Variations of Teleconnection of ENSO and Interannual Variation in Summer Rainfall in the Central United States

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ABSTRACT

Summer rainfall in the central United States has singular interannual variations of a 3–6-yr period. Identifying the causes of these variations assures improvement in predictions of summer rainfall in the region.

A review of previous studies revealed a puzzling situation: the outstanding interannual variations of the summer rainfall in the central United States showed no persistent correlations with known influential interannual variations in the Northern Hemisphere and the El Niño–Southern Oscillation (ENSO). This study was undertaken to identify the cause of this situation and ultimately explain the causes of the observed interannual summer rainfall variations. Its results showed a teleconnection of the ENSO with the summer rainfall in the central United States. The intensity of which has varied over the last 125 years. The teleconnection was active in two epochs, 1871–1916 and 1948–78, and absent in the two epochs 1917–47 and 1979–present. This variation was associated with a multidecadal variation in both sea surface temperature and sea level pressure in the mid- and high-latitude North Pacific. In the epochs of active teleconnection, the circulation in the warm phase of ENSO favored a deformation field in the lower troposphere in the central United States causing wet summers and a reversed circulation in cold phase of ENSO yielding dry summers, a process that partially explains the interannual summer rainfall variations.

The result also showed that the variations of the teleconnection were ''in phase'' with the variation in the average surface temperature of the Northern Hemisphere. When the ''abrupt warming'' of the surface temperature developed in 1917–47 and the most recent two decades, the teleconnection broke down. Because of the limitation in data record length, this observed relationship and the persistence of the variation in the teleconnection need further investigations when additional data are available.

1. Introduction

As the impacts of droughts and floods on agriculture, water resources, and environment capture the attention of the scientific community and demands for water outstrip supplies in growing regions of the world, efforts to study and predict regional and global precipitation variations of interannual and longer timescales have increased. Many studies have examined the relationship between variation in sea surface temperatures (SST) in the Pacific Ocean and interannual variations in annual and seasonal precipitation in various regions of the United States. The rationale is that 1) SST variations in the equatorial Pacific Ocean are dominated by the quasi-periodic El Niño–Southern Oscillation (ENSO) cycle of 3–6 years; and 2) oceans have longer memory than do lands. Through atmospheric teleconnections, the interannual variations in tropical SST can affect similar variations in precipitation over land.

A number of studies have identified statistically significant correlations of SST anomalies in ENSO with interannual variations in precipitation, winter snow, and streamflow in various regions of the United States and Canada (e.g., Ropelewski and Halpert 1986; Harrington et al. 1992; Hereford and Webb 1992; Bunkers et al. 1996; Brown and Goodison 1996; Cayan et al. 1999; Haston and Michaelsen 1994; Roswintiarti et al. 1998). They also showed that the correlations are weakest in the central United States, an area defined as 37.5°–45.0°N, 90°–105°W. There have been no explanations of this result.

Studies of precipitation variations in the central United States have primarily focused on extreme summer rainfall anomalies, that is, droughts and floods. They found three major sources contributing uniquely to these anomalies: the SST anomalies in the tropical Pacific (Trenberth and Guillemot 1996; National Research Council 1998), the SST anomalies in the mid- and high-latitude North Pacific (Namias 1983, 1991), and the low-level southerly jet (LLJ) from the Gulf of Mexico (Helfand and Schubert 1995; Bell and Janowiak 1995; Mo et al. 1995, 1997; Paegle et al. 1996). Additional studies also showed that local moisture supplies and surface conditions sometimes can have a sizeable contribution

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to enhancing summer rainfall anomalies (Brubaker et al. 1993; Kunkel et al. 1994, 1996). These results showed the very different and competing processes important in causing summer rainfall variations in the central United States. They may explain the lack of a persistent and singular relationship between the summer rainfall variations and any single source and process.

These very different influences resulted in complicated interannual variations in summer rainfall in the central United States. This became clear from comparisons of existing studies on the subject. For example, Ting and Wang (1997, see their Fig. 7) showed a good correlation of summer rainfall in the central United States versus SST associated with El Niño using 1950–90 data, whereas Ropelewski and Halpert (1986, see their Fig. 7) found no consistent relationship between the two variations in a dataset covering a different period, 1875–1980. This situation was made even more perplexing when we found significant interannual variations in summer rainfall in the central United States [Hu et al. (2000), see Fig. 2b of this article]. It indicates that the major climate processes that affect the region's interannual variations in summer rainfall are yet to be identified.

There is evidence that the intensity of ENSO varied on decadal scales (Trenberth and Shea 1987). Recently, it was found that ENSO variations also have been modulated by decadal and longer–timescale variations in the North Pacific Ocean region. Latif and Barnett (1996) showed that an unstable process in ocean–atmosphere interactions in the North Pacific resulted in a multidecadal (20 yr) SST variation in that region that affected North America's weather and climate. Zhang et al. (1997) extracted an ''ENSO-like'' interdecadal variation in the North Pacific region based on data for 1900–93. Distinctive SST patterns stood out in different phases of the variation. Moreover, Gershunov et al. (1998, 1999) showed that the North Pacific decadal oscillation (PDO) modulated the winter ENSO teleconnections. Particularly, in high phases of PDO, when winter-season SST in the midlatitude North Pacific was cooler than normal, the El Niño effect on western North American weather was strong and La Niña effect was weak. In low phases, the importance of the effect of El Niño and La Niña reversed.

Similar selective responses of weather and rainfall to different phases of multidecadal or longer-scale variations also were noticed in other phenomena. For instance, the well-known correlation of the ENSO-Indian monsoon intensity varied over the last century; and it weakened in the last 20 years, due partially to increase of surface temperature and eastward shift of the Walker Circulation (Kumar et al. 1999). In North America, Cole and Cook (1998) found variations of correlation between ENSO and summer dryness in the contiguous United States, measured by the Palmer Drought Severity Index, and a weakening of the correlation in the recent decades.

The modulations of teleconnections of ENSO by longer-timescale variations could be an explanation of the puzzling situation in the summer rainfall variations in the central United States. Our hypothesized process is this: In some decades the effect of boreal summer SST anomalies during ENSO was strong on summer rainfall variations in the central United States. In other decades, the effect was weak or absent. In the period when the ENSO teleconnection was active, summer rainfall in the central United States had interannual variations resulting from the ENSO related SST variations. In the decades when this teleconnection was subdued, effect from other processes of different sources dominated the summer rainfall variations in the region. Because of this variation of the teleconnection, there could be a weak or no correlation between ENSO and the central U.S. summer rainfall when the correlation was taken over a period covering several such cycles. In the meantime, an interannual variation of summer rainfall in the region would appear if the processes of the non-ENSO sources also have interannual variations.

There are several major questions to answer in order to validate this hypothesis. 1) Are there variations of teleconnection of the boreal summer tropical Pacific SST anomalies associated with ENSO cycle and the summer rainfall anomalies in the central United States? 2) If so, what processes have regulated them? 3) What processes have affected the region's summer rainfall when ENSO effects weakened? 4) Do those processes in question 3 have significant interannual variations? We will address the first two questions in this study with available data and relevant diagnostics. The last two questions will be answered separately elsewhere.

We describe the data and analysis methods used in this study in the next section. In section 3, we examine summer rainfall variations in the central United States and their relation to SST variations in the Pacific Ocean to address the first and second questions. Composite sea level pressure (SLP) in the boreal summer will be evaluated in section 4, emphasizing the teleconnection of ENSO with summer rainfall variation in the central United States. This discussion is further extended to the dynamic fields in the lower troposphere in section 5. Section 6 contains further discussion and a summary of the study.

2. Data and methods

Data used in this study were monthly total precipitation over the central United States, the global SST, the Northern Hemisphere (NH) SLP, and 850-hPa geopotential height and wind. All data, except for precipitation, were averaged over the three summer months, that is, June, July, and August (JJA), to obtain summer mean values. Summer rainfall was the total of JJA.

The monthly precipitation data were obtained from Dai et al. (1998). They were in a gridded format of 2.5° × 2.5° resolution and covered the globe for 1871–1995.

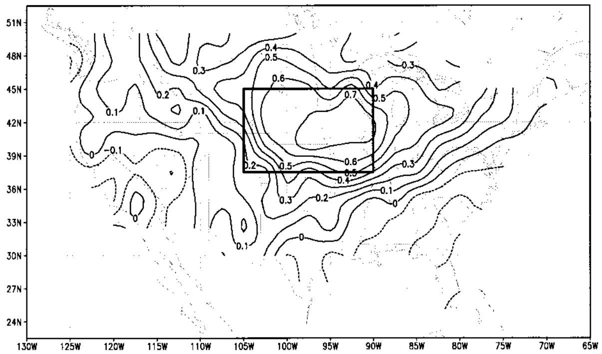

FIG. 1. Distribution of correlation of the central U.S. summer rainfall and gridpoint rainfall in the contiguous United States for the period 1871–1995. The region confined by the thick solid line shows the central United States discussed in this study.

Dai et al. (1997) discussed the procedure used to produce this dataset. We examined the summer rainfall data for the central United States extracted from this dataset against station data in the region and found identical variations (figure not shown). Because the station data were from the same stations with such long records, this consistency test shows that the sampling procedure used in Dai et al. (1997) did not create significant bias variations in the dataset.

Monthly SST data were from the Global Sea-Ice and SST dataset (GISST) version 2.3b from the Met Office (Parker et al. 1995; Rayner et al. 1996). The data covered the period 1871–1998 and had a $1.0° \times 1.0°$ resolution over the globe. Hurrell and Trenberth (1999) showed significant differences between the GISST and National Centers for Environmental Prediction (NCEP) 1961–90 SST data at some regional scales, for example, the Gulf of Mexico area and inter-sea-ice zones in high latitudes. They also showed that the trend of GISST SST after 1981 was different from that of the NCEP SST dataset. They recommended that a 3- or 5-month running mean be applied to the GISST monthly data after 1981 to minimize these differences. Accordingly, we used a 3-month running average of the SST data in our analysis. The SST data were resampled from the $1.0° \times 1.0°$ resolution to a $5.0° \times 5.0°$ resolution to emphasize large-scale features in SST variations.

Data for SLP were obtained from the National Center for Atmospheric Research (NCAR Dataset ds010.1) (Trenberth and Paolino 1980). Data resolution was $5.0° \times 5.0°$ over the NH (15°–90°N), and time length was 1899–1997. We used the NCEP–NCAR reanalysis (Kalnay et al. 1996) geopotential height and wind in evaluation of lower troposphere circulations. The reanalysis data started in 1958.

Both SST and SLP data were detrended in our analysis. The rainfall in the central United States showed no clear trend and thus no detrending was applied to the rainfall data. The SST anomalies (SSTA) were calculated following the procedure described in Zhang et

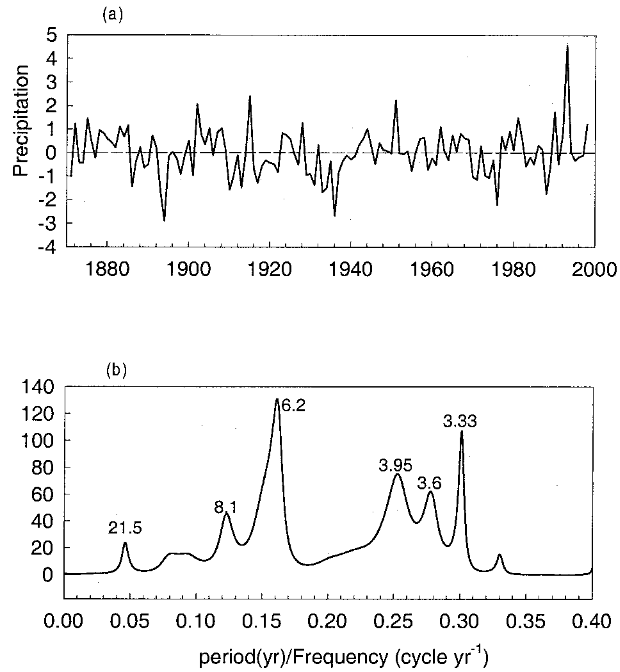

FIG. 2. (a) Normalized time series of central U.S. summer rainfall (1871–1998) [precipitation data for 1871–1995 were from Dai et al. (1998) and for 1996–98 were derived from Higgins et al. (1998)], and (b) its maximum entropy spectrum.

al. (1997). We used composite and statistical evaluations including correlation, moving correlation, and significance tests in diagnosis.

3. Summer rainfall in the central United States and its relation with SST variations

Variations in summer rainfall in the central United States are significantly different from variations in other regions. This is shown in Fig. 1 of simultaneous correlations of the region's average summer rainfall and rainfall at individual grid points of a $2.5° \times 2.5°$ network over the contiguous United States. The length of data record used in this correlation is 125 years (1871–1995). In Fig. 1, the area encompassing Nebraska, South Dakota, Iowa, Illinois, parts of Missouri, and Kansas shows consistent summer rainfall variations (correlation coefficient > 0.6). This area also has distinct interdecadal variations of annual precipitation (Hu et al. 1998). Weak correlations outside this region indicate summer rainfall variations different from that in the central United States. The correlation distribution in Fig. 1 is similar to a one-point correlation map in Ting and Wang (1997), derived using a short data series of 40 yr (1950–90). These results indicate a consistent variation in summer rainfall in the central United States.

Figure 1 also shows negative correlations of variations in summer rainfall between the central United States and the west coastal areas, the southwest monsoon region (Carleton et al. 1990), and the Great Basin.

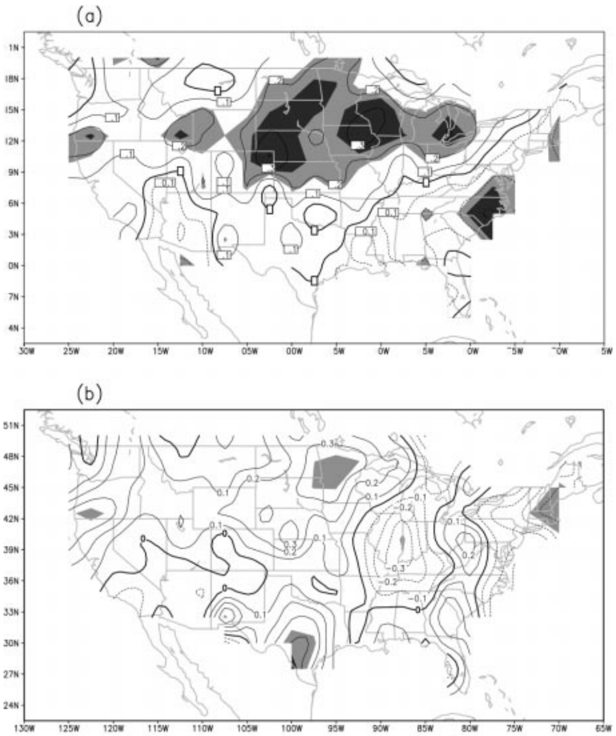

FIG. 3. Correlations of boreal summer Niño-4 and Niño-3.4 SSTA (from the western boundary of Niño-4 to eastern boundary of Niño-3.4) and U.S. gridpoint summer rainfall anomalies over (a) the period 1871–1995 and (b) the period 1917–47.

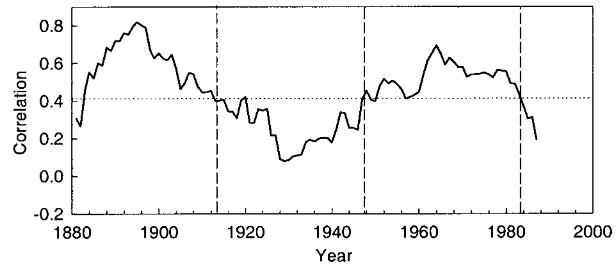

FIG. 4. A 21-yr sliding window correlation of boreal summer Niño-4 and Niño-3.4 SSTA (similar to Fig. 3) and central U.S. summer rainfall. The dotted line is the 95% significance level, based on the Student's *t* distribution for the null hypothesis of no association.

This out-of-phase relationship was found in several previous studies (Tang and Reiter 1984; Mo et al. 1997; Higgins et al. 1997, 1998). However, a comparison of Fig. 1 with the results in the previous studies indicates a much weaker correlation in Fig. 1.[1] Figure 1 also indicates a negative correlation between summer rainfall in the central United States and that in the east and southeast coastal areas of the United States.

Figure 2a displays the temporal variation of the normalized summer rainfall in the central United States (hereinafter ''the summer rainfall''). Figure 2b is the result of a spectral analysis of the time series in Fig. 2a. It shows outstanding variations in the summer rainfall at interannual scales of 3–6 yr.

We examined the correlation between the variations in the summer rainfall and the SST in the central and eastern equatorial Pacific Ocean. The latter is strongly associated in the ENSO cycle with a quasi-periodicity of 3–6 yr, and has been considered as a major process affecting interannual climate variations over the globe. Figure 3 summarizes the results and shows that the correlation and the influence of ENSO have varied over time. An apparently good correlation was found for the period 1871–1995 (Fig. 3a), whereas no significant correlation was found in various segments of that period. One such case is shown in Fig. 3b for the period 1917–47. These results demonstrate that from a statistical perspective, there is no *persistent* influence of the SST variation in ENSO on the summer rainfall variations in the central United States, over the last 125 years.

The variation of the effect of the SST in the tropical Pacific on the summer rainfall is not necessarily inconsistent with the presence of significant interannual variations in the summer rainfall (Fig. 2). Because correlations do not take into account the phase differences in different variations, the interannual variations will remain in the statistical result if other sources will maintain the variations when the ENSO effect weakens.

To measure the temporal variation of the influence of teleconnection of ENSO on interannual variations in the summer rainfall, we calculated 21-yr moving correlations of variations in the summer rainfall versus the SST in the Niño regions. The results (Fig. 4) show a strong correlation (above 95% significance level) in two periods, 1880 to the mid-1910s and 1948 to the late 1970s and early 1980s and an absent correlation in other two periods, the mid-1910s to 1948 and the most recent two decades since 1980.

Before further discussing these changes we want to indicate an interesting observational fact. We have noted that the variation in the correlation in Fig. 4 has been ''in phase'' with the variation in the average NH surface temperature in the last 125 years shown in Ghil and Vautard (1991). Particularly, the two periods with significant positive correlations (Fig. 4) coincided with the time of nearly steady or slightly cooling temperatures in the NH, and the two periods with subdued correlations happened coherently in the time of ''abrupt warming'' of the NH surface temperatures.

According to the variations in Fig. 4, we divided the years 1871–1995 into four periods: 1871–1916, 1917–47, 1948–78, and 1979–95, and will refer to them in the following discussions as epochs 1, 2, 3, and 4, respectively. In Fig. 5, we illustrate for each epoch the spatial correlation of boreal summer SST anomalies (SSTA) in the Pacific and anomalies of the summer

[1] Our further analysis of this difference suggested it was due to an unstable correlation between rainfalls in these regions similar to that discussed in section 6 of this article.

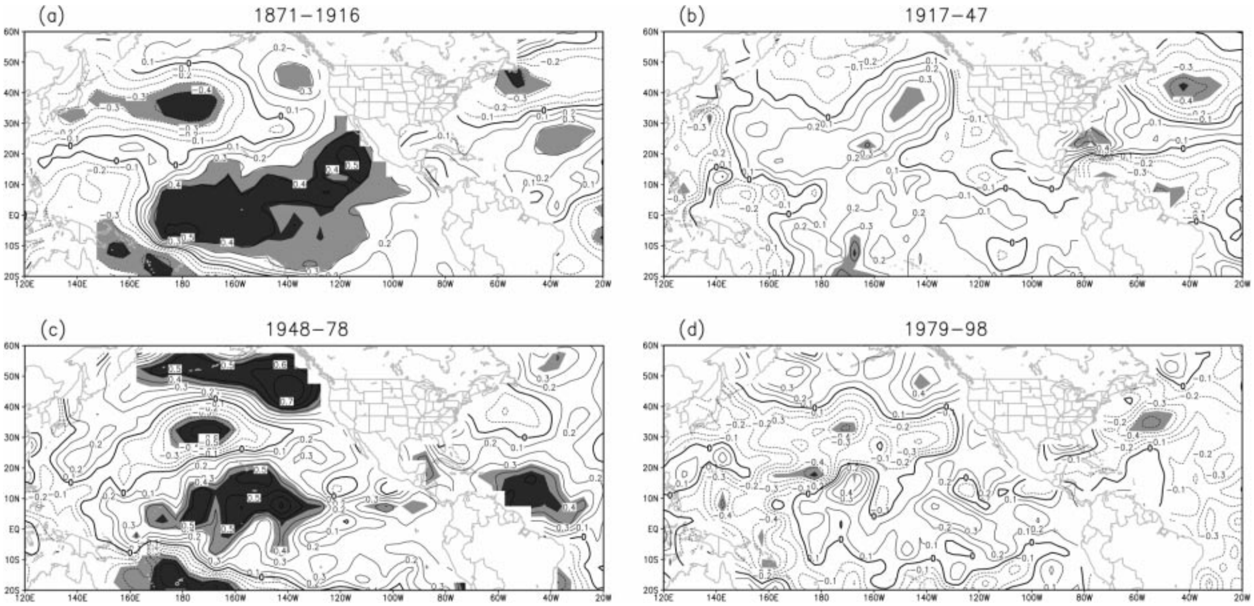

Fig. 5. Correlation coefficients of central U.S. summer rainfall and the boreal summer SSTA for (a) 1871–1916, (b) 1917–47, (c) 1948–78, and (d) 1979–98. Contour interval is 0.1. Light-shaded areas show correlations with greater than 95% confidence level, and dark-shaded areas show correlations with greater than 99% confidence level.

rainfall in the central United States. The first and third epochs (Figs. 5a, 5c) each had a significant positive correlation of the summer rainfall with tropical Pacific SSTA and a similar spatial distribution of the correlation. Positive correlations stretched from the Niño-4 and Niño-3.4 area northeastward to the west coasts of California and Mexico and also were found in the northeastern North Pacific Ocean. The SSTA in most of the midlatitude North Pacific had a negative correlation with the summer rainfall. There are minor differences between the two panels, possibly due to differing climate status in the two epochs (e.g., the latter was warmer than the former).

In great contrast to the above correlation pattern, there was absence of correlations of SSTA in the tropical and North Pacific Ocean versus the summer rainfall in the central United States in the second epoch (Fig. 5b). There was no ''organized'' pattern in the Tropics and low latitudes. Although a sign reverse in correlations was visible in the mid- and high-latitude North Pacific Ocean, the correlation was insignificant. The correlation broke down in this epoch.

Epoch 4 (Fig. 5d) had a correlation distribution similar to epoch 2 with a lack of any correlation pattern in the North Pacific Ocean. We noted that this epoch is about one-half the size in sample years of the other three epochs, because of data availability, and thus may not be fully comparable with the early epochs. Our comparisons of Fig. 5d with Figs. 5c and 5b suggested that the correlation change in epoch 4 reflects a transition away from the third epoch, apparently in synchronicity with the recent strong warming of the NH surface temperature after the late 1970s.

Figure 6 shows the SSTA distributions in the four epochs. From these distributions, we easily find the major features of the SSTA in each epoch. A center of positive SSTA was in the central North Pacific Ocean in both epochs 1 and 3, with surrounding negative SSTA in the east, north, and south in the epochs of high correlation (Figs. 6a and 6c). An opposite SSTA distribution in the mid- and high-latitude North Pacific was in Figs. 6b and 6d for both epochs 2 and 4. The SSTA variations in the eastern North Pacific off the coasts of the southwestern United States were consistent with that in the central North Pacific but with reversed anomalies between opposite epochs. Another noticeable feature in the panels of Fig. 6 is a significant increase in the magnitude of SSTA in the most recent two epochs.

It is necessary to also point out that there were no consistent variations in SSTA in the Atlantic region between these epochs.

The SSTA differences between sequential epochs in Fig. 6 (e.g., SSTA in Fig. 6b minus SSTA in Fig. 6a, and so forth) are shown in Fig. 7 for the Pacific and North America region. The same two features stand out, 1) the flip-flop of the sign of SSTA in the central North Pacific and the eastern North Pacific off the coasts of the southwestern United States in opposite epochs, and 2) an amplification of the SSTA in the recent two epochs.

The shading in Fig. 7 shows statistically significant SSTA changes between the epochs (above 95% signif-

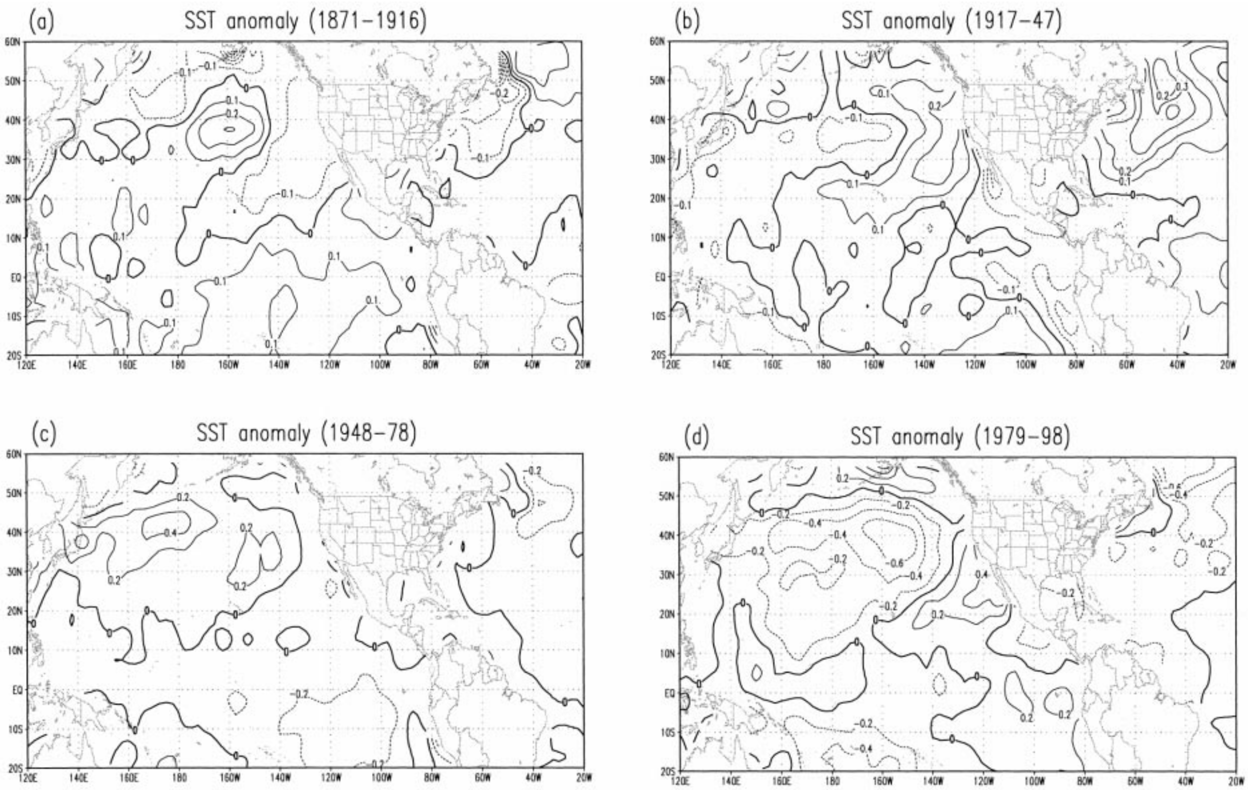

FIG. 6. Mean boreal summer SSTA for (a) 1871–1916, (b) 1917–47, (c) 1948–78, and (d) 1979–98. Contour interval in (a) and (b) is 0.1 K and is 0.2 K in (c) and (d).

icance level based on Student's *t* test). Two regions in the North Pacific Ocean with significant changes are 1) the eastern North Pacific off the coasts of the southwestern United States stretching southwestward to the central tropical Pacific Ocean and 2) the central North Pacific Ocean. These test results assure the variations in SST are significant in the North Pacific Ocean between the opposite epochs and thus, confirm a significant cyclic variation in SSTA in these epochs. The phase of this variation matched very well with changes in the correlation of SSTA in the Pacific and the summer rainfall variations between different epochs in Fig. 5, a result indicates that the variations in SST in the mid- and high-latitude North Pacific Ocean play an important role in facilitating the teleconnection of the tropical Pacific SSTA and the summer rainfall in the central United States.

In summary, the multidecadal variation in mid- and high-latitude SSTA in the North Pacific was consistent with the quasi-periodic variation of the correlation between boreal summer SSTA in the tropical Pacific and the summer rainfall in the central United States. When the SSTA in the mid- and high-latitude North Pacific Ocean had a pattern like that shown in Figs. 6a and 6c for the time periods 1871–1916 and 1948–78, the correlation was high for the SSTA in the tropical Pacific versus the summer rainfall. Effects of the interannual variations of SSTA related to ENSO could reach to the central United States and contribute to interannual variations in the summer rainfall. The correlation of the SSTA in the mid- and high-latitude North Pacific Ocean and summer rainfall in the central United States also was high. When the SSTA in the North Pacific had a pattern like that shown in Figs. 6b and 6d for the time periods 1916–47 and 1979–98, the correlation broke down. During these epochs, the effects of interannual SST variations related to the ENSO cycle in the tropical Pacific on the summer rainfall in the central United States weakened or disappeared. These facts invite the conclusion that the mid- and high-latitude North Pacific Ocean played a ''bridging'' role in the teleconnection of tropical interannual SST variations and the variations of the summer rainfall in the central United States.

Now, what processes could have sustained the interannual variations of the summer rainfall when the teleconnection broke down? This was the third question we raised in the introduction. Our preliminary analysis suggested an increase in correlation of the summer rainfall with local SST variations in the Gulf of Mexico and the Caribbean region when this teleconnection broke down. The effects of the Gulf region, through the well-known LLJ, were strong in many individual extreme precipitation events studied (Brubaker et al. 1993; Helfand and Schubert 1995; Mo et al. 1995). Apparently,

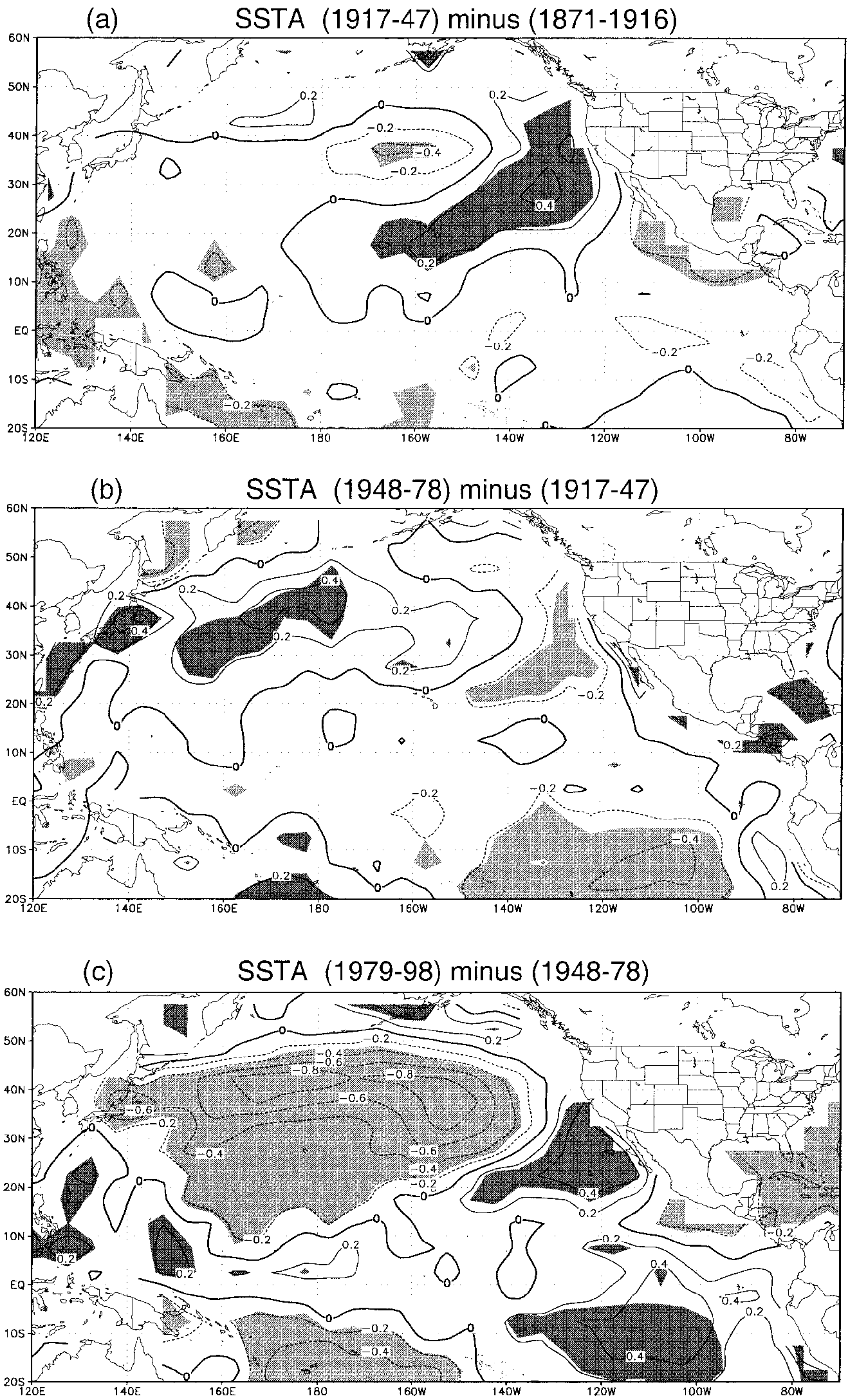

FIG. 7. Differences of summer SSTA from (a) 1917–47 minus 1871–1916, (b) 1948–78 minus 1917–47, and (c) 1979–98 minus 1948–78. Contour interval is 0.2 K. The shaded areas indicate that differences between epochs are above 95% statistical significance level with dark shade for positive regions and light shade for negative regions.

such effects were enhanced when the effects from the North Pacific region weakened. Details of these effects and the questions as to whether these effects lead to interannual variations of summer rainfall in the central United States will be presented separately elsewhere.

4. Low-level features of the teleconnection

In this section, we examine the SLP variation and its relationship with the SST changes shown before. We show in Fig. 8 the average spatial patterns of detrended

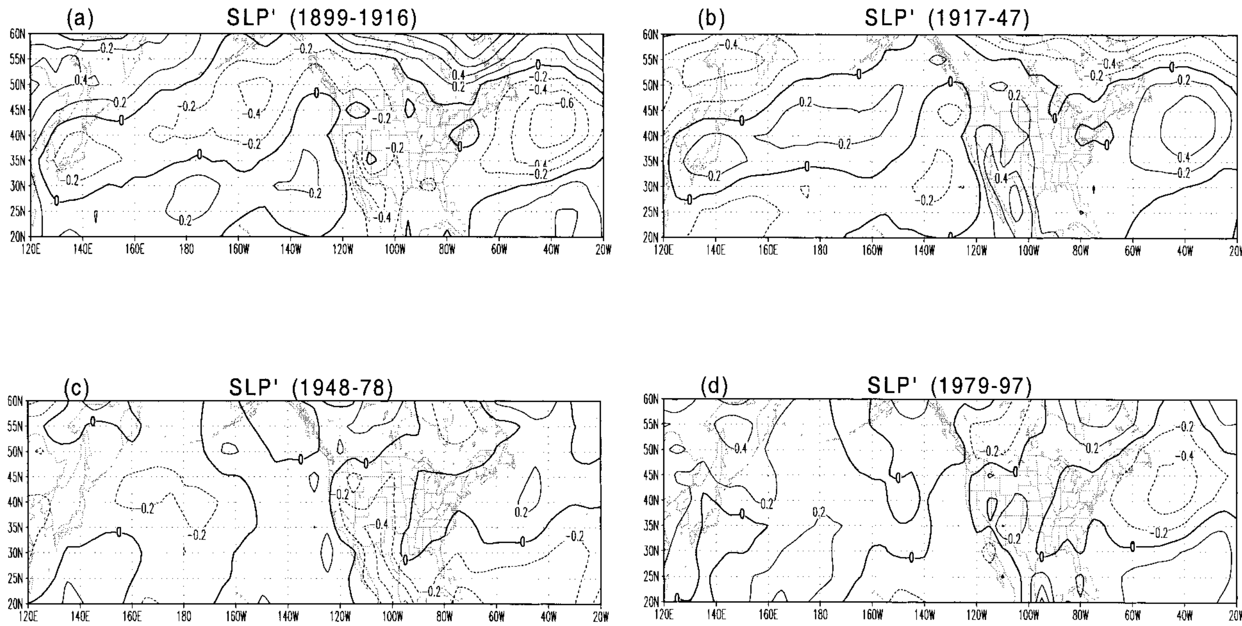

FIG. 8. Mean summer SLP′ of (a) 1899–1916, (b) 1917–47, (c) 1948–78, and (d) 1979–97. The contour interval is 0.1 hPa.

SLP anomalies (SLP′) for the four epochs in Fig. 6. Each panel corresponds to the same lettered (a–d) panel or epoch in Fig. 6. Although sea level is always at or below ground in land areas, SLP may be considered a good ''proxy'' for the low-level atmospheric mass field in land areas. This analysis of SLP facilitates a consistent interpretation of low-level atmospheric mass field for both ocean and land areas.

Figure 8 shows that from the high-correlation epochs (Figs. 8a and 8c) to the low/no-correlation epochs (Figs. 8b and 8d) the gradient of SLP′ reversed in the eastern and central North Pacific. Figures 8a and 8c show positive SLP′ in the eastern North Pacific (off the west coasts of the United States) and negative SLP′ in the region west of this positive SLP′ region. In contrast, Figs. 8b and 8d show negative SLP′ in the eastern North Pacific and positive SLP′ in the central North Pacific. Although the orientation of the SLP′ distributions changed in the later two epochs from the early two, the above feature remained the same. In the region 120°–160°W off the west coasts of the United States, the SLP′ varied from above normal to below normal from Figs. 8a to 8b and repeated the cycle from Figs. 8c and 8d. This cyclic SLP′ variation was in-phase with the variation in SSTA in the region going from cool → warm → cool → warm through the epochs (Fig. 6). In the central North Pacific, an analog variation with opposite signs in the SSTA and SLP′ was clearly shown in Fig. 8.

Over the United States, low pressure anomalies prevailed during epochs 1 and 3 (Figs. 8a and 8c) with a center in northern Mexico and the southwestern United States. These anomalies reversed sign in epochs 2 and 4 (Figs. 8b and 8d).

In the Atlantic region, there was again a lack of consistent variation through the epochs, except for a reversal of the SLP′ field in the most recent two epochs after 1948 (Figs. 8c and 8d).

The changes in SLP′ between the epochs are further shown in Fig. 9. The significant changes suggest alternations in the atmospheric mass and motion fields in the western United States and Mexico and the eastern North Pacific between opposite epochs. The SLP′ pattern in epochs 1 and 3 (see Fig. 8) facilitated effects of the SST variations in the equatorial Pacific to reach the central United States and influence its summer rainfall. Interannual variations of summer rainfall in the central United States developed following that of the SSTA associated with the ENSO cycle. The mechanisms of this teleconnection are beyond the scope of this paper. When the SLP′ pattern reversed in epochs 2 and 4 in association with the changes in SSTA in the North Pacific Ocean (Fig. 7), the teleconnection decreased.

5. Interannual variations of summer rainfall in the central United States

We examined further the interannual variations of summer season low-level geopotential height and motion fields in epochs 1 and 3 when there was a high correlation of the summer rainfall and interannual variations of SST in the tropical Pacific Ocean. Because of limitations on data availability, we used the following method to obtain information of low-level flows in epoch 1. First, we used 1958–78 NCEP–NCAR reanalysis data and developed composites of 850-hPa geopotential and wind for those summers with above- (below-) normal rainfall in the central United States. From comparisons of the 850-hPa geopotential and the SLP′ com-

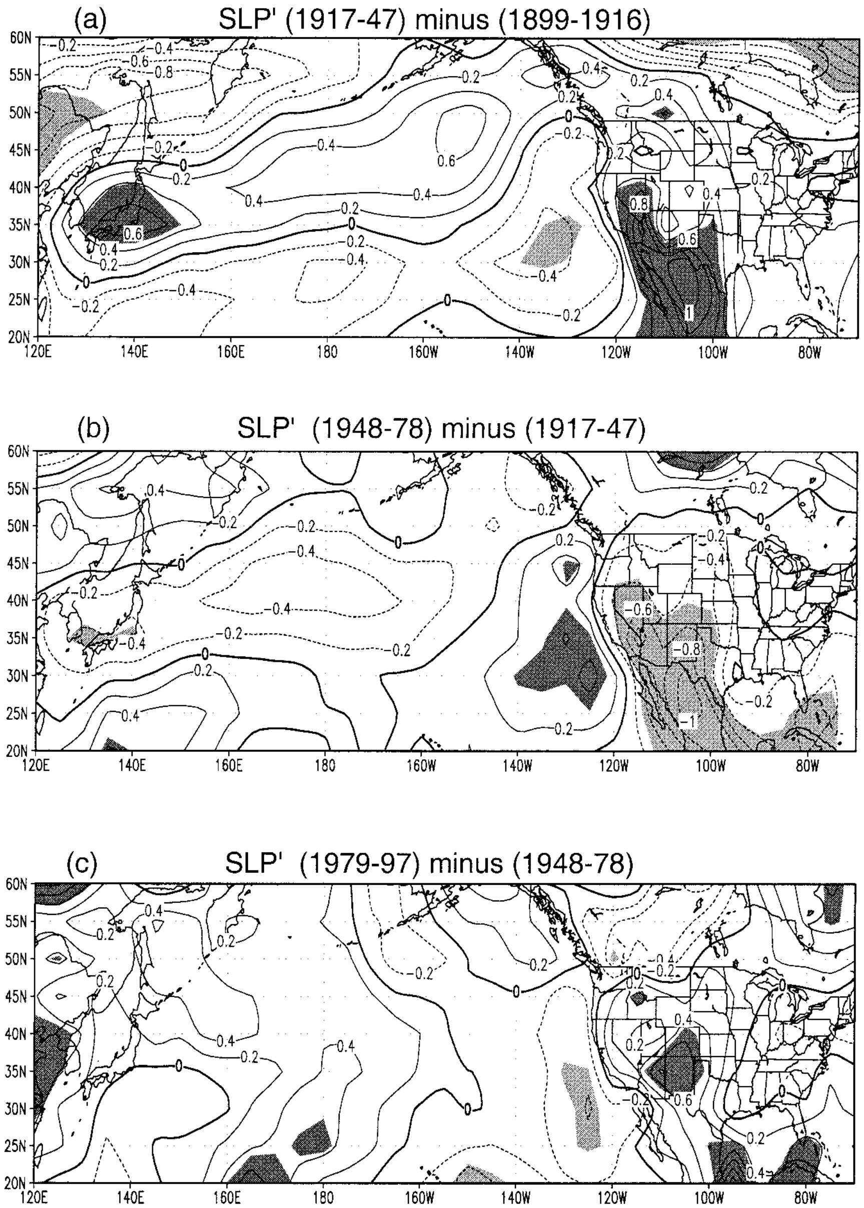

FIG. 9. Differences of summer SLP′ from (a) 1917–47 minus 1899–1916, (b) 1948–78 minus 1917–47, and (c) 1979–97 minus 1948–78. Contour interval is 0.2 hPa. The shaded areas indicate the differences between two epochs are above 95% statistical significance level, with dark shade for positive regions and light shade for negative regions.

posites for the same period we found that the patterns of the two composites were identical, a result that assured the applicability of SLP′ to approximate 850-hPa geopotential anomalies for the same epochs. We then used the SLP′ data in both epochs 1 and 3 and constructed the SLP′ composites of the wet and dry summers. Alternations of these patterns composed the interannual variations of atmospheric dynamic field and rainfall development in the central United States corresponding to interannual variations in SST in the tropical Pacific.

Figure 10a shows differences of anomalies of 850-hPa geopotential and wind between wet and dry summers in 1958–78. Positive geopotential anomalies were in the northeastern Pacific Ocean off the west coast of the United States. A separate and much weaker positive anomaly center was in the northwestern United States. Weak positive anomalies also appeared over Mexico and in the Gulf of Mexico. In between these high anomalies was a negative-anomaly region in southern Texas and northern Mexico. This distribution was consonant with that in Fig. 8c for SLP′. The higher geopotential in the northeastern Pacific Ocean and western United States in the wet summers established a trough with anoma-

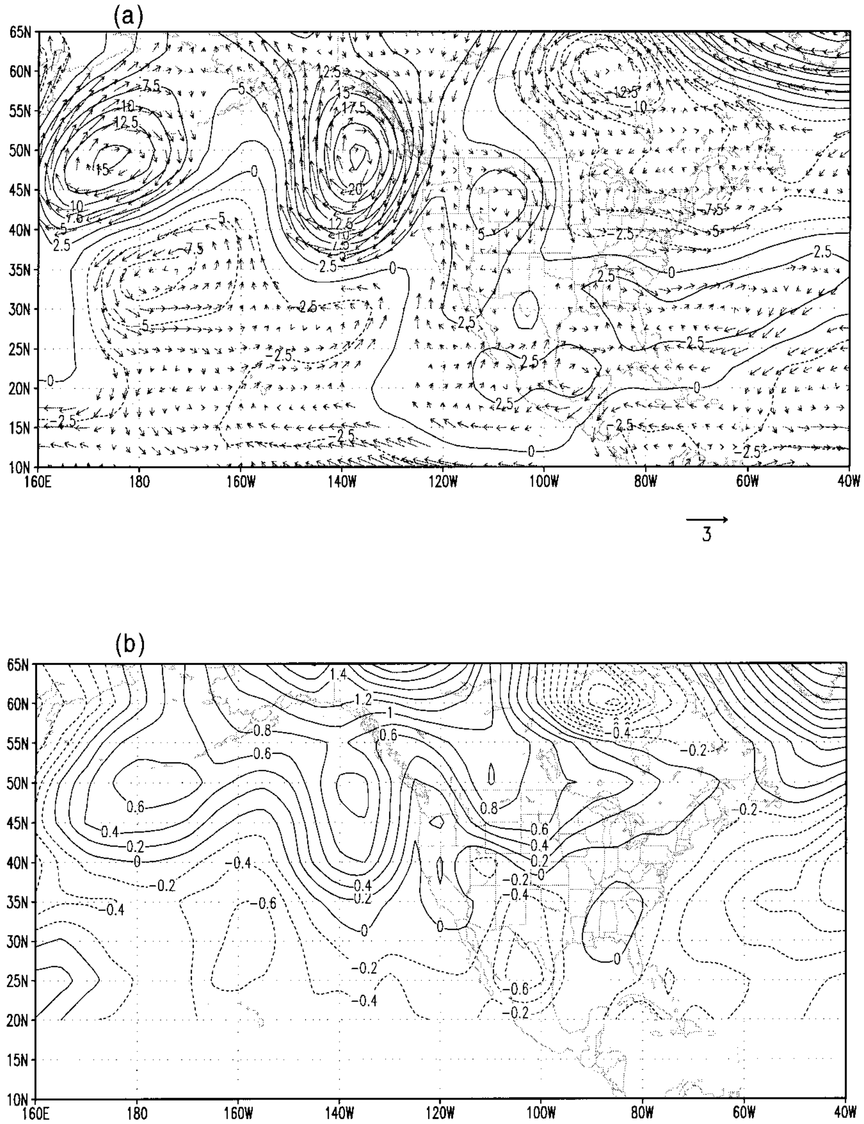

FIG. 10. (a) Summer 850-hPa geopotential height and wind differences between wet and dry years in 1958–78. Contour interval is 2.5 gpm. The wind vectors are not shown where the geopotential height difference is near zero. (b) Summer SLP′ difference between wet and dry years during 1899–1916 and 1948–78. Contour interval is 0.2 hPa.

lously strong northerly wind in the central United States. [A similar anomalous northerly wind was also noticed in 500 hPa for wet summers in the early 1950s (Ting and Wang 1997).] In the meantime, the weak high in the Gulf region directed weak southerly flows to the central United States. This flow configuration favored a stationary front from northeast to southwest across the central United States, a major feature that local forecasters look for in prediction of summer storms.

A reversal of the sign/direction of the geopotential and wind in Fig. 10a yields the anomaly fields typical in dry summers in the central United States. Specifically, the field has low geopotential (low pressures in height surfaces) in the northeastern Pacific Ocean, a weak low in the western United States, and higher pressure in the Gulf region. A ridge is configured in the central United States from these anomalies. The southwesterly wind from the dry plateaus and the desert southwest of the United States and divergence in the central United States suppressed opportunities for storm development and rainfall (Mo et al. 1997; Hu et al. 2000).

Figure 10b shows the SLP′ differences between composites of wet and dry summers for epochs 1 and 3. Except for minor differences in the southwestern United States, the pressure anomaly pattern in the North Pacific Ocean and most of the United States is similar to that in Fig. 10a. Again, this pressure field encouraged strong northerly winds and weak southerly winds to converge from north and south, respectively, to the central United States and produce above-normal summer rainfall. The reversed flow pattern results in dry summers in the region. Because these anomaly patterns and their alternations are associated with the interannual variations in the SST in the tropical Pacific and the ENSO cycle in those epochs, the variations of the summer rainfall in the central United States show significant interannual variations.

6. Summary and concluding remarks

The attributes of the interannual variations in summer rainfall in the central United States are from multiple origins, including interannual variations in the SST in the tropical Pacific Ocean associated with ENSO and the variations in the Gulf of Mexico and the eastern tropical Atlantic region. Although influences from these origins may coexist in the same period and affect the summer rainfall, they have taken turns to dominate the rainfall variations for a period of 30–40 years. Our analyses showed that the correlation of interannual variations of the summer rainfall in the central United States with the SST in the tropical Pacific was high (above 95% significance level) in the epochs 1871–1916 and 1948–78. It languished in the epochs 1917–47 and 1979–present. In the high correlation periods, the teleconnection of the SST in the tropical Pacific related to the interannual variations of the ENSO cycle explained a major portion of the observed interannual variations in summer rainfall in the central United States.

We found that the 30–40 yr alternation of the dominant forces affecting the interannual summer rainfall variations has been in a close association with a multidecadal variation of 60- to 70-yr period in the mid- and high-latitude North Pacific region. The two epochs of high ENSO teleconnection were characterized by an SSTA distribution in the North Pacific Ocean with a broad positive SSTA area in the central North Pacific surrounded by negative SSTA in eastern North Pacific. The two epochs with an absent teleconnection were characterized by a reversal in SSTA in the North Pacific. The SLP′ particularly in the eastern North Pacific showed fairly regular alternations similar to the variations in SSTA. In the high teleconnection epochs, a broad area of negative SLP′ existed in the central North Pacific, coincident with the positive SSTA there. Positive SLP′ were found in the region of negative SSTA. In the nonteleconnection epochs, this SLP′ pattern reversed. A distinct feature in these SLP′ variations between the epochs was a broad area in the eastern North Pacific Ocean off the west coast of the United States stretching south to tropical latitudes. In this area, positive SLP′ was accompanied by negative SSTA in the high teleconnection epochs and negative SLP′ accompanied by positive SSTA in the opposite epochs. The broad positive anomalies of SLP′ in this region encouraged the teleconnection of tropical SST variations to influence circulation and rainfall variations in the central United States. The mechanisms behind this relation are yet to be discovered.

The variation of the teleconnection indicates there has been no persistent effect of ENSO on the summer rainfall in the central United States. This variation attributed to the seemingly conflict results about ENSO effects on rainfall variations in the central United States (e.g., Ropelewski and Halpert 1986; Ting and Wang 1997). Clearly, it must be considered in making predictions of the summer rainfall anomalies based on ENSO development to correctly capture the role of ENSO in the interannual summer rainfall variations.

When the teleconnection was active, the central United States was at a center of a deformation field in wet summers corresponding to the warmer phase of the ENSO and was at a center of a reversed flow field in dry summers corresponding to the cold phase of ENSO. The deformation field resulted from a strong positive anomaly of geopotential in the northeastern Pacific Ocean, a positive anomaly of geopotential in the southeastern portion of the United States and the Gulf region, and weak negative geopotential anomalies in the southwestern and northeastern United States. The reversed flow pattern resulted from a reversed distribution in geopotential anomalies. Thus the interannual variation in tropical Pacific SST associated with ENSO cycle affected the summer rainfall variation in the central United States in this active phase of the teleconnection. This relationship broke down in the inactive phase of the teleconnection.

It is worthy to point out that this phase change in the teleconnection and the multidecadal variation in the North Pacific region has been in a good agreement with the multidecadal variation in the average NH surface temperature in the last 125 years. The two weak cooling periods from 1871 to the mid 1910s and from 1948 to the late 1970s and the two ‘‘abrupt warming’’ periods from 1917 to 1947 and from 1978 to the present (Ghil and Vautard 1991) coincided with the two very active teleconnection and the two nonteleconnection epochs, respectively, in the last 125 years. Whether this was a random coincident or a manifestation of a consistent global climate variation remains to be investigated.

Last, we should caution the readers that the limitation of the data length casts a reasonable doubt on the persistence of the identified multidecadal variation of the teleconnection between ENSO and interannual variations in summer rainfall in the central United States. Further investigations of this issue will be necessary when longer data records become available.

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