A Far-Reaching Footprint of the Tropical Pacific Meridional Mode on the Summer Rainfall over the Yellow River Loop Valley

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(Manuscript received 13 May 2010, in final form 18 November 2010)

ABSTRACT

Over thousands of years, the vicissitudes of the Yellow River and surrounding geographic environment form a unique river loop in northwestern China, namely, the Yellow River Loop Valley. Results from both observations and a climate model indicate that the summer rainfall over the Yellow River Loop Valley region can be remotely controlled by sea surface temperature fluctuations in the eastern subtropical North Pacific. This far-reaching teleconnection is achieved by an atmospheric wave train emanating from the eastern subtropical North Pacific, with a long traveling journey from the western to the Eastern Hemisphere. Furthermore, it is found that the SST forcing pattern resembles the Tropical Pacific Meridional Mode, a mode characterized by an interhemispheric temperature gradient and shift of the intertropical convergence zone in the eastern tropical Pacific.

1. Introduction

Located in the middle of the Yellow River Valley, the Loop region is one of the major agricultural production zones in many Chinese dynasties. Predicting the rainfall over the Yellow River Loop Valley (YRLV) is of significant societal and economic benefit, but few studies so far have been devoted to this specific area. Zhong et al. (2006) analyze some characteristics of the summer rainfall over the YRLV, but what determines its variability remains unclear. In general, the summer rainfall over the eastern part of China is influenced predominantly by the East Asian monsoon system, and its year-to-year variations may involve multiple processes, including surrounding oceanic variability, snow cover over the Tibetan Plateau, and land surface processes.

El Niño–Southern Oscillation (ENSO) has profound influences on the interannual variability of the global climate (Webster et al. 1998) including the East Asian summer monsoon (e.g., Fu and Teng 1988; Ju and Slingo 1995; Zhang et al. 1996; Tao and Zhang 1998). Studies have suggested that ENSO can affect East Asian climate through a Rossby wave–mediated Pacific–East Asian teleconnection in the lower troposphere (Wang et al. 2000; Xie et al. 2009). Based on a singular value decomposition (SVD) analysis, Lau and Weng (2001) found that the summer rainfall variability over China is associated with several different global sea surface temperature (SST) modes, and the 1997/98 rainfall anomalies over China can be explained by the mode representing the growing phase of the El Niño and the mode representing the transition from El Niño and La Niña phases.

The studies mentioned above have pointed out the important role of the central-eastern tropical Pacific
coupled ocean–atmosphere interaction in summer rainfall variability over China. Recent studies have identified a mode characterized by meridional SST gradient coupled with a shift of the intertropical convergence zone (ITCZ) in the eastern tropical–subtropical Pacific, namely, the Tropical Pacific Meridional Mode (TPMM; Chiang and Vimont 2004). This mode differs from the zonal ENSO mode in the seasonal phase lock, time scales, and coupled ocean–atmosphere processes and can trigger an ENSO event (Chiang and Vimont 2004; Chang et al. 2007; Wu et al. 2007, 2010). Here, from both observations and model simulations, we found that the TPMM can profoundly affect the summer rainfall over the YRLV. This far-reaching teleconnection is achieved by an atmospheric wave train emanating from the eastern subtropical North Pacific (ESNP), with a long traveling journey from the western to the Eastern Hemisphere.

The paper is organized as follows. Observational evidences are presented in section 2. Section 3 presents modeling results. The paper is concluded with a summary and some discussions.
2. Observational evidence

The rainfall data used in this study are based on a 160-station monthly rainfall dataset over the mainland of China for the period of 1951–2001 (from the National Climate Center in China). This dataset has been also used by Lau and Weng (2001) in their analysis. The SST data are taken from the National Oceanic and Atmospheric Administration’s (NOAA) extended reconstructed SST (ERSST) (Smith and Reynolds 2003). The atmospheric data used are from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996). To be consistent with the precipitation data, all data used here are from 1951 to 2001.

a. Covariability of the Pacific SST and rainfall over the YRLV

To examine the relationship between the spatial and temporal variability of the rainfall over China and SST anomalies, we first perform a SVD analysis on the summer rainfall over China and simultaneous SST anomalies by following Lau and Weng (2001), but with a focus on the Pacific basin instead of the global SST. The first SVD mode is characterized by an El Niño–like SST pattern and drier-than-normal conditions over North China (Figs. 1a and 1b). This mode accounts for 28.1% of the total squared covariance with the corresponding time series correlated at 0.77 (Fig. 1c). For simplicity, here we define the time series of a SVD mode as the principal component (PC). The second SVD mode, accounting for 17.9% of the total squared covariance, is characterized by anomalous warm SST in the western tropical–subtropical Pacific and wetter-than-normal conditions over the Yangtze River valley (Figs. 1d and 1e). The corresponding PCs correlate at 0.68 (Fig. 1f). These two modes have been identified and discussed in many previous studies (e.g., Weng et al. 1999; Lau and Weng 2001; Li 2008).

The third SVD mode explains 13.5% of the total squared covariance between the rainfall and SST anomalies, and explains 10.4% and 5.9% of the variances of SST and rainfall, respectively. The SST pattern shows a wave train structure in the eastern North Pacific (Fig. 2a), with cold anomalies in the tropical and extratropical North Pacific and warm anomalies stretching from the Gulf of Alaska southwestward to the central subtropical North Pacific, resembling the TPMM (Chiang and Vimont 2004). Coupled with the SST pattern, the heterogeneous rainfall pattern shows positive correlations predominantly localized in the YRLV and some negative correlations in far eastern China between the Yangtze River and the Yellow River Valley (Fig. 2b). The PCs of the rainfall and SST are correlated at 0.72 and display pronounced interannual–decadal variations (Fig. 2c).

To assess the time scales of the coupled ESNP SST–YRLV rainfall mode, we perform the Morlet wavelet transform for the PCs of the coupled mode. The wavelet analysis reveals a pronounced decadal peak around 10 years for both the SST and the rainfall variations (Fig. 3).
Besides 10-yr variability, the rainfall also exhibits substantial biennial variability, while the SST also exhibits bidecadal variability. Overall, the coupled mode is distinguished by the decadal covariability between the ESNP SST and the YRLV rainfall. The decadal variability identified here is consistent with the decadal variation of the TPMM (Wu et al. 2010) as well as the North Pacific basin-scale variation (Qiu et al. 2007).

The ESNP SST–YRLV rainfall mode may be contaminated by the ENSO signals (Wang et al. 2000; Xie et al. 2009). To eliminate the ENSO influences, we subtract the linear regressions of both rainfall and SST anomalies against the prior Spring [March–May (MAM)] Niño-3 SST index from the data and then repeat the SVD analysis. The coupled mode emerges as the second SVD mode, which explains 18.0% of the total squared covariance between the rainfall and SST anomalies, and explains 11.0% and 6.7% of the variances of SST and rainfall, respectively (Fig. 4). For the SST pattern, the wavelike cold–warm–cold SST anomalies from the tropical to extratropical North Pacific remain significant (Fig. 4a), which is very similar to Fig. 2a. For the rainfall pattern, the dominant feature is the strong correlation in the YRLV (Fig. 4b), which is very similar to Fig. 2b. There are 2 PCs that are significantly correlated at 0.75 (Fig. 4c). Compared to SVD analyses under with and without ENSO (Fig. 2 versus Fig. 4), the coupled ESNP SST–YRLV rainfall mode is a mode independent from the impacts of ENSO. It is also noted that some negative rainfall anomalies exists in the Yangtze River valley, which are likely attributed to warm SST anomalies in the western tropical Pacific (Huang and Sun 1992).

To further demonstrate the covariability between the summer rainfall over the YRLV and the SST in the ESNP, we average the summer rainfall anomalies.
observed by 13 weather stations (Huhehaote, Baotou, Xiaba, Tianshui, Minxian, Yulin, Yanan, Xifengzhen, Lanzhou, Yinchuang, Xining, Linxia, and Wuwei) in the YRLV and correlate them with the global SST anomalies (Fig. 5a). The correlation pattern in the Pacific shows a great similarity with the SVD SST mode (Fig. 2a), displaying a north–south dipole in the eastern tropical and subtropical North Pacific. This strongly suggests a potential link between the YRVL rainfall and the eastern tropical–subtropical North Pacific SST variations.

The emergence of correlation in the eastern equatorial Pacific may also indicate a potential mediation of ENSO in the covariability between the summer YRLV rainfall and the TPMM. In this regard, we again eliminate the ENSO influences statistically by subtracting the corresponding linear regression against the Niño-3 SST index from both the SST and the rainfall anomalies and repeat the above correlation analysis (Fig. 5b). In this case, the correlation pattern remains similar to that in the presence of ENSO influences (Fig. 5b versus Fig. 5a). Indeed, the correlation in the ESNP becomes even stronger. This indicates that the ENSO may not be a necessary mediator for the ESNP SST–YRLV rainfall covariability. The wavelet analysis of the time series of the ESNP SST–YRLV rainfall under removal of ENSO (Fig. 4c) further indicates that this coupled mode displays similar decadal (~10 years) covariability as that in the presence of ENSO influences (not shown).

The SST SVD pattern in the eastern tropical subtropical North Pacific region has a strong projection on the TPMM (Chiang and Vimont 2004). To demonstrate that, we first remove the ENSO signal from the SST as described above, and then we perform empirical

![Figure 4](image-url)  
**Fig. 4.** As in Fig. 2, but for the SVD2 of SST and rainfall, with ENSO signals are statistically eliminated by subtracting the linear regressions against the Niño-3 SST index averaged over 5°S–5°N, 150°–90°W in spring (MAM).

![Figure 5](image-url)  
**Fig. 5.** Correlation of the global SST with the YRLV summer rainfall anomaly: (a) with ENSO and (b) without ENSO influences. Contour intervals for SST and rainfall are 0.1 with 0 and ±0.1 contours omitted. Correlations exceeding 0.28 are statistically significant at the 95% level.
FIG. 6. (a) Leading EOF mode of the eastern tropical–subtropical North Pacific summer SST (shaded and white contours with interval at 0.1°C), surface wind (arrows, m s⁻¹) and divergence anomalies (thin black contours, 10⁻⁵ s⁻¹) (obtained by regression against the PC of the leading SST EOF mode), and climatological total precipitation (thick black contours with 5 mm day⁻¹ intervals for values >20 mm day⁻¹). (b) Normalized PC of SST EOF1. (c) Correlation map of the summer rainfall over China and the leading PC of SST EOF1 with 0.1 contour interval with 0 and ±0.1 contours omitted. Shaded areas exceed the 95% significant level based on a Student’s t test.
orthogonal function (EOF) analysis of the eastern tropical–subtropical summer SST. The leading EOF mode is distinguished by a meridional SST gradient with strong SST anomalies in the subtropical region and some modest opposite anomalies in the equator (Fig. 6a). The regression of the surface wind against the PC demonstrates a weakening (strengthening) of the northeastern trade winds associated with a positive (negative) meridional SST gradient (Fig. 6a). The presence of warm anomalies north of the equator sets up an anomalous negative meridional pressure gradient that causes the southwesterly geostrophic wind anomalies to decelerate the northeastern trade winds, thus inducing anomalous convergence northward and a poleward shift of the mean ITCZ (Fig. 6a). The PC exhibits some interannual–decadal variability (Fig. 6b), and is significantly correlated with the PCs of the SST–rainfall SVD mode (Fig. 2c), with spatial correlation coefficients at 0.90 and 0.70, respectively. We further correlate the PC of the TPMM with the summer rainfall over China and reproduce the anomalous rainfall pattern dominated by anomalies over the YRLV and far eastern China between the Yellow River and the Yangtze River (Fig. 6c). The maximum correlation is about 0.5 right at the northernmost of the Loop, indicating about 25% of the local rainfall variability (a local maximum of 55%) can be explained remotely by the TPMM.

To further examine the stability of the coupled ESNP SST and the YRLV rainfall covariability, we calculate an 11-yr running correlation for PCs of the ESNP SST and the YRLV rainfall SVD mode with and without ENSO influences, respectively (Fig. 7). In both cases, the correlations appear to be significant before the early 1970s and after the early 1980s and slightly weaker in between. Overall, despite some decadal variations, the correlation between the ESNP SST and the YRLV rainfall remains significant.

b. Teleconnection from the ESNP to the YRLV

What are the possible mechanisms conveying the remote influences of the ESNP SST to the YRLV rainfall? It is conceivable that anomalous heating associated with the shift of the mean ITCZ induced by the ESNP SST variability can trigger Rossby wave trains to influence remote regions; to demonstrate that, the summer geopotential height anomalies at 300 hPa are regressed against the PC of the TPMM (Fig. 8a). A circumglobal wave train is identified, with the eastern subtropical North Pacific, the eastern United States–western Atlantic, the northeastern Atlantic–western Europe, west-central Asia, northeast Asia, and the North Pacific being positively correlated and with central North America, southern Europe stretching to north-central Asia, and the eastern North Pacific being negatively correlated (Fig. 8a). This circumglobal teleconnection pattern resembles the fifth EOF mode of the summer Northern Hemisphere geopotential height (300 hPa) variations (not shown). It is noted that a southeast–northwest pressure dipole rides over the YRLV. The warm moist air associated with the downstream anomalous anticyclone encounters the upstream anomalous cyclone over the YRLV, setting up sufficient conditions to enhance the rainfall in this area (Fig. 8b). In addition, the convergence of moisture over the north Indian continent will enhance the summer rainfall there (Ding and Wang 2005), which is out of this paper’s scope.

This global-scale teleconnection pattern also shares some similarities with the summer circumglobal teleconnection mode identified in an earlier study by Ding and Wang (2005), although different indices are used. In their study, the heating sources associated with the Indian summer monsoon and ENSO variability are suggested to play a critical role in maintaining this global-scale summer teleconnection. A recent study by Ding et al. (2011) further demonstrates that the shift of ITCZ over the east Pacific may be another source triggering the global-scale teleconnection pattern, although it is somewhat weaker than that of the Indian summer monsoon. Our analysis here provides support for this ESNP forcing mechanism, which can substantiate the summer circumglobal teleconnection. The circumglobal teleconnection triggered by the ESNP SST is indicative of potential Rossby wave propagation along some atmospheric waveguide in the Northern Hemisphere.

To further study the propagation of the Rossby waves, we calculate the stationary Rossby wave activity flux based
on the regressed geopotential height anomaly (Fig. 8a) (Plumb 1985). The wave activity flux over the Northern Hemisphere appears to indicate an eastward wave path from the eastern North Pacific to eastern Asia (Fig. 8c).

To demonstrate the characteristics of the stationary Rossby, we calculate the Rossby wavenumber and group velocity based on the climatological geopotential height at 300 hPa following Ding and Wang (2005). The method was introduced by Hoskins and Ambrizzi (1993). The 300-hPa climatological geopotential height fields are characterized by troughs over the central North Pacific and Atlantic and ridges over North America and north Africa–southern Europe (Fig. 9a). The stationary Rossby wave group velocity core is about 30 m s$^{-1}$ (Fig. 9b) and it takes about 2 weeks to span a circumglobal circle in the midlatitude. As shown in Fig. 9c, the Rossby wavenumber is between 4 and 8 in the midlatitudes of the Northern Hemisphere with a meridional average of 6, consistent with the wave pattern revealed by the regression of the 300-hPa geopotential height anomalies.
against the PC of the ESNP SST (Fig. 8a). A comparison of Figs. 9 and 8 suggests that the Rossby wave propagates eastward approximately along the zonal jet waveguide.

c. Potential predictability of the summer rainfall over the YRLV

The above analysis indicates that the relationship of covariability between the ESNP SST and the YRLV rainfall is robust in summer. However, can we predict the summer rainfall over the YRLV based on the ESNP SST anomaly? To assess the potential predictability, we repeat the SVD analysis between the Pacific SST and summer rainfall over China, but with the SST leading the rainfall by 1–3 months.

The SVD mode of the Pacific SST averaged from May to July and the YRLV summer (June–August) rainfall explains 12.9% of total squared covariance, and exhibits a similar pattern with that at 0 lag (Fig. 10). Despite similarities, it is also noted that the ESNP SST anomalies are somewhat weaker and the equatorial SST anomalies are stronger (Fig. 10a versus Figure 2a). The summer rainfall pattern over China, but with the SST leading the rainfall by 1–3 months.

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The above analyses may suggest that the summer rainfall over the YRLV may be predicted based on the SST anomalies over the eastern subtropical North Pacific by about 2–3 months ahead. However, it remains elusive why the prediction time scale is only about 2–3 months. To address this question, we examine the persistency barrier following Chen et al. (2007) to calculate the autocorrelations of the SST index of the TPMM (Table 1).

The TPMM SST index is obtained by projecting monthly SST anomalies onto the SST SVD mode (Fig. 2a). In Table 1, the autocorrelations statistically significant above the 95% level are highlighted, and the freedom of the time series (estimated by following Bartlett 1935) is shown in brackets. The length of the highlighted column for a particular month can be regarded as a measure of SST memory starting from that month (Lau and Yang 1996). It can be seen that the TPMM SST anomalies initiated in late winter and early summer can persist 6–12 months, while in late summer and early winter it can persist 3–5 months. Therefore, a late-winter barrier exists for the SST anomalies initiated from August to December. It is this barrier that limits the predictive time scale of the YRLV rainfall anomalies based on the TPMM SST anomalies.

3. Modeling results

The observational analyses above suggest that the ESNP SST may be used as a predictor of the summer rainfall in the YRLV with a lead of about one season. To further assess this potential predictability, we design a set of AGCM experiments with the ECHAM5 model.

The ECHAM5 model is developed at the Max Planck Institute for Meteorology, which originally evolved from the spectral weather prediction model of the European Centre for Medium-Range Weather Forecasts (ECMWF) (Roeckner et al. 2003). The model employs a spectral dynamical core. A hybrid coordinate system is used in the vertical direction with the sigma coordinates at the lowest model level gradually transforming into pressure coordinates in the lower stratosphere. The model used here has T42 resolution and 19 vertical levels. The detailed description of the ECHAM5 model is given by Roecker et al. (2003).

The model successfully captures the climatological summer geopotential height with troughs in the central North Pacific and Atlantic (Fig. 11a), a stationary Rossby wave group velocity with a maximum of about 35 m s$^{-1}$ in the midlatitude of the Northern Hemisphere (Fig. 11b), and a stationary Rossby wavenumber with a meridional mean of 5–7 (Fig. 11c).

To dynamically assess the relationship between the ESNP SST anomalies and the summer YRLV rainfall,
two sets of ensemble AGCM experiments are carried. In the first experiment, the AGCM is forced by a fixed SST anomaly from May to September over the ESNP (Fig. 12a). The SST anomaly has the same pattern and magnitudes as the TPMM mode in the eastern subtropical North Pacific (Fig. 6a). The experiment consists of 10-member ensemble runs, with each run starting from a different initial state. In the second experiment, the AGCM is forced by the observed SST anomalies from 1956 to 2001 over the ESNP. This set of experiments consists of 13-member ensemble runs. We will focus on the summer atmospheric teleconnection and the rainfall over the YRLV.

For the fixed positive SST anomaly forcing in the ESNP, the ensemble-mean response of the summer rainfall over China is largely localized over the YRLV extending to the eastern part of Inner Mongolia (Fig. 12b). The positive summer [June–August (JJA)] rainfall anomaly has a maximum amplitude of 0.6 mm day$^{-1}$ right over the YRLV. In addition, drier-than-normal conditions are also found in the southeastern flank of the wetter region (Fig. 12b). In general, the ensemble-mean rainfall response is similar to the rainfall pattern of the SVD mode (Fig. 2b). The ensemble-mean response of the geopotential height at 300 hPa displays a wave pattern with a wavenumber 6 in the midlatitudes of the Northern Hemisphere, with
positive anomalies over the eastern North Pacific, the western North Atlantic, western Europe, eastern China–western Pacific, and with negative anomalies over North America, the eastern North Atlantic, northeastern Asia, and in the vicinity of the Alaska peninsula (Fig. 12c). In general, the simulated dynamic response is similar to the observed atmospheric circumglobal teleconnection (Fig. 8a). Therefore, the modeling results provide strong support for the relationship between the ESNP SST and summer rainfall in the YRLV region derived from the statistical analyses.

With the observed historical SST forcing over the ESNP, the ensemble-mean response of the rainfall over the YRLV significantly correlates with the observed rainfall anomalies, with a maximum correlation at 0.40 (Fig. 13a) and a regional-mean correlation at 0.33 (Fig. 13b), which exceed the 95% significance level using a Student’s t test. This suggests the ESNP SST can impact the rainfall over the YRLV.

To examine the Rossby wave teleconnection mechanism, we calculate the regression of simulated geopotential height anomalies at 300 hPa against summer SST anomalies over the ESNP (Fig. 13b) and also the leading EOF of the simulated summer geopotential height at 300 hPa (Fig. 13c). Despite some notable differences, both the regression and the leading EOF mode show a similar circumglobal wave train, which correlates with the observed (Fig. 8a) at 0.6 and 0.7, respectively. Furthermore, it is found that none of the EOF modes in the control simulation in which the AGCM is forced by climatological SST (not shown) resembles the circumglobal teleconnection pattern. This suggests an important role of the

### Table 1. Autocorrelation of the TPMM index for the period 1951–2001 as functions of the 12 calendar months with a lag time of 1–12 months. Autocorrelation values significant at the 95% level are denoted by bold font. The degrees of freedom of the time series are shown in parentheses.

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4. Summary and discussion

From both observations and climate model simulations, we found that summer rainfall over the YRLV can be significantly impacted by the Tropical Pacific Meridional Mode. The migration of the mean ITCZ associated with the fluctuations of the northeastern trade winds can induce anomalous heating anomalies, which can trigger an atmospheric wave train to convey the ESNP SST influence to the YRLV region. The atmospheric teleconnection bears strong similarities to the summer circumglobal teleconnection identified by Ding and Wang (2005).

Our study here suggests the summer rainfall over the YRLV can be partly predicted by the ESNP SST variations. Then the question is what are the mechanisms driving SST changes in the ESNP? One important forcing is the North Pacific air–sea interaction. The winter North Pacific atmospheric circulation change can induce an SST anomaly in the ESNP in spring, which can be subsequently amplified by wind–evaporation–SST (WES) feedback (Xie and Philander 1994) or seasonal footprint feedback (Vimont et al. 2001; Chiang and Vimont 2004; Chang et al. 2007; Wu et al. 2007, 2010). Studies have indicated that the TPMM peaks in early summer, in contrast to fall for ENSO. This so-called seasonal footprint process (Vimont et al. 2001) can extend the predictability of the summer rainfall over the YRLV by a few months.

Our study here also provides strong support for the TPMM forcing mechanism for the summer global-scale atmospheric teleconnection (Ding et al. 2011). Past studies have identified the important role of ENSO and the Indian summer monsoon variability in driving this teleconnection (Ding and Wang 2005).

Our AGCM experiments also indicate that the TPMM can affect the summer rainfall in the north Indian Ocean (not shown), consistent with the observed stationary wave activity flux in that region associated with the TPMM (Fig. 8c). Therefore, it is possible that the TPMM can influence rainfall over the North Indian Ocean, which can further affect rainfall over northern China. Given the strong coupling between the TPMM and the north–south migration of the ITCZ in summer, the induced diabetic heating can sustain the summer global-scale atmospheric teleconnection mode.

Acknowledgments. This work is supported by the National Science Foundation of China (NSFC) Distinguished Young Investigator Project (40788002). Chun Li is supported by the Open Research Program of LASG and NSFC Grants (40906003, 40830106). We are indebted to both reviewers for their constructive comments, which greatly improved the paper in many aspects. Discussions with Drs. Guoxiong Wu and Bin Wang were helpful.
FIG. 13. (a) Correlation between the simulated summer rainfall and the observed over the YRLV. (b) Time series of the observed (dashed black) and simulated (gray and solid black) summer rainfall anomalies averaged over the YRLV [dashed black box in (a)]. The solid black curve denotes the ensemble mean of the model-simulated rainfall anomalies. (c) Regression of the simulated ensemble-mean 300-hPa geopotential height anomalies against the SST over the ESNP in summer. (d) Leading EOF (13.2%) of the simulated ensemble-mean 300-hPa geopotential height in summer. Shaded areas in (c) exceed the 95% significance level.
REFERENCES


