Spatial and temporal variability of seasonal precipitation in Poyang Lake basin and possible links with climate indices
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ABSTRACT
Based on the precipitation data of 21 meteorological stations in Poyang Lake basin, the temporal and spatial variability of seasonal precipitation was analyzed by wavelet analysis method. This study adopted the cross wavelet transform to analyze the correlation between the seasonal precipitation and climate indices in time and frequency scales, discussed the possible links between its precipitation variations and climate indices, and preliminarily analyzed its mechanism and regular pattern of variation. The results showed that the oscillations in 2–4 years’ and 4–8 years’ bands were the main variation periods of seasonal precipitation in Poyang Lake basin. In the 2–4 years’ band, the years of rainfall peaks appearing in Poyang Lake were basically consistent with the years when El Niño appeared, and the precipitation oscillations in summer appeared more dramatic in space. According to analysis on the cross wavelet power spectra between different seasonal rainfalls and climate indices, certain correlations between climate factors and seasonal precipitation had existed in specific time periods. Large-scale climate oscillations like the El Niño/Southern Oscillation, North Atlantic Oscillation, Indian Ocean Dipole, and Pacific Decadal Oscillation caused the variability of large-scale circulations through their respective independent or inter-coupled climate systems, and affected the precipitation distribution in Poyang Lake basin by changing local climate conditions like the East Asian Monsoon.

INTRODUCTION
During the past decades, global climate change has had a significant impact on the hydrologic cycle (Stocker et al. 2013), causing large-scale fluctuations of water resources systems (Arnell 1999; Stocker et al. 2013). It is virtually certain that, in the long term, global precipitation will increase with increased global mean surface temperature (Stocker et al. 2013). Precipitation is a major driver for the discharge trends and for the large inter-annual-to-decadal variations. A better understanding of the temporal and regional connection between atmospheric circulation and variability of precipitation can provide valuable information for water resources managers, particularly with regard to long-range precipitation forecasting, water resources development, and utilization. Many researchers have detected possible connections between large-scale circulation factors and the climate of China. Precipitation and streamflow in Asia, especially in the Yangtze River basin, has been mainly influenced by the East Asian summer monsoon and El Niño/Southern Oscillation (ENSO) (Blender et al. 2011; Wei et al. 2014). Gong & Wang (1999) studied the teleconnection between ENSO and precipitation in China with statistical analysis, and suggested that the decreasing precipitation in China usually matched
El Niño events and there existed a significant relationship between winter and autumn rainfall and the ENSO in eastern China. Zhang et al. (2007) detected the variability and possible teleconnections between ENSO and annual maximum streamflow from the lower, the middle, and the upper Yangtze River basin, and results indicated that in the lower Yangtze basin, the in-phase relations occurred between annual maximum streamflow and ENSO. Zhu et al. (2007) indicated that Northern China was relatively dry during the developing phase of ENSO, while the Yangtze River valley was relatively wet during the decaying phase of ENSO. Zhang et al. (2014b) indicated that the annual maximum 1-day precipitation amount \((R \times 1\) day) and maximum 5-day precipitation amount \((R \times 5\) day) were influenced mainly by the ENSO events a year earlier in the Poyang Lake basin, and the relations between annual \(R \times 1\) day and \(R \times 5\) day and ENSO were statistically positive.

However, the relationship between precipitation and climatic modes are non-linear and non-stationary, so it is difficult to assess their changes by using statistical methods. Most previous studies are either based on direct correlation to identify the statistical relationship between the climatic variable and the teleconnection pattern indices or more recently on a non-parametric method of spectral analysis. It is now well established that the inter-annual variability of climatic oscillations such as ENSO are non-stationary processes, since their variance changes in frequency and intensity through time (Torrence & Compo 1998; Higuchi et al. 1999), and these non-stationary features are also manifested in precipitation time series (Nolin & Hall-McKim 2006; Pezzi & Kayano 2009). The wavelet transform is a useful time–frequency analysis method that is widely used to analyze non-stationary hydrological time series at different frequencies (Daubechies 1990; Kumar et al. 1997; Torrence & Compo 1998). Wavelet transform could play an important role in analyzing precipitation data, since its local analysis and multiresolution decomposition make the analysis process more efficient and accurate (Kuo & Sheng 2010).

Poyang Lake (28°22′–29°45′N, 115°47′–116°45′E) is located in the middle reaches of the Yangtze River and is the largest freshwater lake in China. Variations in rainfall and other forms of precipitation are one of the most critical factors determining the water resources, floods and droughts, and ecosystem of the Poyang Lake basin. In recent decades, due to climate change, Poyang Lake basin rainfall characteristics have also been changed. Guo et al. (2006), Wang et al. (2009), and Huo et al. (2011) used the Mann–Kendall method to analyze the trend of precipitation of Poyang Lake. Zhang et al. (2014a, 2014b) analyzed the spatial patterns, changing properties and causes of precipitation extremes in Poyang Lake, and investigated the impacts of climate indices on extreme precipitation processes. Li et al. (2012b) compared the difference of Tropical Rainfall Measuring Mission (TRMM) rainfall with rain gauge data at different time scales, and evaluated the usefulness of the TRMM rainfall for hydrological process simulation and water balance analysis in the Xinjiang catchment, one of the five river catchments of Poyang Lake. Yuan et al. (2013) analyzed the spatial distribution of Poyang Lake rainfall based on geostatistical and geographic information system technology. Li et al. (2022a) and Yuan et al. (2014) used the wavelet method to analyze the temporal variability of rainfall data of meteorological stations in Poyang Lake basin. Zhang et al. (2014c) studied variation characteristics of annual precipitation in Poyang Lake basin with the HBV hydrological model, and explored hydrological processes’ driving factors based on BCC-CSM global climate system model forecast data (2014–2050).

However, the spatial and temporal variability of seasonal precipitation considering climatic fluctuations has not been investigated in the Poyang Lake basin. By only analyzing the changes in observed precipitation data, it is difficult to sufficiently relate hydrological changes to a climatic signal, because of the non-linear behavior of both hydrological records and climatic patterns. Therefore, the objectives of this study aim to describe the spatial and temporal variability in Poyang Lake basin precipitation based on non-linear and non-stationary methods, and then to identify possible links to the dominant climatic patterns on the seasonal variability during the past 50 years. A description of the study area and data is first presented, followed by a brief description of the wavelet method. The results from the wavelet analysis are presented next, and finally some conclusions are drawn.

**STUDY AREA AND DATASETS**

**Study area and data**

Poyang Lake basin (28°22′–29°45′N, 115°47′–116°45′E) lies on the northern border of Jiangxi Province in China,
which covers a drainage area of $1.62 \times 10^5 \text{km}^2$, and accounts for 9% of the total area of the Yangtze River Basin. The lake receives water flows mainly from five rivers: Ganjiang, Fuhe, Xinjiang, Raohe, and Xiushui, and discharges into the Yangtze River from a narrow outlet in the north (Figure 1).

The precipitation used in this study was extracted from the National Meteorological Observatory stations in the Poyang Lake basin of the China Meteorological Administration. A total of 21 stations were selected based on the continuous data series from 1960 to 2010. The mean annual and seasonal precipitation time series were constructed according to monthly data. Each year was divided into four seasons, namely, spring (March to May), summer (June to August), autumn (September to November), and winter (December to February the following year). For the time series of each season, the mean precipitation is the average of 3 months’ value. The information concerning meteorological stations used in the study is listed in Table 1, and the spatial distribution is shown in Figure 1.

Climatic patterns

Climatic oscillations have a close relationship with the climate of China. For example, the impact of ENSO on precipitation in China is one of the most important factors in the consideration of short-term climate prediction. Xiao et al. (2014) found that ENSO was the leading driver of

![Figure 1](http://iwaponline.com/hr/article-pdf/47/S1/51/367496/nh047s10051.pdf)
seasonal precipitation variability over the Yangtze River basin, and the spring precipitation was influenced by the Pacific Decadal Oscillation (PDO) and ENSO; the summer and autumn precipitation was influenced by the ENSO and Indian Ocean Dipole (IOD); and the winter precipitation was influenced by the ENSO, IOD, and North Atlantic Oscillation (NAO). Zhang et al. (2014b) pointed out that the Poyang Lake basin was climatically dominated by the Southeast Asian Monsoon. Meanwhile, the East Asian Monsoon had also been significantly influenced by ENSO, NAO, IOD, and PDO (Yuan et al. 2008; Linderholm et al. 2011; Chakravorty et al. 2013; Chen et al. 2013a). Therefore, ENSO, NAO, IOD, and PDO climatic patterns were selected here in order to investigate the specific link between each climatic indicator and seasonal precipitation across the Poyang Lake basin, China.

ENSO stands for El Niño/Southern Oscillation, where Southern Oscillation is the term for atmospheric pressure changes between the east and west tropical Pacific that accompany both El Niño and La Niña episodes in the ocean. There are many indicators that characterize the ENSO. Index ONI (Oceanic Niño Index) provided by the National Oceanic and Atmospheric Administration (NOAA, Oceanic Niño Index: http://www.cpc.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml), is adopted in this study. ONI is a measure of the deviation from the normal sea surface temperature (SST) in the central-eastern Pacific Ocean of Niño 3.4 region. The region extends over a strip of 5° N and 5° S latitude and 120° W–170° W longitude.

NAO refers to a north–south alternation in atmospheric mass between the subtropical Atlantic and the Arctic, and thus involves out-of-phase behavior between the climatological low-pressure center near Iceland and the high-pressure center near the Azores (Stenseth et al. 2005). NAO is the most prominent and recurrent pattern of atmospheric

| Station ID | Station name | Region | Longitude (°E) | Latitude (°N) | Altitude (m) |
|------------|--------------|--------|----------------|---------------|--------------|
| 57883      | Ninggang     | Jiangxi| 113.97         | 26.72         | 263.1        |
| 57894      | Jinggangshan | Jiangxi| 114.17         | 26.58         | 843.0        |
| 57996      | Nanxiong     | Guangdong | 114.32      | 25.13         | 133.8        |
| 57793      | Yichun       | Jiangxi| 114.38         | 27.80         | 131.3        |
| 57896      | Suichuan     | Jiangxi| 114.50         | 26.33         | 126.1        |
| 57598      | Xiushui      | Jiangxi| 114.58         | 29.03         | 146.8        |
| 57993      | Ganzhou      | Jiangxi| 114.95         | 25.85         | 123.8        |
| 57799      | Jian         | Jiangxi| 114.97         | 27.12         | 76.4         |
| 58608      | Zhangshu     | Jiangxi| 115.55         | 28.07         | 30.4         |
| 59102      | Xunmiao      | Jiangxi| 115.65         | 24.95         | 303.9        |
| 58606      | Nanchang     | Jiangxi| 115.92         | 28.60         | 46.7         |
| 58506      | Lushan       | Jiangxi| 115.98         | 29.58         | 1,164.5      |
| 58502      | Jiujiang     | Jiangxi| 116.00         | 29.73         | 36.1         |
| 58813      | Guangchang   | Jiangxi| 116.33         | 26.85         | 143.8        |
| 58911      | Changting    | Fujian | 116.37         | 25.85         | 310.0        |
| 58715      | Nancheng     | Jiangxi| 116.65         | 27.58         | 80.8         |
| 58519      | Poyang       | Jiangxi| 116.68         | 29.00         | 40.1         |
| 58527      | Jingdezhen   | Jiangxi| 117.20         | 29.30         | 61.5         |
| 58626      | Guixi        | Jiangxi| 117.22         | 28.30         | 51.2         |
| 58726      | Qixianshan   | Fujian | 117.83         | 27.95         | 1,401.9      |
| 58634      | Yushan       | Jiangxi| 118.25         | 28.68         | 116.3        |
variability over the middle and high latitudes of the northern hemisphere, especially during the cold season months (November–April). The NAO index used here is taken from the NOAA (http://www.cpc.ncep.noaa.gov/products/precip/CWlink/pna/norm.nao.monthly.b5001.current.ascii.table).

IOD is a coupled ocean–atmosphere phenomenon, being defined as the SST anomaly between the western equatorial Indian Ocean (50°E–70°E, 10°S–10°N) and the southeastern equatorial Indian Ocean (90°E–110°E, 10°S–0°N), and is also referred to as Dipole Mode Index (DMI) (Saji et al. 1999; Ashok et al. 2001; Saji & Yamagata 2003). The IOD is taken from the Japan Agency for Marine Earth Science and Technology (JAMSTEC; http://www.jamstec.go.jp/frsgc/research/d1/iod/iod/dipole_mode_index.html).

The PDO is defined as the leading empirical orthogonal function of monthly anomalies (deviations from the climatological annual cycle) of sea surface temperature in the Pacific poleward of 20°N (Davis 1976; Mantua et al. 1997). PDO is most frequently referred to as a long-lived El Niño-like pattern of the Pacific climate variability (Zhang et al. 1997). The PDO index is taken from the Earth System Research Laboratory (ESRL) of NOAA (http://www.esrl.noaa.gov/psd/data/correlation/pdo.data).

**METHODOLOGY**

**Continuous wavelet transform**

A wavelet transform is a useful time–frequency analysis method that is widely used to analyze non-stationary hydrological time series at different frequencies (Daubechies 1990; Kumar et al. 1997; Torrence & Compo 1998). A wavelet transform can be classified into two categories, namely, continuous and discrete. A continuous wavelet transform $W_n$ is the convolution of a vector $x$ (with time dimension $n$) with a wavelet function $\Psi$:  

$$W_n(s) = \sum_{n}^{N-1} x_n \psi \left( \frac{n'-n}{s} \delta t \right)$$

(1)

where $s$ is the scale or dilation, $n'$-$n$ shows the number of points from the time series origin, $\delta t$ is the time interval, $N$ is the number of points, and the overbar indicates the complex conjugate.

We chose the complex Morlet wavelet function, which is commonly used for the wavelet base function:

$$\psi_0(\eta) = \eta^{-1/4} e^{i\omega_0 \eta} e^{-\eta^2/2}$$

(2)

where $\omega_0$ is a non-dimensional frequency and $\eta$ is a non-dimensional time.

A cone of influence (COI) is used to avoid wavelet edge effects. More details of the wavelet transform used in hydrology are found in the references, such as those by Torrence & Compo (1998) and Labat et al. (2000).

To examine fluctuation in power over a range of scales (a band), one can define the scale-averaged wavelet power as the weighted sum of the wavelet power spectrum over different scales (Torrence & Compo 1998). The wavelet power spectrum is shown in Figure 2 for the average seasonal rainfall in the Poyang Lake basin using the Morlet wavelet, which represented the average variance over a range of scales. The thick contours depicted the 95% confidence level of local power relative to red noise. The U-shaped line is the COI beyond which the result might be distorted by edge effects. Figure 2 reveals that the seasonal rainfall had more kinds of quasi-periodic oscillations, which mainly focused on 2–4 and 4–8 years’ bands in summer, autumn, and winter. This suggested the choice of the scale-averaged wavelet power to further examine fluctuations in power over specific ranges of wavelet periods (bands).

Scale-averaged wavelet power is defined as the weighted sum of the wavelet power spectrum over scales $s_1$ to $s_2$:

$$W_n^z = \frac{\delta_i \delta_j}{C_n} \sum_{s_{ij}} \frac{|W_n(s)|^2}{s_{ij}}$$

(3)

where $\delta_i$ is a factor that dictates the scale resolution (chosen as 0.1), and $C_n$ is a reconstruction factor specific to each wavelet form; $C_n = 0.776$ for the Morlet. It is possible to explore activity around a period of 8 years. However, for the span of the available precipitation data, it is not reasonable to consider wavelet periods much beyond 8 years. Two bands of wavelet...
periods are examined in detail: 2–4 and 4–8 years. The power spectrum produced for a given time series is the product of the natural process involved and noise. The contour lines identify peaks of greater than 95% confidence for a red-noise process with a lag-1 coefficient of 0.23 following the Monte Carlo analysis of Torrence & Compo (1998) based on the univariate lag-1 autoregressive process.

**Cross wavelet transform**

Cross wavelet transform combines wavelet transform and cross spectrum analyses. Thus, it can examine two time series correlations that are expected to be linked in some way (Grinsted et al. 2004).

Two time series, \(X = \{x_1, x_2, \ldots, x_n\}\) and \(Y = \{y_1, y_2, \ldots, y_n\}\), are given. We can define the cross wavelet spectrum as \(W_{XY}(s) = W_X^Y(s)W_Y^X(s)\) according to the continuous wavelet transform \(W_X^Y(s)\) and \(W_Y^X(s)\), where \(W_Y^X(s)\) is the complex conjugate of \(W_X^Y(s)\). We further define the cross wavelet power as \(|W_{XY}(s)|\). Given two stationary random processes, the standardization of the cross wavelet transform can be defined as a wavelet cross correlation coefficient such as:

\[
 r(X, Y) = \frac{\sum_{i=1}^{n} (W_X^Y(s) - \bar{W}_X^Y(s))(W_Y^X(s) - \bar{W}_Y^X(s))}{\sqrt{\sum_{i=1}^{n} (W_X^Y(s) - \bar{W}_X^Y(s))^2} \sqrt{\sum_{i=1}^{n} (W_Y^X(s) - \bar{W}_Y^X(s))^2}}
\]

(4)

Thus, the wavelet squared coherency is defined as the absolute value squared of the smoothed cross wavelet spectrum normalized by the smoothed wavelet power spectra:

\[
 R_n^2(s) = \frac{|S(s^{-1}W_{XY}^n(s))|^2}{S(s^{-1}|W_{XY}^n(s)|^2) \cdot S(s^{-1}|W_{X}^n(s)|^2)}
\]

(5)

where \(S\) is a smoothing operator of the Morlet wavelet, which is given as:

\[
 S_{time}(\omega)_{|s} = (W_n(s) \ast c_2 \mathcal{F}^s)|_s
\]

\[
 S_{time}(\omega)_{|s} = (W_n(s) \ast c_2 \prod (0.6s))_{|s}
\]

(6)

Figure 2 | Morlet wavelet power spectrum of average seasonal rainfall in Poyang Lake basin: (a) spring, (b) summer, (c) autumn, and (d) winter.
where $c_1$ and $c_2$ are normalization constants and $\Pi$ is a rectangle function.

**Significance test for correlation analysis**

A red-noise background spectrum was chosen for the significance test to examine the suitability of the cross wavelet transform. Many hydrological time series can be modeled as red noise. Red noise is a stationary stochastic process. The joint probability distribution of red noise does not change when it is shifted in time or space. The parameters of red noise, such as mean and variance, also do not change over time or position. A simple example of red noise is the AR(1) model

$$x_n = \alpha x_{n-1} + z_n \quad (7)$$

where $\alpha$ is the lag-1 autocorrelation, $x_0 = 0$, and $z_n$ follows a normal distribution with the mean equal to zero. The discrete Fourier power spectrum of (6) can be normalized as:

$$P_k = \frac{1 - \alpha^2}{1 + \alpha^2 - 2\alpha \cos(2\pi k/N)} \quad (k = 0, 1, 2, \ldots, N/2) \quad (8)$$

where $k$ is the frequency index.

Therefore, (7) can be modeled as a red-noise series by choosing the lag-1 autocorrelation especially if $\alpha = 0$ is the white-noise series.

We can deduce (7) to be composed of two time series. If two time series have theoretical Fourier spectra, then the distribution of the cross wavelet power with Fourier red noise spectra $P^X_k$ and $P^Y_k$ is

$$D\left(\frac{|W_n^X(s)W_n^Y(s)|}{\sigma_X\sigma_Y} < p\right) = \frac{Z_v(p)}{v} \sqrt{\frac{P^X_k P^Y_k}{p^X_k p^Y_k}} \quad (9)$$

where $\sigma_X$ and $\sigma_Y$ are the respective standard deviations, $v$ is the degree of freedom, and $Z_v(p)$ is the confidence level associated with probability $p$. At a confidence level of 95%, $v = 1$ and $Z_1(95%) = 2.182$ for the real wavelet. For the complex wavelet, $v = 2$ and $Z_2(95%) = 3.999$.

**RESULTS**

**Power Hovmöller of Poyang Lake basin seasonal rainfall**

The scale-average wavelet power spectra represented the average variance over a range of scales (or a certain band). It provides an efficient way to examine the fluctuations in power over a desired band. By averaging the wavelet power spectra at multiple locations, one could assess the spatial and temporal variability of a field of data (Torrence & Compo 1998; Anctil & Coulibaly 2004; Coulibaly & Burn 2004, 2005; Coulibaly 2006). This representation provides an efficient approach to examining irregular variations at different longitudes. Figure 3 shows a power Hovmöller (time-longitude diagram) of the wavelet variation for seasonal rainfall of Poyang Lake basin in the 2–4 years’ band at each longitude. The original time series at each longitude was the average rainfall between 24°N and 30°N. At each longitude, the wavelet power spectrum was computed using the Morlet wavelet, and the scale-averaged wavelet power over the 2–4 years’ band was calculated from formula (3). The time series were then combined into a two-dimensional contour plot. The 95% confidence level was computed using the lag-1 auto-correlation at each longitude.

The average wavelet power of the spring rainfall time series is shown in Figure 3(a). The left diagram in Figure 3(a) is the Hovmöller diagram of the spring precipitation in each station at different longitudes, where the regions encircled by thick lines indicate that they had passed the significance test of 95% confidence level. In the bands with significant oscillations, both in different years and at different longitudes, especially in 1970–1975, the significant regions run through the whole river basin. It indicates that in the past 50 years, the spring rainfall of the whole river basin in 1970–1975 was at the peak. The right diagram in Figure 3(a) shows the mean power spectra in different temporal ranges corresponding to the Hovmöller diagram of spring precipitation. The wavelet power spectrum of spring rainfall in 1970–1975 reached the maximum value, and was consistent with the Hovmöller diagram analysis. Moreover, the oscillations in different degrees existed in the years 1980–1985, 1990–1995, and around 2000. However, in the 21st century, the precipitation appears to be in a gradual reduction trend.
The bottom diagram in Figure 3(a) shows the average power spectra in different longitude ranges corresponding to the Hovmöller diagram of spring precipitation. The precipitation variation in different spaces is clearly shown in the diagram. The region with a high power spectrum indicates its precipitation oscillation was very dramatic.

Similar regular patterns could also be found for other seasons. Figure 3(b) is the spatial–temporal distribution diagram of summer precipitation in Poyang Lake. In the temporal scale, there were big fluctuations in summer precipitation from the mid-1990s to the early 21st century and certain fluctuations in the mid-1960s. In the spatial scale, obvious precipitation oscillations appeared in the regions of 114.5°E and 115.5–116°E. The regular pattern of summer precipitation represented in the temporal scale is relatively uniform, but relatively concentrated in the spatial scale. Figure 3(c) is the autumn precipitation distribution diagram. It can be seen from the Hovmöller diagram and power spectrum diagram that less autumn precipitation in Poyang Lake was observed in the early 1960s, late 1970s to mid 1980s, and around 1995. In the years 1995–2005 and 1970–1975, there was much autumn precipitation in Poyang Lake. In the winter diagram of Figure 3(d), the rainfall is mainly concentrated in two big time nodes. The biggest occurred in the years 1995–2000 or so, and the second biggest in the 1960s. The statistics of the years with significant oscillation under the 2–4 years’ band of each station are listed in Table 2. The significant years in Table 2 correspond one-to-one with the significant regions in Figure 3, and thus are more favorable to identify the years with a large precipitation process and the years with less precipitation.
Table 2 | Seasonal rainfall in the 2–4 years' band period of locally significant wavelet power spectrum at 95% confidence level

| Station ID | Station name | Location (E, N) | Period of locally significant wavelet power |
|------------|--------------|-----------------|---------------------------------------------|
| 57883      | Ninggang     | 113.97, 26.72   | Spring 1963–1970, 1977, 1980–1985 1998–2004 |
| 57894      | Jinggangshan | 114.17, 26.58   | Summer 1969–1977, 1980–1985 1998–2005 |
| 57996      | Nanxiong     | 114.32, 25.13   | Autumn 1970–1982, 1991–1994 1961–1971 |
| 57793      | Yichun       | 114.38, 27.8    | Winter 1969–1976, 1991–1992 1965–1971 |
| 57896      | Suichuan     | 114.5, 26.33    | Spring 1964, 1969–1976, 1982–1984, 1991–1992 1966–1969, 1994–2003 |
| 57598      | Xiushui      | 114.58, 29.03   | Summer 1965–1974, 1991–1992 1967–1970, 1983, 1993–1999 |
| 57993      | Ganzhou      | 114.95, 25.85   | Autumn 1969–1976, 1991–1992 1965–1971 |
| 57799      | Jian         | 114.97, 27.12   | Winter 1969–1977, 1980–1985, 1992–1993 1961–1963, 1980–1982, 1998–2004 |
| 58608      | Zhangshu     | 115.55, 28.07   | Spring 1962–1964, 1969–1978, 1980–1985 1976–1980, 2000 |
| 59102      | Xunniao      | 115.65, 24.95   | Autumn 1963–1966, 1972–1976, 1990–1994 1961–1971, 1997–2000 |
| 58606      | Nanchang     | 115.92, 28.6    | Winter 1963–1967, 1987–1991 1977–1980, 1997–2009 1963–1968, 1979–1983, 1992–1994, 2003–2007 |
| 58306      | Lushan       | 115.98, 29.58   | Summer 1966–1975, 1996–2003 1968–1981 |
| 58502      | Jiujiang     | 116, 29.73      | Autumn 1968–1975, 1996–2003 1969–2000 |
| 58813      | Guangchang   | 116.33, 26.85   | Winter 1966–1975, 1996–2003 1966–1969, 1995–1999–2006 |
| 58911      | Changting    | 116.37, 25.85   | Spring 1968–1977 1964–1969, 1988–1991, 2002–2004 |
| 58715      | Nancheng     | 116.65, 27.58   | Summer 1970–1977, 1981–1986 1975–1982 1986–1993, 1995–2003 1962–1967, 1993–2000, 2005–2009 |
| 58519      | Poyang       | 116.68, 29      | Autumn 1970–1975, 1991–2002 1992–1995 |
| 58527      | Jingdezhen   | 117.2, 29.3     | Spring 1969–1975, 1991–2003 1969, 1979–1982, 1992–1998 |
| 58626      | Guixi        | 117.22, 28.3    | Winter 1969–1977 1995–1998 1994–2001 1962–1970, 1993–2001, 2003–2004 |
| 58726      | Qixianshan   | 117.83, 27.95   | Summer 1969–1977, 1980–1985 1990–1999 1987–1991 |
| 58634      | Yushan       | 118.25, 28.68   | Autumn 1965–1984, 1994–1997, 2000 1989–1999 1971–1975, 1987–1989, 1994–2001 |

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According to the results of ENSO studies (Amarasekera et al. 1997; Mason & Goddard 2001; Alexander 2002; Ward et al. 2010), ENSO would become the main mode dominating the tropical Pacific region in the 21st century and have a closer correlation with precipitation variation in terms of regional scale due to the increase in water vapor supply. In the 2–4 years’ band, the years of rainfall peaks appearing in Poyang Lake were basically consistent with the years when El Niño appeared, such as in 1963–1964, 1965–1966, 1972–1973, 1982–1983, and 1997–1998, when a wide range of El Niño presented. The 2–4 years’ band was also similar to the El Niño appearance period.

In order to analyze the spatial distribution of seasonal rainfall under the 4–8 years’ band conditions, the time-longitude diagrams of seasonal precipitation in Poyang Lake basin were drawn as shown in Figure 4. Compared with the time-longitude diagram of the 2–4 years’ band (Figure 3), except for the slightly dramatic summer precipitation oscillation, the remainder of the seasonal precipitation oscillations appeared much gentler within the whole spatial and temporal ranges. Figure 4(a) shows the spatial–temporal distribution diagram of spring precipitation. In the temporal scale, the precipitation extremes in certain ranges occurred in the 1960s and mid-1980s to mid-1990s; and in the spatial scale, certain precipitation oscillation extremes existed at 114.5°E, 115°E, 115.5°E, 116.5°E, and 117.5°–118°E. The summer precipitation oscillation in Figure 4(b) was obviously greater, which also verified the characteristic of Poyang Lake basin being significantly influenced by the summer monsoon. Especially in the period 1990–2005, the precipitation oscillations ran through the whole river basin. In addition, the rainfall extremes of
certain degrees appeared in the regions near 115° E and 117.5° E in the 1970s or so. Figure 4(c) is the time–longitude diagram of autumn precipitation. In the regions at 114.5–115° E, 115.5–116° E, 116–116.5° E, and 117.5° E, the inter-annual oscillation continuing through 1965–1985 was very significant. In the band of Poyang Lake over 4–8 years, the oscillation of winter precipitation (Figure 4(d)) only existed in local regions at 115.5° E (1994–2003) and 116° E (1984–1999). Likewise, the statistics of the years with significant oscillation under the 4–8 years’ band of each station is listed in Table 3.

Relationship between seasonal rainfall and climate indices

The cross wavelet power spectra reflect the dependency relationships of the correlation to the time and frequency. Figures 5–8 illustrate the cross wavelet power spectra between the spring, summer, autumn, and winter precipitation amounts in Poyang Lake basin and the corresponding ONI, NAO, IOD, and PDO of each season. The regions encircled by thick lines indicate that they had passed the significance test of 95% confidence level. The relative phase relationship is shown as arrows (with in-phase pointing right, anti-phase pointing left, and climatic factor leading precipitation by 90° pointing straight down). To avoid the boundary effect and false wavelet high-frequency information, the region within the wavelet influence cone (the arc in the diagram) is the effective spectral value.

It can be seen from Figure 5 that the spring precipitation amount and climate factors in spring have certain positive correlations at different temporal scales. It can be seen from Figure 5(a)–5(d) that the main climate factors affecting the spring precipitation include ENSO, NAO, IOD, and PDO,

### Table 3 | Seasonal rainfall in the 4–8 years’ band period of locally significant wavelet power spectrum at 95% confidence level

| Station ID | Station name | Location (E, N) | Period of locally significant wavelet power |
|------------|--------------|-----------------|--------------------------------------------|
|            |              |                 | Spring | Summer | Autumn | Winter |
| 57883      | Ninggang    | 113.97, 26.72   | 1996–1999 | 2000–2006 | 1980–1986 |
| 57894      | Jinggangshan| 114.17, 26.58   | 1999–2007 | 1980–1985 |
| 57996      | Nancheng    | 114.32, 25.13   | 1984–1994 |
| 57793      | Yichun      | 114.38, 27.8    | 1961–1964, 1985–1996 | 2000–2001 |
| 57896      | Suichuan    | 114.5, 26.33    | 1983–1989 | 1968–1972, 1999–2006 | 1967–1979 |
| 57598      | Xiushui     | 114.58, 29.03   | 2000–2002 | 1984–1999 | 1970–1985 | 1986–1992 |
| 57993      | Ganzhou     | 114.95, 25.85   | 1961–1964, 1985–1996 | 2000–2001 |
| 57799      | Jian        | 114.97, 27.12   | 1965–1974, 2000–2007 | 1978–1986 | 1998 |
| 58608      | Zhangshu    | 115.55, 28.07   | 1985–1990 | 1995–2000 | 1977–1982 | 1994–2003 |
| 59102      | Xunmiao     | 115.65, 24.95   | 1985–1992 | 1997–2007 | 1966–1977, 1998–2004 |
| 58606      | Nanchang    | 115.92, 28.6    | 1992–2002 |
| 58506      | Lushan      | 115.98, 29.58   | 1974–1980 | 2001–2009 | 1986–1999 |
| 58502      | Jiujiang    | 116, 29.73      | 2001–2009 |
| 58813      | Guangchang  | 116.33, 26.85   | 1966–1970, 1998–2003 | 1966–1983 | 1986–1988, 1997–1998 |
| 58911      | Changting   | 116.37, 25.85   | 1989–1995, 2000–2007 | 1994–2006 | 1977–1981, 1997–2004 |
| 58715      | Nancheng    | 116.65, 27.58   | 1960–1966 |
| 58519      | Poyang      | 116.68, 29      | 1969–1973, 1988–1998 | 1961–1963 |
| 58527      | Jingdezhen  | 117.2, 29.3     | 1992–2001 | 1971–1974 |
| 58526      | Guixi       | 117.22, 28.3    | 1969–1970, 1989–2001 | 1970–1984 | 1996–1998 |
| 58726      | Qixianshan  | 117.83, 27.95   | 1960–1967 | 1969–1973, 1995–1996 | 1966–1972 |
| 58634      | Yushan      | 118.25, 28.68   | 1989–1998 | 1976–1978 |
and are mainly concentrated in the 2–4 years’ band. For ENSO (Figure 5(a)), the spring precipitation amount in Poyang Lake basin and the ONI in spring has a positive correlation (phase angle of around 315°) in the 2–4 years’ band during 1966–1975, and a positive correlation (phase angle of around 315°) in the band scale of about 4 years during 1981–1992. NAO (Figure 5(b)) has a negative correlation (180° ± 10°) to the precipitation in 1970–1976 and is mainly concentrated in the 2–3 years’ band in the time domain. IOD (Figure 5(c)) has a negative correlation (225°) both in 1985–1996 and the 3–4 years’ band. PDO (Figure 5(d)) has a positive correlation (0–30°) to the precipitation in the 3–4 years’ band during 1970–1973 and a positive correlation (−30°–0°) in the 3–5 years’ band during 1980–1990.

Figures 6–8 illustrate the cross wavelet power spectrum between the summer, autumn, and winter precipitation
amount and the ONI, NAO, IOD, and PDO of corresponding seasons, respectively. The influence of each summer climate factor (Figure 6) on precipitation is concentrated mainly in the 3–6 years’ band in the time domain and in several periods of the late-1970s to early-1980s and late-1980 to 1990s in the time axis. There were large and significant regions in autumn (Figure 7). The influence of ENSO (Figure 7(a)) on precipitation is concentrated in several time periods, 1968–1972, 1975–1985, and 1987–2000, with the temporal scale mainly in the 2–6 years’ band. NAO (Figure 7(b)) has a significant influence on precipitation in the larger scales such as the 6–8 and 12–16 years’ bands. There were significant regions in the 2–4 years’ band during 1990–2000 and in the 5–6 years’ band during
1969–1984 in the case of IOD (Figure 7(c)). PDO (Figure 7(d)) has a relatively small influence on precipitation and only existed in the individual periods of the early 1970s and late 1990s. The winter (Figure 8) ONI, NAO, IOD, and PDO has certain influences on precipitation in different time periods. ENSO is mainly concentrated in the mid-to-late 1980s and mid-to-late 1990s; NAO is mainly concentrated around 1980 and around 1998; IOD is mainly concentrated in 1994–1999 or so; and PDO is concentrated in the mid-to-late 1990s. According to the above analysis, certain correlations between the climate factors and seasonal precipitation existed in specific time periods.

**DISCUSSION**

Located in the East Asian Monsoon region, the precipitation in China is mainly caused by water vapor convergence (Zhou et al. 1999; Zhou & Yu 2005), while the water vapors’ transfer is closely related to the large-scale atmospheric circulation situation. As the main mode of inter-annual climate variability of the ocean–atmosphere system in the tropical Pacific region, ENSO has a significant influence on China’s climate. It is generally believed that the East Asian winter and summer monsoons might be weak in El Niño years (the ENSO positive phase). In an El Niño event’s development stage, there is a large amount of summer precipitation in the Yangtze-Huai River basin and less precipitation in northern China and the southern Yangtze River regions. In an El Niño event’s decline stage, the summer precipitation in the aforementioned regions basically appears in an opposite trend, with frequent floods in the southern Yangtze River regions. China’s precipitation variation corresponding to a La Niña year’s (the ENSO negative phase) development stage, and the decline stage has a totally opposite distribution trend.

The occurrence of an El Niño event creates the main atmospheric circulation anomalies favorable for much rain in summer in the Yangtze River basin. In the case of an El Niño anomaly of tropical Pacific subsurface sea temperature in the fore-winter, an anticyclonic circulation anomaly would be generated in the East Asian continental coast, South China Sea and ocean surface to the east of Taiwan in the next summer, with a strong abnormal cyclonic circulation over Japan in the north, a cyclonic circulation over the high-altitude Lake Baikal and an anticyclonic circulation over the Kuril Islands in the east. Such a circulation situation is favorable for the cold air in the north and the cold wet air current in the northeast direction to be transferred to the south and to converge with abundant water vapor at the western end of the subtropical high in the Yangtze–Huai River basin, causing much precipitation in summer in the Poyang Lake basin.

The NAO in winter and spring plays an important role in the anomalous distribution of China’s summer precipitation, with prominent contribution to the summer precipitation in the middle and lower reaches of the Yangtze River especially. According to the studies by Wei et al. (2008), the NAO index in May and the precipitation amount in summer in the Yangtze River basin have a significant negative correlation over the 0.05 significance level. Among the numerous influencing factors, like SST and solar activity, the NAO in spring had the most stable correlation with the summer precipitation in the middle and lower reaches of the Yangtze River, which is a negative correlation maintained since the 1950s. Linderholm et al. (2011) found that some major patterns of summer climate over China, including the East Asia summer monsoon, were highly connected with the inter-annual variation of summer NAO, supporting a teleconnection between the North Atlantic region and East Asia previously found in winter. Feng et al. (2011) pointed out that in the NAO’s negative phase year, the westerly jet in summer was close to the south, the western Pacific subtropical high was close to the west and south, and the southwest water vapor transfer from the west Pacific Ocean was stronger, causing much precipitation in summer in the southern Yangtze River regions and southern China regions. On the contrary, in NAO’s positive phase year, there was less precipitation in southern China regions.

The observation data and studies show that the Indian Ocean has SST anomaly dipole phenomena similar to the Pacific ENSO events that have significant influences on the Asian monsoon and the atmospheric circulation and precipitation anomalies in the monsoon region. Chen et al. (2015b) pointed out that the IOD events were the comprehensive results of the coupling function of two independent ocean–atmosphere systems in the tropical
Pacific Ocean–Indian Ocean scale and Indian Ocean scale, respectively, where the former was closely related to the Walker circulation anomaly caused by the ENSO event and the latter was connected to the Mascarene high-pressure anomaly in the southern Indian Ocean. According to the results of the studies by Yuan (2006), in the IOD’s positive phase year, the precipitation amount increased in summer in the southern Yangtze River regions, especially the Poyang Lake basin; and in the IOD’s negative phase year, the precipitation in the Poyang Lake basin was less. It was discovered by analyzing the circulation field that in the summer, the southwestern monsoon and South China Sea monsoon was strengthened, the subtropical high was close to the south and weakened, and the precipitation in southern China increased. The preliminary influence mechanism was obtained by analyzing the vertical circulation anomalies, namely, the latitudinal SST distribution anomalies in the Indian Ocean in the IOD’s positive phase year that caused the Walker circulation weakening and the longitudinal Hadley circulation weakening as a result.

As a decadal variability signal with the most significant latitude in the North Pacific Ocean, PDO is the most important factor regulating the climatic teleconnection relationship in the East Asia regions. In winter of the PDO’s positive phase year, the precipitation significance anomalies in most Yangtze River basin regions including the Poyang Lake basin are less. This is mainly due to the high Mongolian high anomalies in winter and strong East Asian winter monsoon causing most regions in China to be affected by the northwest air current, which is not conducive to the generation of precipitation. In summer of the PDO’s positive phase year, the precipitation anomalies in the regions of the middle and lower reaches of the Yangtze River were great, thus forming the precipitation anomaly distribution pattern similar to the ‘droughts in the north and floods in the south’ (Huang et al. 1999; Zhu & Yang 2003; Yang et al. 2005). PDO is one of the important reasons resulting in China’s summer precipitation appearing in the decadal shift of ‘droughts in the north and floods in the south’ after the late-1970s (Yang et al. 2005). In the PDO’s positive phase year, the West Pacific subtropical high was close to the south, the geo-potential height anomaly at the south side of Lake Baikal to Lake Balkhash in the East Asia continent increased, and the Lake Balkhash upper trough weakened to cause the systematic northwest air currents in the upper level of the troposphere. Moreover, the pressure abnormalities in the Mongolia Plateau and the northwest China regions became higher and caused the East Asian summer monsoon weakening and the large amount of water vapor from the Bay of Bengal, South China Sea, and Western Pacific Ocean to be transferred to the regions of the middle and lower reaches of the Yangtze River, thus resulting in significantly high precipitation in the Poyang Lake basin (Zhu & Yang 2003).

Many studies have indicated that PDO has a modulation effect on ENSO, thus affecting the precipitation in southern China. In the PDO’s negative phase year, the ENSO event in the development stage did not have an obvious influence on China’s summer precipitation generally. In the PDO’s negative phase year, the influence of the ENSO event in the decline stage on China’s summer precipitation was mainly represented by significant precipitation in the Yangtze River basin, where the regions of the middle and lower reaches of the Yangtze River, including the Poyang Lake basin, were typically represented. While in the PDO’s positive phase year and in the case of ENSO in the decline stage, the significant precipitation anomalies in most regions of the Yangtze River basin were greater than usual. Especially after the late 1980s and in the summer of an El Niño event in the decline stage, the Yangtze River basin was liable to suffer whole-basin and large-range flood disasters, with the most serious cases occurring in the summer of 1998 especially. This phenomenon was also very obvious in the Poyang Lake basin.

CONCLUSIONS

This study investigated the temporal and spatial variability of seasonal precipitation variations in the Poyang Lake basin, and analyzed the correlation between the seasonal precipitation and climate indices in time and frequency scales. Wavelet transform was used to describe the periodic oscillations. Then, we extended it to different longitude ranges, developed the Hovmöller diagrams for the seasonal precipitation in each meteorological station at different longitudes, and conducted significance tests on them.
The study revealed that oscillation in a band of less than 8 years was the main variability period of the seasonal precipitation in the Poyang Lake basin. In the 2–4 years’ band, the years of rainfall peaks appearing in Poyang Lake were basically consistent with the years when El Niño appeared. In the spatial scale, the summer precipitation oscillation was relatively dramatic. According to the analysis of climate indices, the occurrence of El Niño events created the main atmospheric circulation anomalies favorable for much rain in summer in the Yangtze River basin, and was very likely to cause much rain in summer in the Poyang Lake basin. Similar to the occurrence of ENSO events, NAO, IOD and PDO affected large-scale circulations to produce variability through their respective independent or inter-coupled climate systems, and changed the precipitation distribution in the Yangtze River basin and Poyang Lake basin by affecting local climate conditions such as the East Asian monsoon.

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