhang 2008; Schiemann et al. 2009). It has been long known that the seasonal shift of upper-level westerly jet marks an abrupt seasonal transition of the atmospheric circulation regimes in East Asia (Yeh et al. 1959). Numerous researchers have studied the upper-level jet as its southward and northward shift affects the location of rainband over East Asia, with the southward shift causing the rainband moving southward (Lau et al. 1988; Webster and Yang 1992; Lau et al. 2000). The meridional shift of the jet also has a close relationship with Asian

monsoon onset (Lau et al. 2000) and the interannual variability of rainfall in China (Lau et al. 1988; Ding 1992; Zhou and Yu 2005). Lu (2004) found that the southward movement of the jet strengthens convective activities along the rainband over eastern East Asia. It has been well documented that meridional displacement is the most prominent feature of the subtropical westerly jet during the summer season (Lin and Lu 2005; Du et al. 2016). Many previous studies observed a close relationship between the position of upper-level jet and rainfall over China, Japan, and the Korean Peninsula (Laing and Wang 1998; Lau et al. 2000; Lu 2004; Xuan et al. 2011; Li and Lu 2017). Conclusively, the southward shift of the jet during summer indicates that it will be a rich Meiyu year, and vice versa. This relationship between the position of the jet and rainfall can be considered as an impact of the extratropical circulation anomalies on the East Asian summer monsoon.

Studies have suggested that the seasonal transition of the East Asian monsoon has a close relationship with changes in the position and intensity of the subtropical jet over East Asia. As these studies were focused on East Asia, less attention has been paid to the connection between subtropical westerly jet and monsoon rainfall over the Indian subcontinent. As the northern part of the Indian subcontinent is situated at the south of the upper-tropospheric subtropical westerly jet, this region experiences complex climate variability in summer (Lu et al. 2020). Since the SWJ has three different core regions and the one over East Asia has a large influence there. We may expect the core region over central Asia located between 40 and 80°E to affect NWISM, different from the one over East Asia which has its maximum between 80 and 130°E. A more detailed analysis is thus needed to investigate the interannual variability of the jet and the NWISM rainfall. This work aims to analyze how the position and intensity of the upper-tropospheric westerly jet modulate the mean summer rainfall over the NWISM region. EOF analysis is employed to construct an index based on the position and intensity of SWJ hereafter SWJI. We examine the correlation between the SWJI and NWISM rainfall to explore their possible connection. Using precipitation and reanalysis data we seek to answer the following questions:

- 1) What large-scale dynamic and thermodynamic patterns are associated with the different phases of SWJI?
- 2) Are there any connections between different modes of climate variability, such as the Arctic Oscillation, and the SWJI?

The datasets and methods used in this study are proposed in section 2. The large-scale anomalies associated with the different phases of SWJI are described in section 3. The relationship between the SWJI and the Arctic Oscillation is discussed in section 4. The main results are summarized in section 5.

2. Data and methods

2.1 Data

Monthly mean wind, geopotential height, and temperature from surface to 100 hPa on $2.5^{\circ} \times 2.5^{\circ}$ grid from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR, Kalnay et al. 1996) for the summer season (June-July-August) during 1951-2015 was used to quantify the large-scale patterns. For moisture transport, we used daily data to compute the monthly mean to reduce the offset of opposite wind vectors.

High-resolution ($0.25^{\circ} \times 0.25^{\circ}$) daily gridded rainfall data from Asian Precipitation-Highly Resolved Observational Data Integration Toward Evaluation (APHRODITE, Yatagai et al. 2012) product V1101 during 1951-2007 and APHRODITE-2 V1801R1 during 2008-2015 were used in this study. The APHRODITE precipitation dataset covers the Monsoon Asia region (-15°S - 55°N , 60°E - 150°E) and has better performance as compared to several other existing data sets over South Asia (Kim et al. 2019). It has been widely used in studies of precipitation characteristics over the NWISM and surrounding regions (Dong et al. 2016; Dong et al. 2017). The anomalies are calculated as the deviations relative to the climatological mean of 1951-2015.

Besides, the Arctic Oscillation (AO) monthly index during 1951-2015 (https://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/monthly.ao_index.b50.current.ascii.table) is adopted to probe its potential linkage to the NWISM rainfall. Summer AO index is computed by averaging the corresponding monthly values in JJA.

2.2 Method

Figure 1 shows the distribution of 200-hPa zonal wind during summer overlaid with the mean summer rainfall over the NWISM region. Two active cores of the westerly jet can be found over the NWISM region with one located around 50°E and the other one around 90°E . Its axis, denoted by the green solid curve, is located

around 40°N. The summertime precipitation features maximum centers over central India and the foothill of the southern Himalayas. The leading modes of 200-hPa zonal wind obtained by EOF analysis within a domain involving the main body of the SWJ (indicated by the red rectangle in Fig. 1) are shown in Fig. 2. The results reported in this study remain almost unchanged when we change the domain used for EOF analysis (e.g., 25-55°N, 0-150°E). As shown in Fig. 2a, the spatial pattern of the first leading EOF mode (EOF1) is characterized by a north-south dipole pattern, with opposite anomalies bounded by 40°N, i.e., the positive and the negative anomalies are almost evenly distributed along the climatological jet axis. Therefore, this dipole pattern indicated by EOF1 and the associated PC1 may reflect the year-to-year meridional displacement of the westerly jet. On the other hand, the second mode of EOF (EOF2) shows a monopole structure centered along the climatological jet position, which together with PC2 reflects the intensity of the jet.

Taking both the intensity and position of the westerly jet into consideration, we define a SWJI as the normalized difference between the regionally averaged 200-hPa zonal winds between a northern area (42-52°N, 45-70°E) and a southern area (28-38°N, 45-70°E) weighted by the area mean 200-hPa zonal winds within the core jet region (34-44°N, 45-70°E).

$$SWJI(t) = \frac{[U_{north}(t) - U_{south}(t)]}{\sum_{t=1951}^{t=2015} U_{core}(t)}$$

By subtracting the mean and dividing by the standard deviation over 1951-2015, the standardized subtropical westerly jet index (SWJI) therefore, represents fluctuations in the subtropical westerly jet's position as well as its intensity.

We then identified years with standardized SWJI greater (smaller) than 0.8(-0.8) as strong positive (negative) SWJI years in this study. As a result, 16 strong positive and 16 strong negative SWJI years are identified (Fig. 3b). Based on the positive and negative SWJI years, composites of winds, geopotential heights, temperature, wave activity flux, and moisture flux anomalies are constructed to study the impact of SWJ on the summer rainfall over the NWISM region. The wave activity flux is calculated following Plumb (1985) to reveal the potential mechanism for the remote impact of AO on its downstream regions. The student's *t* test is applied to assess the statistical significance of results such as correlation and regression analysis.

3 Results

3.1 Impact of the SWJI on rainfall anomalies

Figure 3a shows the correlation between the SWJI and precipitation over South Asia and the surrounding regions. The key feature is the significant positive correlations in the tropics and negative correlations in the extratropical region. It shows a dipole structure as the negative correlation exists north of 37°N and a positive correlation exists to the south of this position. This meridional dipole of significant correlations indicates a close relationship between the meridional displacement and intensity of the upper-level SWJ and summer rainfall over this region. More specifically, a robust correlation can be seen over eastern and central India. The correlation is significant over central and southwestern India as well. It also shows a good correlation with summer rainfall over the whole Pakistan with the northern part showing a greater magnitude. This is also evident in the corresponding time series. As shown in Fig. 3b, the time series of SWJI is strongly correlated with mean precipitation over central India and Pakistan (indicated by the blue rectangle in Fig. 1) with a correlation of 0.66 ($p < 0.001$), suggesting that the SWJI is influential on the summer precipitation over the NWISM region.

On the other hand, significant negative correlation coefficients are observed over Central Asia and Xinjiang. Xinjiang is located in northwest China where the interannual and interdecadal variation of precipitation has different features from that in eastern China, dominated by the East Asian monsoon climate (Zhao et al. 2014). The positive correlation coefficients over the NWISM region and negative correlation coefficients over Central Asia and Xinjiang region show a dipole pattern similar to the EOF1 pattern shown above.

3.2 Upper and lower level circulation anomalies associated with the SWJI

To understand how the SWJ modulates the rainfall over this region, we examined changes in the atmospheric dynamic conditions associated with positive and negative SWJI years. Figure 4a and 4b show the composite of rainfall and 200-hPa circulation anomalies during positive and negative SWJI years, respectively. A remarkable feature in Fig. 4a is the anomalous anticyclonic circulation over the mid-latitude region located to the east of Caspian Sea. This anticyclonic anomaly has a large horizontal scale stretching from the Mediterranean region up to about 85°E. During positive SWJI years, the easterly anomalies mainly appear over Pakistan and

191 northwest India. While for negative SWJI years, on the contrary, there are anomalous
192 westerlies over the NWISM region (Fig. 4b) and an anomalous cyclonic circulation
193 over the mid-latitude region. It is important to note that the anomalous westerlies over
194 west Central Asia extend southward and intrude to the NWISM region.

195 Figure 4c and 4d show circulation anomalies at 700 hPa during positive and
196 negative SWJI years, respectively. It is evident that during positive SWJI years, there
197 is a low-level cyclonic anomaly over the NWISM region. On the contrary, an
198 anticyclonic anomaly is noted during negative SWJI years. It has been reported that
199 the low-level anticyclonic circulation over the NWISM region further diverges the
200 moisture-bearing winds and hence contributes to the reduced summer rainfall over
201 this region (Mishra et al. 2020). The cyclonic anomaly during positive SWJI years
202 indicates a strengthened monsoon trough over the NWISM region, while the
203 anticyclone anomaly during negative SWJI years indicates the weakening of the
204 monsoon trough. Corresponding to the positive precipitation anomaly over Central
205 Asia during negative SWJI years, there is an upper-level large cyclonic anomaly. The
206 anomalous cyclonic circulation over Central Asia results in anomalous prevailing
207 southerlies over this region. On the other hand, anticyclonic anomaly during positive
208 SWJI years over Central Asia results in anomalous northerlies over Central Asia,
209 which prevents warm and wet air penetration from lower latitudes to the Central
210 Asian region.

211 To demonstrate the propagation of the wave activity associated with the
212 changes in the position and intensity of SWJ, we computed the stationary wave
213 activity flux (SWAF) following Plumb (1985). Figure 5 shows the propagation of
214 wave activity overlaid with 200-hPa geopotential height anomalies. The SWAF
215 usually propagates eastward, which is consistent with the characteristics of the wave
216 train described by Hoskins and Karoly (1981). Large SWAF emanates from the
217 Mediterranean Sea region and then propagates both southward and eastward. During
218 the positive SWJI years, a low geopotential height anomaly is present over central
219 Europe and the Rossby wave transmits from the west of this low center to the high
220 center located around Central Asia. A distinct stationary wave pattern is found around
221 40°N, where the wave activity flux vectors point toward the Caspian Sea and Central
222 Asia. The SWAF is larger over the Mediterranean Sea and the Caspian Sea. The wave
223 activity is weaker at the center of the low geopotential height anomaly. However, to
224 the west of this center, the wave is stimulated and propagates downstream and

convergence of wave activity flux can be seen over central Asian high. During the negative SWJI years, central Europe is occupied by a positive geopotential height anomaly. The stationary wave extending from the Mediterranean region is not distinct over Central Asia and the SWAF is weaker. The eastward propagation of SWAF weakens to the east of 50°E. Two branches of wave activity flux emanate from the southwest of this positive height anomaly over central Europe. One branch transports wave activity flux northward and the other branch transports wave activity flux southward to the low geopotential height anomaly over Central Asia. These two branches converge over the central Asian low and an anticyclonic anomaly occurs over the NWISM region. Comparing Fig. 5a and 5b shows that the NWISM region was a wave activity divergent center in positive SWJI years.

Yadav et al. (2018) also noticed that excess monsoonal rainfall over northeast India was accompanied by a negative geopotential height anomaly over northwest Europe and a positive geopotential height anomaly over the Mediterranean region and Central Asia. This positive geopotential height anomaly reduced the intensity of the jet to the south of 40°N. On the contrary, they reported a positive geopotential height anomaly over northeastern Europe and a negative geopotential height anomaly over the Mediterranean region and Central Asia during deficit monsoonal rainfall years. This negative height anomaly over Central Asia increased the intensity of the jet to the south of 40°N and therefore weakened the monsoon over northeast India.

Climatologically, the northward gradient of tropospheric temperatures resulting from higher temperatures over land and lower temperatures over ocean leads to the Indian summer monsoon. Figure 6 shows a latitude-pressure cross-section of anomalous meridional and vertical wind vectors and temperatures along 60-70°E during positive and negative SWJI years. This region passes through the Arabian Sea, the Thar desert region, and the Hindukush-Himalaya mountain ranges. During positive SWJI years, a strong anomalous descent is evident over the NWISM region between 25 and 30°N. Along with the descent, a warm temperature anomaly is located over the Hindukush Himalaya mountain ranges. The warm anomalies persist throughout the whole troposphere with a maximum centered at 300-hPa. These warm anomalies co-occur with the anticyclone (Fig. 4a). Watanabe and Yamazaki (2012) argued that the adiabatic heating strengthened by the anomalous descent causes an increase in temperature at the middle and lower levels, resulting in a heat low over the Thar desert region and adjoining areas. Tropospheric temperature is one of the factors

governing the monsoon over the Indian subcontinent. Liu and Yanai (2001) showed that a good monsoon over the NWISM region is associated with a high tropospheric temperature over the Eurasian continent. Earlier studies have also shown the importance of high tropospheric temperature in the organization of a rainy season over the Indian subcontinent during the summer monsoon (Goswami and Xavier 2005). Therefore, warmer tropospheric temperatures over land will enhance the north-south temperature gradient, resulting in more monsoonal rainfall over the NWISM region. On the contrary, negative tropospheric temperature anomalies are observed during negative SWJI years. These cold anomalies are associated with the southward intruding westerly winds (Fig. 4b). The tropospheric cooling, associated with dry air, suppresses the monsoon by reducing the land-ocean thermal contrast and the convective instability over the NWISM region (Li and Yanai 1996; Krishnan et al. 1998; Krishnan et al. 2009; Samanta et al. 2016).

Figure 7 shows the moisture flux transport and divergence during positive and negative SWJI years integrated from 1000 to 300 hPa. During positive SWJI years, there is a cyclonic anomaly over the Indian sub-continent indicating a strengthened monsoon trough and more water vapor transport to Pakistan as a result. This is further established by the moisture flux divergence, as the negative anomalies which represent inward moisture flux, dominate over the Indian subcontinent. While in the case of negative SWJI years, an anticyclonic anomaly dominates the Indian subcontinent region. Along with this, there is strong outward moisture flux presence over the study region. This is consistent with reduced summer rainfall during negative SWJI years. During positive SWJI years, low-level westerly flow leads to more precipitation over northwest Indian subcontinent. Generally, the moisture produced by evaporation over the northern Arabian Sea and the Bay of Bengal is transported through low-level westerly winds. So, weakened westerlies during negative SWJI years weakens the monsoon. Overall, the atmospheric circulation plays a crucial role in transporting moisture and sustaining water vapor for the formation of rainfall.

4. Discussion

The above analysis examined the connection between interannual variability of NWISM and the SWJI, but what regulates the variability of the SWJI. We devoted particular attention to examining whether Arctic Oscillation (AO) has any influence on the SWJI. AO is an important large-scale global circulation pattern between the

Arctic and the mid-latitude in the Northern Hemisphere. The AO index, defined as the leading empirical orthogonal function of Northern Hemisphere sea level pressure anomalies poleward of 20°N, was characterized by an exchange of atmospheric mass between the Arctic and mid-latitudes (Thompson and Wallace 1998). The correlations between monthly and seasonal AO index and SWJI are computed (Fig. 8). During the spring season, weak but significant correlations are noted between the AO and the SWJI ($r=0.23$, $p<0.05$). These correlations become more robust in June, with the correlation coefficient increased to 0.55 ($p<0.05$).

It is important to know the physical mechanism modulating the SWJI. Using a regression method, we further studied how the AO during the spring/June influences the summer variability of the upper-level westerly jet. The regression of 500-hPa winds and geopotential height with AO-MAM and AO-June shows an upper-level ridge over Central Asia (Fig. 9). During the spring season, it is found that the meridional gradient of the mean tropospheric temperature reduces over northwestern Eurasia and increases over Central Asia (Fig. 10a). As a result, there is a deceleration of zonal wind over Eurasia and enhanced zonal wind over Central Asia (Fig. 11a). The wave activity flux shows that the wave train emanates from the northeast Atlantic to Central Asia. In June, the meridional temperature gradient becomes more pronounced over these regions (Fig. 10b). The reduced meridional gradient of mean tropospheric temperature occurs over north-western Eurasia, Iran, Afghanistan, Saudi Arabia, and central Pakistan. While enhanced meridional temperature gradients are observed over Central Asia. The associated zonal winds decrease over Eurasia, Iran, and Saudi Arabia while an increase is observed over Central Asia, which occurs to the north of the climatological westerly jet position in summer (i.e., 40°N). This indicates that AO influences the position and speed of the jet from its climatological position during summer. The wave activity flux shows that the stronger wave train extending from the North Atlantic is much more pronounced during June.

As the SWJ's position and intensity modulates the mean summer rainfall over the NWISM region, an important question is what is the mechanism that causes this interannual oscillation in the upper-level jet stream during summer. The above results suggest a connection between AO, SWJI, and NWISM.

Our results show that a wave train starts to develop in the spring season. We have observed a weak but significant correlation between the AO index in the pre-monsoon season with SWJI. Previous studies have also observed the relationship

between the wave train and the AO index in June (e.g., Ding and Wang 2005), but no significant correlation was reported in the spring season. So, they concluded that there is some degree of linkage between the AO and the circumglobal wave train, but they did not study the mechanism further. We have found that these seasonal correlations are significant at or above 95% confidence level. The mechanism of the AO and the NWISM rainfall teleconnection is linked through an atmospheric bridge, which causes the changes in SWJ.

The relationship between the AO-SWJ-NWISM proposed here provides an unraveling pathway. AO is a crucial driver of circulation anomalies that start in early spring and modulate the SWJ during the summer season. During the positive phase of the AO, the dipole structure of the Eurasian trough and Central Asia pressure ridge gets reinforced and there emerges a second trough and ridge meridional dipole over eastern China. As a result, an intensified upper-level jet is found at the southern edge of the Eurasian region. This northward strengthened jet is favorable for summer rainfall over the NWISM region. Note that this wave train does not resemble the typical CGT or Silk Road Pattern (SRP; Lu et al. 2002), which is a zonally oriented teleconnection pattern in boreal summer over the Eurasian continent, with alternative pressure ridges and trough. Instead, it shows a meridional dipole structure in pressure ridges and troughs along the climatological position of the westerly jet.

5. Summary and conclusion

The present study investigates the effect of SWJ on summer rainfall over the NWISM region and the possible mechanisms underlying this relationship. Our results show that the SWJ considerably influences the interannual variations of the NWISM. During positive SWJI years, an anticyclonic anomaly is observed over Central Asia. This upper-level anticyclonic anomaly results in enhanced convection over Pakistan and India as suggested by Ding and Wang (2005). On the contrary, during negative SWJI years, a cyclonic anomaly is observed over Central Asia, which reduces the meridional thermal gradients due to the intrusion of cold westerly winds and suppresses rainfall. Watanabe and Yamazaki (2012) showed that upper-level anticyclone generates anomalous descent over the Thar desert region and promotes the development of a heat low over this region due to strong subsidence and surface sensible heating. The upper-level anticyclonic anomaly is also considered a component of positive CGT pattern and favorable for strong monsoon over the

NWISM region (Ding and Wang 2005). The upper tropospheric heating associated with anticyclonic anomaly over the Afghanistan-western plateau is considered to be a prerequisite for the establishment of the summer monsoon over the NWISM region (Murakami and Ding 1982). Thus, our results are in agreement with Watanabe and Yamazaki (2012), suggesting that the upper-level anticyclonic anomaly caused by the northward shift of the jet influences the NWISM rainfall by enhancing the tropospheric temperature. However here we studied the interannual variability instead of intraseasonal variability. We have also discussed the coupling between the upper and lower troposphere and how the low-level circulation modifies and enhances rainfall over the NWISM region.

By performing detailed diagnostics using observed and reanalysis data, the present study provides a mechanistic understanding of mid-latitude interactions with the NWISM rainfall on the interannual time scale. Our results indicate that the variability of SWJ is significantly linked with the AO. This variability is part of a larger-scale pattern. With a focus on the monsoon and subtropical jet interaction, the study does not rule out the influence of other factors like MJO, ENSO, and IOD on the NWISM rainfall variability. It is well known that MJO, IOD, and ENSO are deeply associated with the interannual fluctuation of the NWISM rainfall. However, it is equally important to recognize the mid-latitude influence on the variability of NWISM rainfall. The study highlights a new pathway through which different processes in mid-latitude can interact with the NWISM rainfall and affect its interannual variability.

Our study has also highlighted the importance of the AO in modulating fluctuations of NWISM rainfall by regulating the SWJ. So far, the NWISM system has been known to be influenced by North Atlantic Oscillations, ENSO and IOD. Here, we have explained the physical mechanism between the AO and the NWISM rainfall. This is another climate mode, which can be used to forecast the interannual variations of NWISM rainfall. This study offers a perspective on understanding the interannual variability of NWISM rainfall in the context of variations in upper-level jet and the remote influence of the AO. Positive phases of the AO induce a dipole meridional wave train structure along the climatological upper-level jet. This causes the intensification of SWJ to the north of its climatological position, which in turn affects the interannual variability of NWISM.

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Data availability statement. The APHRODITE precipitation products used in this study are available at <https://www.chikyu.ac.jp/precip/english/products.html>. The NCEP reanalysis can be acquired from <https://psl.noaa.gov/data/gridded/data.ncep.reanalysis.html>. The AO index can be downloaded from https://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/monthly.ao.index.b50.current.ascii.

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Figure Captions

Fig. 1 Mean summer precipitation (shading; units: m d^{-1}) and 200-hPa zonal wind (contour; units: m s^{-1}) at intervals of 2 m s^{-1} over the South Asia summer monsoon region. The red box shows the domain used for the EOF analysis in this study while the blue box shows the NWISM region used to extract summer rainfall time-series shown in Fig. 3. The green thick line shows the subtropical westerly jet axes during the summer season defined as the maximum of the zonal winds

Fig. 2 The first (a) and second (b) leading mode of EOF analysis of the summer zonal wind at 200 hPa over $25\text{--}55^\circ\text{N}$ and $40\text{--}100^\circ\text{E}$. The corresponding PC time series are shown in (c) and (d), respectively. The green line in (a) and (b) marks the axes of westerly jet during the summer season

Fig. 3 (a) Spatial distribution of correlation coefficients between SWJI and summer rainfall over the NWISM and surrounding regions. (b) Time series of the standardized SWJI (left y-axis) and the summer rainfall (right y-axis) averaged over NWISM (blue rectangle shown in Fig. 1) during 1951–2015 based on APHRODITE data. Red stars in (b) show positive SWJI years while blue circles show negative SWJI years

Fig. 4 Composite distribution of summer rainfall anomalies (shading; units: mm/d) and 200 hPa zonal and meridional wind (vectors; units: m/s) during (a) strong positive and (b) strong negative SWJI years. The 700 hPa anomalies for strong positive and negative SWJI years are shown in (c) and (d), respectively. The green line in (a) and (b) marks the mean axes of the westerly jet during positive and negative SWJI years, respectively

Fig. 5 Composite distribution of 200 hPa mean wave activity fluxes (vectors; units: m^2/s^2) and geopotential height anomalies (shading; units: m) during (a) strong positive and (b) strong negative SWJI years

Fig. 6 Latitude-pressure cross section of anomalous meridional and vertical wind (vectors) and anomalous temperature (shading; units: $^\circ\text{C}$) averaged over $60\text{--}70^\circ\text{E}$ during (a) strong positive SWJI and (b) strong negative SWJI years. The unit of meridional wind speed is m s^{-1} . The unit of vertical wind is Pa s^{-1} , and the vertical wind component is multiplied by 100. The topography is shaded by black in (a) and (b)

Fig. 7 Composite moisture flux divergence (shading; units: $10^{-6} \text{ kg m}^{-2} \text{ s}^{-1}$) and transport (vectors; units: ms^{-1}) integrated from 1000–300 hPa during (a) strong positive SWJI and (b) strong negative SWJI years

Fig. 8 The time series of standardized SWJI and the Arctic Oscillation index during (a) MAM and (b) June. Both correlation coefficients are significant at 95% confidence level

Fig. 9 Regression of summer geopotential height (shading; units: m) and horizontal winds (vectors; units: m s^{-1}) at 500 hPa against (a) AO-MAM and (b) AO-June index.

Fig. 10 Regression of summer air temperature averaged between 600 and 200 hPa (contours; units: $^\circ\text{C}$), and its meridional gradients (shading; units: $10^{-7} \text{ }^\circ\text{C m}^{-1}$) against (a) AO-MAM and (b) AO-June index. Stippling indicates meridional gradients regression coefficients are statistically significant at 95% confidence level

Fig. 11 Regression of 200-hPa zonal wind (shaded; units: m s^{-1}) and Rossby wave activity flux (vectors; units: m^2/s^2) against (a) AO-MAM and (b) AO-June index.

Stippling indicates meridional gradients regression coefficients are statistically significant at 95% confidence level. The vectors of Rossby wave are plotted where the scales are $> 0.4 \text{ m}^2 \text{ s}^{-2}$

Figures

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