Occurrence of droughts and floods during the normal summer monsoons in the mid- and lower reaches of the Yangtze River

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[1] The daily precipitation data at 720 stations over China for the 1957–2000 period during summer (May–August) are used to investigate the droughts-floods coexistence (DFC) phenomenon during the normal summer monsoons. A droughts-floods coexistence index on seasonal timescale over the mid- and lower reaches of the Yangtze River (MLYRV) is defined to quantify this phenomenon and the associated ocean-atmospheric features in the strong DFC years are examined statistically. Results demonstrate that the occurrence of the strong summer DFC in the MLYRV is of an increasing trend for the period of 1957–2000. The strong summer DFC in the MLYRV is often accompanied by the anomalously subseasonal oscillation of the western Pacific subtropical high, the low-level westerly winds anomalies over the equatorial oceanic areas from the Indian Ocean to the western Pacific and the northward cross-equatorial winds anomalies near Sumatra and Somalia during summer, the strong Southern Hemisphere annual mode during the preceding November through January, high sea surface temperature in the oceanic areas from the Arabian Sea to the South China Sea, and El Niño or the developing phase of El Niño in the 6 preceding months. All these offer some predictive signals for the summer DFC in the MLYRV.
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1. Introduction

[2] Previous studies suggest that the summer rainfall in the mid- and lower reaches of the Yangtze River valley (MLYRV) is of close relationship with the advance and retreat of the East Asian Summer Monsoon (EASM) [*Tao and Chen*, 1987; *Lau et al.*, 1988; *Ding*, 1992; *Wang*, 1994; *Chang and Chen*, 1995; *Lau and Yang*, 1997]. It is also found out that the severe summer droughts or floods in the MLYRV are associated with the singularities of the large-scale atmospheric circulation [*Chen*, 1994; *Wang and Xu*, 1997; *Gong and Wang*, 1999; *He et al.*, 2001; *Nan and Li*, 2003]. The majority of the previous literatures on severe droughts and floods put their emphasis on the seasonal mean rainfall [*Simmonds et al.*, 1999; *Webster et al.*, 1998; *Barlow et al.*, 2002; *Matsumoto*, 1997], while few consider the subseasonal variation of rainfall that is also of great importance. For example, if precipitation of some summer is predicted to be normal, it might be misunderstood that there would be neither droughts nor floods in the summer. Nevertheless, if both droughts and floods happened within the summer, the seasonal precipitation could also be normal. Such a misunderstanding is due to the ignorance of the subseasonal variation of precipitation. As a result, the decision maker and the public may not be warned and prepared for severe floods and droughts in advance, which would cause disastrous loss of lives and property. The summer droughts-floods coexistence (DFC) in the MLYRV is just such an occasion of extremely large rainfall subseasonal variation, yet with a normal seasonal mean. In this paper, the DFC refers to the occurrence of both droughts and floods during the normal summer monsoons. We try to investigate the ocean-atmospheric features in the anomalous DFC summers with normal seasonal precipitation, using ERA-40 reanalysis data sets and Reynolds sea surface temperature (SST) data.

[3] According to the classification of precipitation area given by Ting and Wang [*Ting and Wang*, 1997], from the standard deviations of the summer precipitation of Chinese 720 stations for the 1957–2000 period, we select the station, Gaoyou, which is of the largest rainfall variation in the MLYRV, as the base point. Based on the one-point correlation map of the summer precipitation between the base point, Gaoyou, and Chinese 720 stations (not shown), we then use the mean precipitation at the 42 stations, which are of the same rainfall variation as that at the base point, their correlation coefficients exceeding the 95% confidence level based on the Monte Carlo, to represent the summer precipitation in the MLYRV. The summer in the MLYRV is defined as May–August (MJJA) in this paper.

[4] In order to describe the DFC phenomenon quantitatively, a DFC index (DFCI) is defined as follows:

$$DFCI = S_{STD} \cdot 10^{(0.5-|R_{STD}|)}$$

where S_{STD} and R_{STD} are, respectively, the normalized average summer no-rain days and precipitation of the whole MLYRV. S_{STD} reflects the concentration intensity of the summer rainfall. The larger the S_{STD} is, the more concentrated the summer rainfall is, and *vice versa*. $10^{(0.5\ -\ |RSTD|)}$ is the weight coefficient which magnifies the weight of the normal-precipitation summers and reduces that of the severe floods or droughts summers. It should be emphasized that those indices over 0.5 standard deviations and under −0.5 standard deviations are defined as the high DFCI (or strong DFC) and low DFCI (or weak DFC) summers, respectively. The absolute magnitude of the summer precipitation anomalies within 0.5 standard devia-

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Table 1. Normalized Precipitation and No-Rain Days of the High and Low DFCI Summers in the MLYRV From 1957 to 2000[a]

High DFCI Summers				Low DFCI Summers			
Year	DFCI	$\Delta R/\sigma_R$	$\Delta S/\sigma_S$	Year	DFCI	$\Delta R \bullet \sigma_R$	$\Delta S \bullet \sigma_S$
2000	2.70	−0.02	0.91	1977	−3.21	0.05	−1.23
1986	2.27	0.08	0.88	1975	−1.50	0.29	−1.03
1995	1.83	0.01	0.60	1993	−1.49	0.51	−1.70
1990	1.30	−0.59	1.58	1970	−1.44	0.15	−0.72
1997	0.94	−0.72	1.50	1965	−1.31	0.16	−0.67
1988	0.84	−0.44	0.69	1973	−1.04	−0.38	−0.90
1981	0.73	−0.93	1.84	1957	−0.85	0.38	−0.75
1971	0.67	−0.28	0.37	1974	−0.71	0.41	−0.68
				1963	−0.69	0.8	−1.63
				1989	−0.66	0.1	−0.31

[a] ΔR and ΔS refer to the summer precipitation anomaly and no-rain days anomaly, respectively, and σ_R and σ_S are the corresponding standard deviations.

tions is regarded as normal, while the anomalies over 0.5 standard deviations and under −0.5 standard deviations are defined as floods and droughts, respectively.

2. Results

2.1. Precipitation Distribution Features of the Summer DFC in the MLYRV

[5] To verify whether the DFCI is able to describe the occurrence of droughts and floods during the normal summer monsoons in the MLYRV, first we calculate the normalized precipitation and no-rain days of the high and low DFCI summers from 1957 to 2000 in the region (Table 1). It can be seen from Table 1 that the absolute value of the precipitation anomaly in 5 summers of all the 8 high DFCI summers is within 0.5 standard deviations, which indicates the precipitation of these 5 summers is normal, and there are 4 among these 5 summers having anomalously more no-rain days (no-rain days anomaly above 0.5 standard deviations). Although the precipitation anomalies of the other 3 summers are less than −0.5 standard deviations, their corresponding no-rain days anomalies are more than 1.5 standard deviations, much larger than the absolute value of the precipitation anomaly. All these indicate that the precipitation in the high DFCI summer is more concentrated than normal, which implies both droughts and floods occur. Among the 10 low DFCI summers, 8 summers have the normal precipitation and 9 summers have more raining days (no-rain days anomaly less than −0.5 standard deviations), which indicates the precipitation in the low DFCI summer is more uniform. Therefore, high DFCI summers often see both droughts and floods occur in spite of the normal seasonal mean precipitation, and *vice versa*. Figures 1a and 1b are the summer climatological pattern and the high DFCI summers composite pattern of Zeng-Li monsoon indices [*Li and Zeng*, 2002]. There is almost no difference between these two patterns, which illustrates that in spite of the occurrence of both droughts and floods, the high DFCI summers are of the same monsoon distribution as the summer climatological pattern. In other words, the high DFCI summers have the normal monsoons on seasonal timescale. The low DFCI summer have, too (not shown). Thus, DFCI might be regarded as a quantitative index describing the DFC phenomena during the normal summer monsoons in the MLYRV. In this paper, we mainly investigate the high DFCI (or strong DFC) summers situations.

[6] Figure 2 describes the temporal variation of the summer DFCI in the MLYRV from 1957 to 2000. The DFCI exhibits significant variability not only on interannual timescale, but also on decadal timescale. The polynomial regression trend line illustrates that the summer DFCI has an increasing trend from 1957 to 2000, relatively low before 1980s and high after 1980s. It implies that the high DFCI summers have emerged more frequently in the MLYRV since 1980s, which is consistent with the result from Table 1 that 7 of all the 8 high DFCI summers are after 1980s and 8 of all the 10 low ones are before 1980s. It suggests that summer precipitation in the MLYRV is of larger subseasonal variation after 1980 than before 1980.

2.2. Simultaneous and Preceding Atmospheric Circulation Features of the Strong DFC Summers

[7] Figure 3 displays the features of 500 hPa winds of the high DFCI summers. In May (Figure 3a), the western Pacific subtropical high (WPSH) is centered at 20°N, 155°E, extending westward to the South China Sea (SCS); and its north and west sides covered with significant anomaly areas over the western Pacific between 5°–30°N. Another significant anomaly areas are over the mid- and high latitude region of the East Asia. These features imply the WPSH is more southward than the climatological position and the intensity of the cold high over the East Asian continent is strong. In June the WPSH has a northward shift (Figure 3b), with its center at 25°N, 157°E and the significant anomaly areas mainly at its north side. A large area of significant anomaly emerges in the East Asian

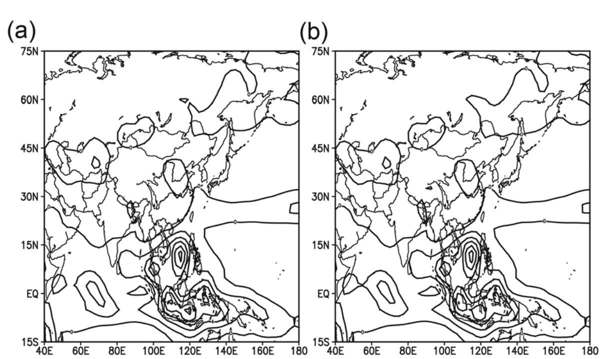

Figure 1. (a) Summer climatological pattern and (b) high DFCI summers composite pattern of Zeng-Li monsoon indices.

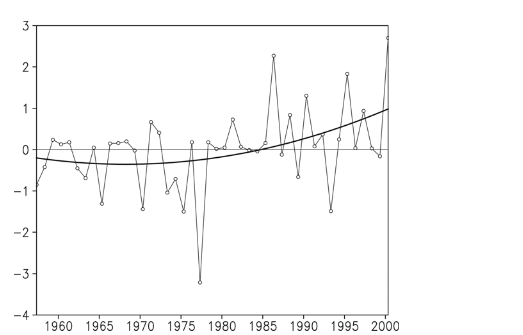

Figure 2. The normalized summer DFCI time series (the thin line with open circles) in the MLYRV from 1957 to 2000 and its polynomial regression trend (the bold line).

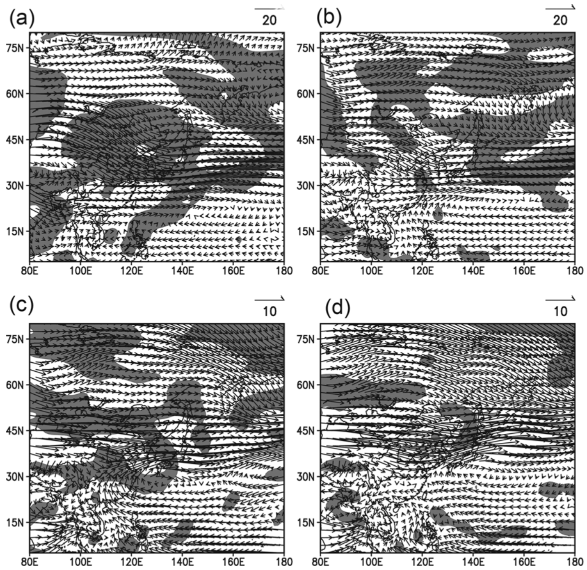

Figure 3. The composite 500 hPa winds fields for (a) May, (b) June, (c) July, and (d) August of the high DFCI summers in the MLYRV. The winds anomalies (high DFCI summers minus summer climatological mean) in the shaded areas exceed the 95% confidence level based on the Student-*t* test.

continent near 30°N in July (Figure 3c) and the WPSH has a remarkable northwestward shift, its center at 30°N, 135°E. In August the WPSH withdraws eastward (Figure 3d), with its center at 30°N, 145°E, while the significant anomaly emerging in the western Pacific. According to the above analysis, the WPSH exhibits a wide range of spacial activities in the strong DFC summers in the MLYRV, which implies that the anomalously subseasonal oscillation of the WPSH might be one factor contributing to the strong DFC in the MLYRV. Figure 4a illustrates that there is a large area of positve correlation near the equator. It indicates that in the strong DFC summers, low-level westerly winds anomalies often appear over the near-equatorial oceanic areas from Indian Ocean to the western Pacific, which offers necessary condition for the northward propagating of Rossby wave from the Southern Hemisphere (SH) to the Northern Hemisphere (NH) [*Tomas and Webster*, 1994]. There are two regions with anomalously positive correlation: one near Sumatra and the other one near Somalia (Figure 4b). It demonstrates that the strong DFC summers are usually corresponding to southerly winds anomalies in these two regions which reinforce the cross-equatorial flows from the SH [*Tao and Chen*, 1987].

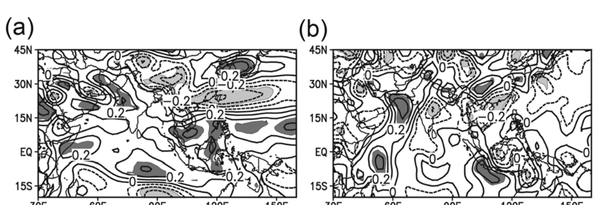

Figure 4. Correlation coefficients between the summer DFCI and the summer 850hPa (a) u and (b) v components of winds fields. The shaded areas exceed the 95% confidence level based on the Monte Carlo test.

[8] In order to investigate the relationship between the summer DFC in the MLYRV and the previous large-scale atmospheric circulation, the correlation coefficients between the DFCI and the Southern Hemisphere annual mode (SAM) index [*Nan and Li*, 2003], the Northern Hemisphere annual mode (NAM) index [*Li and Wang*, 2003a], the North Pacific Oscillation (NPO) index [*Wallace and Gutzler*, 1981] and the North Atlantic Oscillation (NAO) index [*Li and Wang*, 2003b] in the 6 preceding months (from the preceding November through April) are calculated respectively. The SAM in the preceding November, December and January (NDJ) are of the most significant relationship with the summer DFC in the MLYRV. Their correlation coefficients reach 0.35, 0.42 and 0.38, respectively, which exceed the 95% confidence level based on the Monte Carlo test. The correlation pattern between the summer DFCI and the composite SH sea level pressure of the preceding NDJ shows that the polar cap is covered by the significant negative correlation values and the mid-latitude is the significant positive correlation region (not shown), which means the higher the summer DFCI is, the larger the pressure gradient between the mid- and high latitude regions in the SH is and the stronger the SAM is. Therefore, the preceding NDJ SAM might be regarded as a predictor for the following summer DFC in the MLYRV.

2.3. Preceding and Simultaneous SST Features of the Strong DFC Summers

[9] The correlation coefficient map between the summer DFCI in the MLYRV and SST of the six preceding months (Figure 5) illustrates that from the preceding November through April, anomalously positive correlation prevails over the oceanic areas including the western Pacific, the SCS, the Bay of Bengal, and the Arabian Sea, which indicates that the high DFCI summers in the MLYRV are usually accompanied by the preceding high SST in the above oceanic areas. In addition, we noticed that in the preceding November (Figure 5a), there are significant positive correlation zones in the equatorial central Pacific and the North Pacific, propagating eastward with time (not shown). Subsequently significant positive correlations appear in the oceanic areas from the West Coast of North

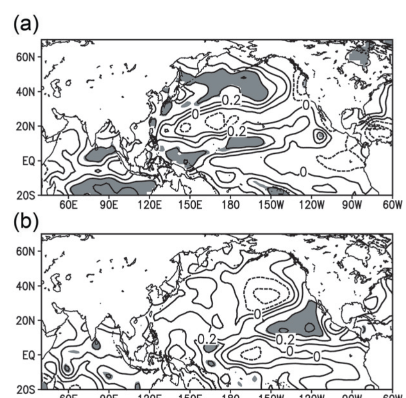

Figure 5. The correlation coefficients between the DFCI in summer in the MLYRV and SST in the preceding (a) November and (b) April. Contour interval is 0.1. The shaded areas exceed the 95% confidence level based on the Monte Carlo test.

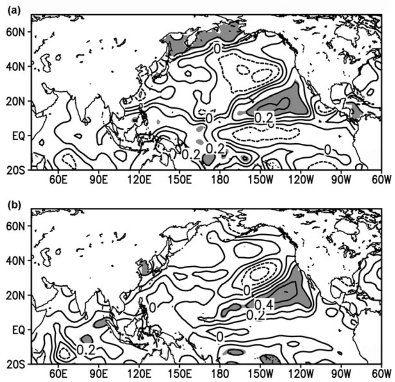

Figure 6. The correlation coefficients between the DFCI in summer in the MLYRV and SST in (a) May and (b) August. Contour interval is 0.1. The shaded areas exceed the 95% confidence level based on the Monte Carlo test.

America to the equatorial eastern Pacific and negative correlation prevails in the North Pacific in April (Figure 5b). This correlation pattern implies the SST in the equatorial eastern Pacific has a tendency to increase in the preceding months of the strong DFC summers in the MLYRV. During the contemporary period, anomalously positive correlation prevails along the West Coast of North America and propagates from low latitude to high latitude from May through August (Figure 6), which indicates that the strong DFC summers are often related to the high SST propagating from the tropical eastern Pacific to extra-tropical eastern Pacific along the West Coast of North America continent. In fact, among the 8 high summer DFCI years in the MLYRV, 6 are El Niño years. Hence, the strong DFC summers in the MLYRV seem to be related to El Niño or the developing phase of El Niño.

3. Summary and Discussion

[10] This paper focuses on the simultaneous and preceding ocean-atmospheric features of the strong summer DFC in the MLYRV. The results show that the occurrence of strong summer DFC displays an increasing trend from 1957 to 2000 and the anomalous subseasonal oscillation of the WPSH might be one factor contributing to the strong summer DFC in the MLYRV. The anomalously low-level westerly winds anomalies prevailing over the near-equatorial Indian Ocean and the western Pacific offer the necessary condition for the cross-equatorial propagation of Rossby wave from the SH to the NH. Strong SAM during the preceding NDJ is often followed by the strong summer DFC in the MLYRV. In the preceding NDJ of the strong DFC summers in the MLYRV, high SST in the oceanic areas from the Arabian Sea to the SCS and El Niño or the developing phase of El Niño are often observed. All these phenomena offer some predictive signals for the strong summer DFC in the MLYRV. Nevertheless, the physical mechanism of the interannual variability of summer DFC is still not clear. Based on the results thus far, the cross-equatorial propagating of the preceding strong SAM signal from the SH to the NH, associated with the anomalous subseasonal oscillation of the WPSH, might also be a factor contributing to the strong summer DFC in the MLYRV. High SST developing in the tropical eastern Pacific might also play a role in the strong summer DFC in the MLYRV. This paper examines the summer DFC in the MLYRV on seasonal timescale. There is no doubt that if the timescale gets shorter, the possible mechanisms related to the DFC might be different. These occasions need further investigations and will be discussed in other papers.

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