East-west SST contrast over the tropical oceans and the post El Niño western North Pacific summer monsoon

Toru Terao
Faculty of Informatics, Osaka-Gakuin University, Osaka, Japan

Takuji Kubota
Graduate School of Engineering, Osaka Prefecture University, Osaka, Japan

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[1] The circulation and SST anomalies during the post El Niño Asian summer monsoon season were examined through data analysis and linear equatorial β plane model calculations. Over the Philippine Sea, a negative precipitation anomaly and low-level anti-cyclonic anomaly were found. Over the western equatorial Pacific, a low-level easterly anomaly prevailed. An east-west SST anomaly contrast dominated over the tropical Indian and Pacific Oceans, with positive anomalies over the Indian and western Pacific Oceans and negative anomalies over the central to eastern Pacific. The β plane model demonstrated that the anti-cyclonic anomaly over the Philippine Sea was a Rossby response to the negative precipitation anomaly found in this region. The easterly anomaly along the equator was part of a Kelvin response to the SST anomaly contrast. On the northern side of this anomalous easterly, a negative vorticity anomaly developed. This could induce the moist Rossby wave over the Philippine Sea.

[2] In the present study, an investigation is conducted on the features of the SST and atmospheric circulation anomalies that are significant in the post El Niño summer (June, July, and August), using data from 1980 to 1998, which includes three major El Niño events (1982/83, 1987/88, 1997/98). As shown in Figure 1, these El Niño events that matured during the boreal winter were accompanied by the succeeding rapid SST cooling in the eastern Pacific. Therefore, in the present study, the post El Niño years are defined as 1983, 1988, and 1998.

[3] In order to investigate the variations in the circulation fields, the NCEP/NCAR reanalysis data [Kalnay et al., 1996] were utilized. For the sea surface temperatures, Global Sea-Ice and SST dataset (GISST) compiled by the UK Met Office were used. Use was also made of the CPC Merged Analysis of Precipitation (CMAP) for the monthly precipitation dataset [Xie and Arkin, 1997].

1. Introduction

[4] The relationship between the El Niño/Southern Oscillation (ENSO) and the precipitation associated with the Asian summer monsoon has been of major interest to researchers. During the 1997/98 winter, an unprecedentedly amplified El Niño event followed by a quick transition to the La Niña that matured in the 1998/99 winter was observed. Lau and Wu [2001] pointed out that the Asian summer monsoon during this transition was characterized by numerous disasters resulting from the heavy precipitation. Lau and Wu [2001] also isolated a coupling mode between the pattern of the Asian summer monsoon precipitation and SST variations, which became dominant, especially in the summer season, during the rapid cooling of the eastern equatorial Pacific after the mature phase of El Niño (hereafter post El Niño summer) such as in 1998. This mode consists of heavy rainfall over some regions in East and South Asia, and suppressed rainfall and anomalous anti-cyclonic flow over the Philippine Sea.

[5] Many studies have pointed out that such a low-level anti-cyclonic anomaly over the Philippine Sea is a significant signal for anomalous precipitation over East Asia in the post El Niño summer [Wang et al., 2001; Lu and Dong, 2001]. There are discrepancies, however, among the discussions on the formation mechanism of this anti-cyclonic anomaly. Wang et al. [2001] proposed that the local negative SST anomaly in the western North Pacific can play a crucial role in maintaining this anomalous low-level anti-cyclone, based on the earlier study by Wang et al. [2000]. This SST anomaly is evident in the boreal winter and spring before the post El Niño summer. In summer season, however, it is not significant [Lu, 2001]. The nature of the post El Niño western North Pacific summer monsoon circulation should be further addressed.

[6] Composited anomalies for the post El Niño summers are shown in Figure 2. In the composited precipitation anomaly map (Figure 2a), it is found that significant negative precipitation anomalies prevail over the Philippine Sea. Over the eastern equatorial Indian Ocean, a large positive precipitation anomaly is found. On the other hand, in several small areas near Japan, the southern part of China, and the west coast of India, above normal precipitation areas appear.

[7] Over the Indian and Pacific Oceans, an east-west SST anomaly contrast develops (Figure 2b). To the west of 140°E, positive SST anomalies prevail in contrast with the central to eastern Pacific Ocean, where negative SST anomalies develop. This is consistent with the fact pointed out in past studies [Chiu and Newell, 1983; Kubota and Terao, 2003], that SST anomalies in the Indian and western Oceans, with positive anomalies over the Indian and Pacific Oceans and negative anomalies over the tropical Indian and Pacific were examined. The nature of the post El Niño western North Pacific summer monsoon circulation should be further addressed.
Pacific Oceans tend to be positive in the summer after the maturing phase of El Niño. [8] In the lower troposphere over the Pacific Ocean, two zonally elongated subtropical anomalous highs are seen in the Northern and Southern Hemispheres (Figure 2c). Between these highs, a strong easterly anomaly prevails along the equator (Figure 2c). The wind speed of the anomalous easterly averaged over 120°–150°E and 10°S–10°N is 1.7 ms⁻¹. In the subtropical regions, the wind velocities appear to be in geostrophic balance.

[9] On the equator, the geopotential anomaly is the largest (more than 4 m) at about 150°W, and tends to decrease gradually to the west (Figure 2c). This eastward pressure gradient along the equator corresponds to the strong easterly wind anomaly dominating the equatorial area between 120°E and 150°W.

[10] Along 15°–20°N (Figure 2c), the anomalous ridge extends westward to 90°E, which corresponds to the low-level anti-cyclonic circulation anomaly pointed out in earlier studies [Wang et al., 2001; Lau and Wu, 2001]. The location of this anti-cyclonic anomaly center is slightly to the northwest of the center of the negative precipitation anomaly over the Philippine Sea shown in Figure 2a. This suggests that the anti-cyclonic anomaly is a Rossby response to the negative diabatic heating of the precipitation anomaly.

[11] In Figure 2b, the negative SST anomaly is not significant over the Philippine Sea, which supports the comment by Lu [2001], noting that a significant SST signal is not found over the Philippine Sea.

[12] As was shown by Xie et al. [2003], the above normal SST over the South China Sea in this season can be attributed to the anti-cyclonic circulation prevailing above. Thus, the distribution of SST anomaly and the circulation anomaly that were pointed out above is self-consistent over this area.

[13] For the post La Niña summers, no significant anomalies with opposite signs are found (figures not shown). The main difference is in the temporal evolution of the SST over the central to eastern Pacific. While the SST anomaly decreases rapidly in time toward the summer season in the post El Niño years, the SST anomaly does not change in the post La Niña spring.

3. Equatorial θ Plane Model

[14] To evaluate the first baroclinic stationary response of atmospheric circulations to thermal forcing, a simple
linear equatorial $\beta$ plane model [Matsuno, 1966; Gill, 1980] is employed. The two spatial patterns of forcing functions are shown in Figure 3. The $Q_a$ forcing function is designed to model the tropical large scale east-west SST contrast with a node at 140°E (Figure 2a). On the other hand, $Q_b$ models the anomalous cooling caused by the suppression of the convective activity over the Philippine Sea (Figure 2b).

[15] For $Q_a$, the equivalent depth is set to 20 m, corresponding to a diabatic heating of a 4500 m vertical scale. On the other hand, the equivalent depth for $Q_b$ is set to 100 m, reflecting the vertical scale (~10,000 m) of deep convection. Here, the vertical scale of anomalous sensible heating responsible for the SST anomaly is estimated with consideration of the vertical scale of the lower tropospheric temperature anomalies associated with the SST anomalies. These anomalies are reported to be confined within 3 km above sea level [Kubota and Terao, 2003]. It is reasonable to assume that the equivalent depth for $Q_a$ should be much smaller than that for $Q_b$. Both the Newtonian cooling and Rayleigh friction are assigned as $(5 \text{ days})^{-1}$. The buoyancy frequency is assumed to be $1.0 \times 10^{-2} \text{s}^{-1}$ over the model domain. The results shown below are not to any extent altered if somewhat different parameters are used.

[16] Figure 4a shows the circulation anomaly pattern that should arise from the east-west thermal forcing contrast $Q_a$. Over the cooling area in the subtropics in both hemispheres, a positive geopotential anomaly with a trough along the equator prevails. On the equator, an eastward geopotential gradient can be seen over the area shown. A strong anomalous easterly wind prevails along the equator, originating from the cooling area, and accelerating until it arrives at the edge of heating area. It is clear that this result captures the large scale characteristics of the circulation anomaly pattern shown in Figure 2c. The magnitude of the heating rate that exerts the observed wind speed in the lower troposphere ($-1.7 \text{ ms}^{-1}$ averaged over 120–150°E and 10°S–10°N) is about 0.2 K day$^{-1}$.

[17] On the other hand, the local cooling off the equator $Q_b$ induces an anti-cyclonic circulation anomaly (Figure 4b), which is shifted somewhat to the northwest of the forcing. The circulation anomaly, extending approximately 30° to the west of the forcing area, is recognized as a Rossby response to the localized cooling. The resemblance to the observed precipitation and geopotential anomaly patterns over southeast Asia and the western North Pacific (Figures 2a and 2c) is obvious. The heating rate responsible for the observed wind speed ($-1.6 \text{ ms}^{-1}$ averaged over 120–150°E and 10–20°N)
is, in this case, about 1 K day$^{-1}$, which corresponds to the suppression of diabatic heating due to the decrease in the precipitation rate of about 3 mm day$^{-1}$. Figure 2a shows that the observed precipitation anomaly over the Philippine Sea is $-2.6$ mm day$^{-1}$, indicating that the observed precipitation anomaly can induce the anti-cyclonic anomaly that prevails over the Philippine Sea with the observed amplitude.

[18] Thus, it is confirmed that the circulation anomaly pattern that appears in the post El Niño summer (Figure 2) is understood to be a mixture of the Kelvin response to the large scale east-west SST contrast, and the Rossby response to the zonally elongated negative precipitation anomaly over the Philippine Sea.

[19] There remains, however, the question of how does the persistent precipitation anomaly over the Philippine Sea exist without any geographically fixed forcing. As was shown in the previous section, no local SST anomaly could be found that coincides with the suppressed precipitation over the Philippine Sea. Another geographically fixed forcing should exist that is responsible for the suppressed precipitation.

[20] In order to determine this, the model results should be revisited (Figure 4a) for east-west heating contrast $Q_a$ (Figure 3a). On the northern side of the low-level easterly wind prevailing on the equator, a zone of negative vorticity can be seen as a part of the thermally forced persistent Kelvin response as shown in Figure 4c.

[21] Such negative vorticity may induce a negative vertical mass flux anomaly from the lower boundary layer through a reduction in the Ekman pumping. Under such a situation, convection is suppressed if the convective activity is climatologically strong, which in turn exerts a Rossby response. Thus, it is suggested that the lower tropospheric negative vorticity anomaly over the western North Pacific, exerted by the east-west SST anomaly contrast, is a plausible cause of the anti-cyclonic anomaly over the Philippine Sea. The result of Wang and Xie [1997] is supportive to this mechanism, where the moist Rossby wave was shown to become unstable in this area due to the vertical easterly shear and boundary effect.

[22] The reason why such a Rossby mode does not appear on the southern side of the Kelvin mode should be noted here. First, in the Southern Hemisphere, the moist Rossby wave cannot develop, since the humidity is less than that of the Northern Hemisphere [Wang and Xie, 1997]. Second, it should be remarked that this region is climatologically dominated by weak southerly winds during this season (1.3 ms$^{-1}$ averaged over 120°–150°E and 20°S–0°). This flow carries the anomalous anti-cyclonic vorticity produced by the Kelvin mode from the subtropics in the Southern Hemisphere toward the equator.

4. Summary

[23] El Niño events that mature in the boreal winter tend to accompany the succeeding rapid cooling of the SST over the equatorial eastern Pacific. The impact of this rapid SST transition on the circulation in the summer monsoon season was investigated using NCEP/NCAR reanalysis data from 1980 to 1998, and a simple equatorial $\beta$ model.

[24] In the summer monsoon season during the rapid cooling of the eastern equatorial Pacific after the El Niño winter (post El Niño summer), an east-west SST anomaly contrast prevails over the Indian and Pacific Oceans. Above normal SSTs were found over the Indian-western Pacific Oceans and below normal over the central to eastern Pacific Ocean (Figure 2b). Corresponding to this, a strong easterly anomaly along the equator developed from the western North Pacific to the maritime continent, as a Kelvin response to the east-west SST anomaly contrast. Such anomaly patterns of SST and circulation become fully evident after June. In spring, the SST contrast already exists but is less significant. While the Kelvin response to this SST anomaly is evident, the Rossby response is not established yet (figures not shown). It should also be noted here that similar circulation anomalies appear for other post El Niño summers prior to 1980.

[25] Over the western North Pacific, the reduction of convection over the Philippine Sea and a lower tropospheric anti-cyclonic circulation anomaly prevailed. Wang et al. [2001] proposed a local mechanism due to the positive feedback process between the negative SST anomaly and the positive sea level pressure anomaly [Wang et al., 2000]. As has been shown in Figure 2b, however, no significant local negative SST anomaly could be found.

[26] Instead, another plausible mechanism was proposed that accounts for the formation of the anti-cyclonic anomaly over the Philippine Sea. In this process, not the local SST anomaly, but the large scale zonal SST anomaly contrast was important. This exerted a Kelvin response with a strong easterly flow from the central Pacific to the Maritime Continent along the equator, with an anomalous negative vorticity on its northern side. This in turn, through the suppression of convection due to the reduction of Ekman pumping, exerted the anti-cyclonic anomaly over the Philippine Sea as a Rossby response.

[27] Thus, the anti-cyclonic anomaly over the Philippine Sea occurs concurrently with the east-west SST anomaly contrast over the Indian and Pacific Oceans. To further clarify the mechanism of the anti-cyclone as a delayed response to El Niño, the temporal evolution of global SST anomaly associated with El Niño should be addressed more. Wang [2002] addressed the importance of the Walker circulation in the temporal evolution of global SST anomaly. The Kelvin response to SST analyzed in the present study, which is another aspect of the Walker circulation, may give a hint to the evolution of ENSO.

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References


T. Kubota, Graduate School of Engineering, Osaka Prefecture University, 1-1, Gakuen-cho, Sakai, Osaka 599-8531, Japan. (kubota@aero.osakafu-u.ac.jp)

T. Terao, Faculty of Informatics, Osaka-Gakuin University, 2-36-1, Kishibe-Minami, Suita, Osaka 564-8511, Japan. (terao@s.osaka-gu.ac.jp)