Sea Surface Temperature in the South China Sea —an Index for the Asian Monsoon and ENSO System—

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Abstract

Interannual variability of the sea surface temperature anomalies (SSTA) over the South China Sea (SCS) is recognized as an index for the Asian monsoon and ENSO system because of its special geographical location for that system.

The following results are obtained by the statistical analysis of the observational data.

- (1) In the northern winter, the SCS SSTA are quite sensitive to the longitudinal shift of global wind anomalies associated with the equatorial Pacific SSTA. This fact is related to that the SCS SSTA and the neighbor SSTA have strong biennial oscillation.
- (2) When the global wind anomalies are shifted eastward in the winter (BO-type years), the tropical eastern Pacific SSTA tend to change in the following spring. On the other hand, when those wind anomalies are shifted westward (LF-type years), the eastern Pacific SSTA tend to be maintained through the year. The associated differences between the BO and LF-type years are found in the seasonal change of the low-level tropical wind anomalies from the preceding summer through winter.
- (3) The northern summer SCS SSTA seem to be controlled by *in-situ* low-level wind anomalies. Furthermore, easterly anomalies over South Asia and the tropical western Pacific and westerly anomalies over East Asia are found in the lower atmosphere for the positive SCS SSTA. It is also shown that the summer SCS SSTA have a statistical relationship with the equatorial central Pacific SSTA in the preceding winter. This fact suggests a relationship between the summer Asian monsoon and the winter phase of ENSO.

1. Introduction

The South China Sea (SCS) sea surface temperature (SST) from boreal winter through summer has remarkable characteristics associated with the Asian monsoon; large variability of the biennial oscillation (Shen and Lau, 1995), significant time-lagged correlation with summer Indian rainfall (Tomita and Yasunari, 1996), and a close relationship with Chinese rainfall (Shen and Lau, 1995).

1.1 SCS and biennial oscillation

Chen *et al.* (1992) found in long-term observational data that Chinese summer rainfall and surface temperature show a clear signal of the biennial oscillation. According to Shen and Lau (1995), the summer SCS SST anomalies (SSTA) show high correlation with the Chinese summer rainfall. The SCS is the area where the SSTA have the maximum spectral densities in the 2–3 years periods (the biennial oscillation) together with the Timor Sea (Fig. 6 of Shen and Lau, 1995). Kawamura (1988) also showed that the biennial oscillation is pronounced

Fig. 1. Power spectral densities of the SCS, and the Niño4 SSTA in DJF during 1963–94. The SSTA are first normalized with respect to their standard deviations.

in the SCS SSTA and it has a peak in the cold season. On the other hand, the 3–6 years periods (the low-frequency oscillation) are dominant in the tropical central and eastern Pacific. Figure 1 shows the power spectral densities of the SCS SSTA and the Niño4 SSTA for the December–January–February (DJF) average (the June–July–August is also referred to as JJA hereafter), which is obtained from 32 years of data during 1963–94. Note that the average from December in 1962 through February in 1963 is denoted as the DJF average in 1963 in this paper. The shaded area in Fig. 2 is referred to as the SCS hereafter, and the Niño4 areas are defined in this paper as the regions of (160°E – 150°W , 2°S – 2°N). Their interannual variabilities are normalized by their standard deviations. The power spectral densities of the DJF SCS SST are concentrated within 2–3 year. On the other hand, the Niño4 SST have a large peak at 4–6 year.

Figures 3a and 3b shows year-to-year variations of the SCS SSTA and the Niño4 SSTA in DJF for the two separate periods of 1980–94 and 1965–79, respectively. Those are normalized by their standard deviations. (hereafter, all graphs for interannual variability in this paper are shown after variables are normalized by their standard deviations). From the figure, we can see how the biennial and low-frequency oscillations are related to each other specifically. The polarities of the SCS SSTA are the same as those of the Niño4 SSTA for about two thirds of all years. For the other one third years, the polarities of the SCS SSTA and the Niño4 SSTA are inverse. The latter years are represented with the character ‘L’ in the lower part of the figure. For most of those years, the SCS SSTA change their

polarities toward the next winter while the Niño4 SSTA tend to keep their polarities.

The low-frequency oscillation in the eastern Pacific SSTA and the associated atmospheric phenomena has been studied theoretically. According to the delayed oscillator mechanism (Saurez and Schopf, 1988; Battisti and Hirst, 1989), the coupled dynamics between the ocean and the atmosphere in the Pacific is responsible for the El Niño phenomena with the low-frequency oscillation. Jin *et al.* (1994) showed that the Pacific ocean-atmosphere coupled system is able to produce the biennial oscillation as well as the low-frequency oscillation when the annual cycle is taken into account. Meehl (1987, 1993) proposed a biennial mechanism that local ocean heat content is intensified as a long-term memory mechanism after a season of local convective maximum, and that annual movement of the active convection region explains the large-scale structure in the biennial oscillation.

Tomita and Yasunari (1996) focus on the SCS SSTA to propose a possible seasonal process for the biennial oscillation of the El Niño and Southern Oscillation (ENSO)/monsoon system. In that proposed process, the SCS SSTA pass the information of the equatorial central Pacific SSTA in winter to the following summer Indian monsoon. The role of the SCS SSTA is supported by the fact that the SCS SSTA tend to be persistent from winter through to the next summer.

We are interested in the large-scale background responsible for the significant biennial oscillation and the persistency of the SCS SSTA. The persistency of the SCS SSTA is not the case for all years. For about one third of the years, the persistency from winter through the following summer is broken. Those years completely coincide with the ‘L’ years denoted in Fig. 3. We can see this fact by comparing the DJF SCS SSTA in Fig. 3 and the JJA SCS SSTA, which will be shown later in Fig. 8. A question is — what are the differences in the atmospheric circulations between the persistent SCS SSTA years and the non-persistent SCS SSTA years?

The biennial oscillation is not confined to the SCS, but is a global phenomenon. Yasunari (1991) and Yasunari and Seki (1992) discussed the role of the Asian monsoon on the biennial nature of the global climate system. Meehl (1994) studied the coupled land-ocean-atmosphere processes associated with the tropical biennial oscillation, and concluded that the northern winter circulation over the Eurasian Continent is altered by the tropical convective heating anomalies over the biennially oscillating tropical SSTA.

Yasunari (1985) found two types of eastward-propagating modes in the tropics, the SO mode (40–60 month periods) and the QBO mode (20–30 month periods), which have similar zonal struc-

Fig. 2. Climatic SST in JJA. The SCS region is shaded.

tures but different phase speeds. Rasmusson *et al.* (1990) concluded that the ENSO-related variance is concentrated in 24–30 month (biennial) and 48–60 month (low-frequency) bands. Their singular spectrum analysis indicates that those two time scales are associated with two distinct modes of the variability. Barnett (1991) obtained the two complex eigenmodes with the quasi-biennial and low-frequency periods. In the above studies, the two modes are treated separately, then the interface between the two modes is discussed. Yasunari (1985) suggested that a coupling (decoupling) of the two modes is associated with the major (minor) ENSO events over the eastern Pacific. Rasmusson *et al.* (1990) and Barnett (1991) also obtained similar results.

The variance of the biennial oscillation is large in the eastern equatorial Pacific SSTA rather than the SCS SSTA (Kawamura, 1988; Shen and Lau, 1995; Tomita and Yasunari, 1996). The biennial component in the eastern equatorial Pacific SSTA has its maximum from autumn to winter. This season for the maximum is slightly earlier than that of the SCS SSTA. The biennial component coexists with stronger low-frequency components in the DJF eastern Pacific SSTA. The DJF SCS SSTA shows a single peak of the biennial oscillation in its power spectrum (Fig. 1).

We will see the in-phase and out-of-phase fields of the biennial mode and the low-frequency mode. The biennial mode and the low-frequency mode are assumed to be represented by the polarities of the DJF SCS SSTA and the DJF Niño3 SSTA, respectively, in this paper. We will present the real fields without decomposition into the two modes.

1.2 SCS and change of Niño3 SSTA

Since the Niño3 SSTA evolution in spring is sometimes difficult to forecast, this season is called 'predictability barrier' (Webster and Yang, 1992). Except for this season, the Niño3 SSTA can be predicted with a certain skill (Latif and Graham, 1991), and has a high auto-correlation until almost the next spring (see Fig. 6 of Kitoh, 1988).

Tomita and Yasunari (1993) classified the ENSOs into the two types of the BO-ENSO and the LF-ENSO by occurrence duration. The BO-ENSO has shorter duration and shows the biennial nature. The LF-ENSO has longer duration and contains the low-frequency oscillation. They compare the SSTA and wind stress anomalies between the two ENSO types from their initiation through to their termination. They conclude that the most significant difference between the BO-ENSO and the LF-ENSO is whether a large clockwise eddy in the western Pacific appears or not. They mention that this eddy seems to contribute to the termination of the BO-ENSO through the modulation of the oceanic Mindanao Dome (Masumoto and Yamagata, 1991). The understandings of the BO-ENSO and the LF-ENSO lead us to consider a part of mechanism for the ENSO evolution.

Figures 4a and 4b shows the change of Niño3 SSTA in spring for 1980–94 and 1965–79, respectively. Niño3 is defined here as the region of (150°W–90°W, 2°S–2°N). The change of Niño3 SSTA in spring (referred to as Niño3-DSST) is defined here as the change of Niño3 SSTA from DJF to the following JJA. We can say that the 'L' years denoted in Fig. 4, which are the same as those in Fig. 3, are related to the small Niño3-DSST. This finding is consistent with the suggestion by Yasunari (1985)

Fig. 3. Interannual variabilities of the SCS, and the Niño4 SSTA in DJF during (a) 1980–94 and (b) 1965–79. The SSTA are normalized with respect to their standard deviations. The ‘L’ is defined in the text.

Fig. 4. Interannual variabilities of the change of Niño3 SSTA in spring during (a) 1980–94 and (b) 1965–79. The SSTA are normalized with respect to their standard deviations. The ‘L’ is defined in the text.

and Rasmusson *et al.* (1990); the interface between the biennial and low-frequency oscillations is closely related to the occurrence of the ENSO events.

We are interested in the large-scale background associated with the change of Niño3 SSTA in spring. We present the comparative analysis of the atmospheric circulation anomalies and the SSTA between the “exceptional years (‘L’ in Fig. 3 and Fig. 4)” and the “normal years”.

The atmospheric data from the former National Meteorological Center (NMC) and the outgoing longwave radiation (OLR) data from the National Oceanographic and Atmospheric Administration (NOAA) are used for the composite analysis during 1980–94. Besides, the SST data for 1963–94 is analyzed. Those are compiled by the Japan Meteorological Agency (JMA). The above data sets are interpolated on grids 5° in longitude by 4° in latitude. The local *t*-test (Chervin and Schneider, 1976) is applied to the obtained statistical results. The fields with over 95 % significance are shaded in the

figures of this paper, if necessary.

The large-scale backgrounds related to the summer SCS SSTA are studied in Section 2. The classification of years are defined in Section 3. The large-scale backgrounds related to the winter SCS SSTA are shown in Section 4. The characteristic seasonal change for the classified years is presented in association with the ENSO evolution in Section 5. Section 6 is for the conclusions and discussion.

2. Summer SCS SSTA and the Asian monsoon

The northern summer circulations associated with the SCS SSTA are studied statistically. The summer SCS SST is climatically around 29°C (Fig. 2). The summer SCS SST seem to satisfy necessary conditions to produce the anomalous atmospheric circulations by regulating the convective activities on the spot.

Figure 5a shows OLR differences between the positive and negative SCS SSTA years during 1980–94. The positive SCS SSTA years chosen are 1983, ’87,

Fig. 5. (a) OLR composite in JJA for the positive SCS SSTA, (b) SSTA. The shaded areas are significant.

'88, '92, and '93. The negative SCS SSTA years chosen are 1981, '82, '84, '85, '86, '89 and '94. The magnitudes of SCS SSTA in these years are over a half of their standard deviation (see Fig. 8). Negative OLR anomalies implying more precipitation are not clearly observed over the positive SCS SSTA. Instead, significantly positive OLR anomalies are extended east of the Philippines. The active convection anomalies are observed over the equatorial Pacific and the western Indian Ocean. There are significant negative anomalies near Japan.

Figure 6 shows the 850 hPa wind differences between the positive and negative SCS SSTA years. The easterly wind anomalies expand from the tropical western Pacific through South Asia. The low-level wind anomalies over the SCS seem to be a small part of the global-scale anomalies. It is shown in Fig. 7 that the wind speed over the SCS is closely related to the JJA monsoon wind index (Webster and Yang, 1992) from year to year. The summer monsoon wind index is defined here as the zonal wind shear between 850 hPa and 200 hPa, averaged within (60°–110°E, 2°–18°N) during JJA.

The above statistical study strongly suggests that the summer SCS SSTA does not control the atmo-

sphere, but is controlled by the atmospheric circulations related to the summer South Asian monsoon. The *in-situ* low-level wind seems to be responsible for regulating the SCS SSTA. Actually, we can see in Fig. 7 that the SCS SSTA are negative when the wind speed over the SCS is large, and *vice-versa*. It is supposed that a large wind speed usually contributes to a cooling of tropical SST by enhanced evaporation and ocean mixed layer mixing. Local activity of high-frequency disturbances may be dependent on the time mean wind fields.

Figure 6 also shows that the 850 hPa wind anomalies over South Asia are only part of the significant wind anomalies associated with the JJA SCS SSTA. Other major parts of the wind anomalies are found in the tropical western Pacific (east of the Philippines) and East Asia (south of Japan). The SCS is located climatically near the junction of those three wind branches.

The patterns of the circulation anomalies over East Asia are similar to ones studied by Kurihara (1985), Nitta (1987) and Nitta and Hu (1996). They show that those summer East Asia circulation anomalies are related to the western Pacific SSTA. Murakami and Matsumoto (1994) showed that the

Fig. 6. The same as Fig. 5, except for 850 hPa wind.

Fig. 7. Interannual variabilities of the SCS SSTA, the windspeed over the SCS and the South Asia monsoon index in JJA during 1980-94. These are normalized with respect to their standard deviations.

SCS is the boundary between the South Asia monsoon and the western Pacific monsoon. They concluded that the driving force for the western Pacific summer monsoon is very different from the Southeast Asian summer monsoon; the effects of zonal asymmetry in the Pacific SST at 10°N-20°N are important for the western Pacific monsoon.

We find that the low-level zonal wind anomalies over South Asia, the tropical western Pacific and East Asia fluctuate concurrently with the JJA SCS SSTA to a first approximation (not shown).

Figure 5b shows the composited SST differences

between the positive and negative JJA SCS SSTA years. The significantly positive SSTA are found mostly over the northern Indian Ocean besides the SCS. Any SSTA do not seem to be closely connected to anomalous OLR on the spot. There seems to be no single candidate in the summer SSTA to directly explain the anomalous atmospheric circulations associated with the summer SCS SSTA. The anomaly patterns of the SST and the OLR for the positive SCS SSTA are comparable to the composite ones for the weak South Asia monsoon (Li and Yanai, 1996).

The variabilities of the DJF Niño4 SSTA during 1980-94 and 1965-79 are shown together with the JJA SCS SSTA in Figs. 8a and 8b, respectively. The variability of the DJF Niño4 SSTA is similar to that of the JJA SCS SSTA. Particularly, the signs for those SSTA are the same for all years during the two periods. Ju and Slingo (1995) concluded from their AGCM experiments that the Asian monsoon represented by the summer monsoon wind index, is sensitive to the phase of the ENSO in the preceding winter and spring. The close relationship between the JJA SCS SSTA and the DJF Niño4 SSTA can be understood from the results of Ju and Slingo (1995) and the fact that the JJA SCS SSTA are mostly controlled by the South Asia monsoon activities. This view is discussed in Subsection 4.5.

As indicated by Fig. 7, the qualitative results of the statistical studies in this section are not quite sensitive as to whether to choose the persistent SCS SSTA cases or not (not shown). The results suggest that the summer SCS SSTA are not physically persistent from the previous winter, but created by the summer monsoon circulations.

Fig. 8. Interannual variabilities of the JJA SCS SSTA and the DJF Niño4 SSTA during (a) 1980–94 and (b) 1965–79. The SSTA are normalized with respect to their standard deviations.

3. BO-type and LF-type years

The “exceptional years” and the “normal years” in the introduction are named the BO-type years and the LF-type years, respectively. Those are taken after the BO-ENSO and the LF-ENSO in Tomita and Yasunari (1993). The BO-ENSO events belong to the BO-type years, and the first years of the LF-ENSO events mostly belong to the LF-type years.

The classification is schematically summarized in Fig. 9. For the BO-type years, the DJF SCS SSTA have the same sign as the DJF Niño4 SSTA. For the LF-type years, the DJF SCS SSTA have the inverse sign to the DJF Niño4 SSTA. We note again that the polarities of the JJA SCS SSTA completely coincide with those of the DJF Niño4 SSTA for all years during the two periods 1965–1979 and 1980–94. There are a few exceptional years for the coincidence when the DJF Niño3 SSTA are used instead of the DJF Niño4 SSTA. Furthermore, as shown in Fig. 9, the BO-type years are divided into the BO(+)-type years and the BO(–)-type years, and the LF-type years consist of the LF(+)-type years

and the LF(–)-type years. All years during the two periods 1965–79 and 1980–94 are classified and analyzed, based on the deviations from the respective 15-year averages:

BO(+)-type year: 1983, '88, '90, '91, '92, and 1966, '69, '73, '78, '79,

BO(–)-type year: 1981, '82, '84, '86, '89, and 1965, '72, '74, '76,

LF(+)-type year: 1980, '87, '93, and 1968, '70, '77,

LF(–)-type year: 1985, '94, and 1967, '71, '75.

We will statistically compare the global SSTA and the atmospheric anomalies between the BO and LF-type years. More specifically, the field differences between the BO(+)-type years and the BO(–)-type years and the corresponding differences for the LF-type years are obtained first, then those are compared. The above study process is schematically shown on the right-hand side in Fig. 9. Figure 9 also includes the statistical result for the Niño3-DSST schematically.

4. Winter SCS SSTA and the biennial oscillation

4.1 SSTA

Figure 10a shows the 1980–94 DJF SSTA composite for the BO-type years, which is defined as the SST differences between the BO(+)-type years and the BO(–)-type years. Figure 10b is for the LF-type years. Positive anomalies in the tropical central and eastern Pacific are common for the BO and LF-type years. The positive SCS SSTA for the BO-type years and the negative SCS SSTA for the LF-type years are expected from the year type definition.

Notable differences between the BO and LF-type years are the distribution and magnitude of the tropical eastern Pacific SSTA. For the BO-type years, large positive anomalies expand from the dateline through the eastern Pacific, and those are latitudinally concentrated around the equator. On the other hand, for the LF-type years, small positive anomalies are located over the equatorial central Pacific. This characteristic does not contradict the SSTA differences between the BO-ENSO winter and the first winter of the LF-ENSO, shown by Tomita and Yasunari (1993). The SSTA near the Timor Sea are positive and negative for the BO- and LF-type years, respectively. Those polarities are the same as those of the respective SCS SSTA.

Figures 10c and 10d are the same as Figs. 10a and 10b, except for 1965–79. The field significance may be tested by comparing the two SSTA composites during 1965–79 and 1980–94. The SSTA characteristics for the BO and LF-type years during 1980–94 are also found during 1965–79.

Fig. 9. Classification of years is shown schematically. The BO(+), BO(-), LF(+), and LF(-)-type years are defined. See the detail in the text.

Fig. 10. The DJF SSTa for (a) the BO-type years and (b) the LF-type years during 1980-94. The shaded area is significant. (c) and (d) are the same as (a) and (b), except for during 1965-79.

Fig. 10. (Continued)

4.2 Low-level wind and vorticity

The DJF 850 hPa wind composite for the BO-type years in Fig. 11a is compared with that for the LF-type years in Fig. 11b. For the BO-type years, large westerly anomalies prevail from 160°E through the eastern Pacific along the equator, and easterly anomalies are over the maritime continents. For the LF-type years, the whole structure of the above wind anomalies can be found, but is shifted westward; significant westerly anomalies are found east of 150°E, and significant easterly anomalies are located over the Indian Ocean.

The inverse direction of the wind anomalies is apparent over the SCS between the BO and LF-type years. Southerly anomalies are found over the SCS for the BO-type years, and northerly anomalies are found for the LF-type years. This fact is consistent with the SCS SSTA differences between the two year types. Northeasterly anomalies are clear from the central Pacific in the northern subtropics toward

the maritime continents for the BO-type years, and form anti-cyclonic wind anomalies over the Philippines. Southerly anomalies are observed from the SCS through south of Japan. The cyclonic wind anomalies are located off the equator in the central Pacific in the Northern and Southern Hemispheres. For the LF-type years, northerly wind anomalies diverge from the East China Sea toward the SCS and the tropical western Pacific. Inverse wind anomalies between the BO and LF-type years are detectable around the Timor Sea.

The DJF 850 hPa wind composite for the BO-type years is similar to the winter surface wind composite of the biennial mode (Rasmusson *et al.*, 1990). That is also similar to the winter wind anomaly composite associated with the typical positive SCS SSTA (Tomita and Yasunari, 1996). Moreover, the wind anomalies for the BO and LF-type years are found in the wind stress anomalies for the BO-ENSO winter and the first winter of the LF-ENSO (Tomita

Fig. 11. The DJF 850 hPa wind anomalies for (a) the BO-type years and (b) the LF-type years. The shaded areas are significant. (c) The dark/light shaded area represents almost the inverse/same wind direction between the BO and LF-type years. The light shaded area is enclosed by dashed lines. See the detail in the text.

and Yasunari, 1993), respectively.

The tropical 850 hPa wind composite for the BO-type years is also similar to that in the case of a warm East Asia, which is shown by Hanawa *et al.* (1989). Southerly wind anomalies south of Japan for the BO-type years are consistent with that fact.

The basic structure of the wind anomalies is more clearly identified in the vorticity anomalies. Figures 12a and 12b show the DJF 850 hPa vorticity anomalies for the BO and LF-type years, respectively. For both year types, the cyclonic vorticity anomalies are

found off the equator in the central Pacific. The anti-cyclonic anomalies exist east of the SCS for the BO-type years and west of the SCS for the LF-type years.

From a global aspect, the westward shift of the vorticity anomalies is confirmed for the LF-type years as compared with those for the BO-type years. This means that the wind over the SCS is very sensitive to a small longitudinal shift of the global wind anomaly structure. We can also specify the area around the Timor Sea as another area sensitive to a

Fig. 12. The DJF 850 hPa vorticity anomalies for (a) the BO-type years and (b) the LF-type years. The shaded areas are significant. The units are 10^{-6} s^{-1} .

longitudinal shift of those wind anomaly structures.

4.3 Biennial and low-frequency oscillations

The SSTA and wind anomalies for the BO and LF-type years are summarized from the viewpoint of the relationship between the biennial and low-frequency oscillations. The light shaded area in Fig. 11c is where the 850 hPa wind anomalies are in almost the same direction between the BO and LF-type years; the angles between the wind anomaly vectors in Fig. 11a and those in Fig. 11b are less than 45° . The dark shading is where the angles between them are larger than 90° , close to the inverse direction. The equatorial 850 hPa wind anomalies from the central Pacific through the eastern Pacific are almost in the same direction between the BO and LF-type years. The wind directions are nearly inverse over the SCS and the Timor Sea. Most of the subtropical Pacific is included in the region of the inverse wind direction.

It is reasonable to suppose that differences in the low-level wind direction rather than differences in the wind magnitude result in qualitative change in the oceanic state. According to Shen and Lau (1995), the SSTA oscillation with dominant 2–3 year periods is found over the SCS and near the Timor Sea. The SSTA oscillation with dominant 4–6 year periods is found over the central and eastern Pacific. Those areas with the 2–3 and 4–6 year periods coincide with the regions where the wind anomaly vectors are in almost the inverse direction and the

same direction, respectively, between the BO and LF-type years (Fig. 11c).

4.4 Upper-level circulations

Figures 13a and 13b are the same as Figs. 11a and 11b, except for 200 hPa. The common characteristics for the BO and LF-type years are easterly anomalies along the equator and a pair of anomalous anti-cyclonic circulations in the subtropics over the eastern Pacific. Those are probably the response to the convective anomalies in the equatorial central Pacific (shown later in Fig. 14). Westerly anomalies and a pair of cyclonic circulations are located over the tropical western Pacific for the BO-type years. Similar anomalies are found, but shifted westward over the Indian Ocean for the LF-type years. In the comparison between the BO and LF-type years, the global features of the upper-level circulation anomalies are consistent with the low-level ones.

Westerly/easterly zonal wind anomalies are observed at 200 hPa over South Asia or Southeast Asia in the previous winter of the weak/strong monsoon years (Webster and Yang, 1992, Ju and Slingo, 1995). As shown so far, the upper-level westerly anomalies over South or Southeast Asia in the winter and the weak summer South Asia monsoon are found commonly for the BO and LF-type years. The differences between the BO and LF-type years are the longitudinal position of the global wind anomalies over South Asia and Southeast Asia in winter.

Fig. 13. The same as Fig. 11, except for the DJF 200 hPa wind anomalies.

4.5 OLR

Figures 14a and 14b are the same as Figs. 12a and 12b, except for OLR. The OLRs for 1990, 91 and 92 of the BO-years are not included because of partly missing data. Negative OLR anomalies implying active convection are found in the equatorial central Pacific for the BO and LF-type years. These are associated with warm SSTA in the central Pacific, and distributed symmetrically around the equator. The OLR anomalies over the Indian Ocean and the western Pacific are not necessarily symmetric about the equator. Positive OLR anomalies in the northern subtropics are common for the BO and LF-type years. The positive OLR anomalies are possibly related to the depressed shallow convective cloud in a cold surge. The positive OLR in the subtropics for the LF-type years seems to be shifted westward as compared with one for the BO-type years. The positions of the OLR anomalies in the northern subtropics seem to be related with those for the low-level vorticity anomalies (Fig. 12).

Soman and Slingo (1997) concludes that the summer Asian monsoon is influenced by the latitudinal position and strength of the convection maximum over the western Pacific in the preceding spring. They also show that the convection maximum over

the western Pacific is suppressed by the Walker circulation during El Niño years, and enhanced by the *in situ* SSTA during La Niña years. We could say that the weakened summer Asian monsoon for the BO and LF-type years are consistent with warm Niño4 SSTA and the positive OLR anomalies over the northern western Pacific and the northern Indian Ocean in the winter. Actually, these positive OLR anomalies are maintained over those regions through the following spring.

5. ENSO evolution and the classified years

The concept of the BO and LF-type years is related to the Niño3-DSST (defined in Section 1 as the change of Niño3 SSTA in spring). The statistics of the Niño3-DSST are already shown in Fig. 4.

Figures 15a and 15b shows the longitude-time map of the equatorial SSTA (2°S – 2°N) for the BO and LF-type years, respectively, during 1980–94. The positive DJF SSTA over the central and eastern Pacific decrease with time and become negative toward the end of the second year for the BO-type years. The positive SSTA for the BO-type years appear in the previous summer. On the other hand, for the LF-type years, the positive DJF SSTA in the central and eastern Pacific tend to be maintained

Fig. 14. The same as Fig. 12, except for the OLR anomalies.

Fig. 15. The longitude-time section map for the equatorial SSTA of (a) the BO-type years and (b) the LF-type years. The shaded areas are significant. The time starts at the top of the figure from the preceding June, and ends at the bottom in the second December.

Fig. 16. The same as Figs. 15a and 15b, except for the equatorial 850 hPa zonal wind anomalies.

through the second year. For the LF-type years, the SSTA appear in the preceding autumn. The SSTA around 90°E in the Indian Ocean are significantly positive from the winter through the second summer for the BO-type years.

It is of interest what in the BO-type years is responsible for large Niño3-DSST as compared with small ones for the LF-type years. The SSTA are the results of the complex interaction between atmosphere and ocean. Differences in oceanic state and dynamics are not explored in this study. Only parts of the atmosphere-ocean interaction, that is, the low-level wind anomalies, are statistically studied.

Figures 16a and 16b shows the longitude-time map of the 850 hPa zonal wind anomalies along the equator (2°S – 2°N) for the BO and LF-type years, respectively. The significant westerly wind anomalies are clearly found from the preceding summer through the winter over the central and eastern Pacific for the BO-type years.

It is shown by Kawamura (1988) that the westerly low-level wind anomalies over the central Pacific in autumn are the characteristics of the biennial eastward propagation mode in the warm phase of the eastern Pacific SSTA. Easterly anomalies appear at 90°E in the first autumn, and then move eastward into the maritime continents through the winter.

The longitudinal shift of the equatorial wind anomalies from the first autumn through the winter is shown similarly in the biennial mode of Rasmusson *et al.* (1990). They also pointed out that the wind anomalies at 90°E in autumn are strongly phase-locked to the annual cycle. The easterly anomalies at 90°E in the second summer correspond to the equatorial part of the weak South Asia monsoon jet.

For the LF-type years, the significant westerly wind anomalies over the central Pacific are not found in the preceding summer, but from winter through late spring. Easterly anomalies are located in the Indian Ocean during the winter season, and slightly move eastward in late spring.

Based on the delayed oscillator mechanism, the equatorial 850 hPa zonal wind anomalies for the BO-type years seem to contribute to large Niño3-DSST. The westerly wind anomalies over the central Pacific in the preceding summer may create the cold Rossby wave off the equator in the central Pacific. The easterly wind anomalies over the western Pacific in winter may enhance the cold Kelvin wave in the western Pacific. The equatorial Kelvin wave can cross the Pacific in less than 3 months. The fastest Rossby wave is three times slower (Cane, 1992). Both waves might contribute to the breakdown of the positive Niño3 SSTA in the following spring.

The equatorial zonal wind anomalies for the above two plausible conditions are missing for the LF-type years. The equatorial 850 hPa westerly wind anomalies are found later than the first autumn. It is indicated from the estimated speed of oceanic equatorial waves that a Rossby wave created in autumn or winter does not contribute to the Niño3 SSTA change in the following spring. Spring is basically the preferred season for ENSO's onset and termination due to the climatic atmospheric and oceanic condition (*e.g.*, Tziperman *et al.*, 1997).

Masumoto and Yamagata (1991) showed that a positive curl of the surface stress of the north-east Asian winter monsoon develops the Mindanao dome off the Philippine coast from autumn through winter. Tomita and Yasunari (1993) showed that the wind stress anomaly field has a specific clockwise eddy off the east coast of the Philippines in the northern winter immediately after the BO-ENSO. They concluded that the anomalous clockwise eddy plays an important role in ENSO termination through the Mindanao dome. Tomita and Yasunari (1996) proposed one of the biennial oscillation mechanisms, where they again refer to the wind curl and the Mindanao Current off the Philippines (Lukas, 1988).

From the DJF 850 hPa vorticity anomalies shown in Fig. 12, we can roughly compare the DJF oceanic north-south flow through the Sverdrup equation between the BO and LF-type years. The negative 850 hPa vorticity anomalies for the BO-type years are translated into the equatorward flow anomalies in the ocean east of the Philippines. Those oceanic flow anomalies may create the cold water pool in the tropical western Pacific. No corresponding flow is expected from the vorticity anomalies for the LF-type years. The negative 850 hPa vorticity anomalies for the BO-type years are closely connected to the easterly wind anomalies in the equatorial western Pacific discussed already in this section.

6. Conclusions and Discussion

We study in this paper the large-scale background fields associated with the SCS SSTA. Interannual variability of the SCS SST is recognized as an index for the Asian monsoon and ENSO system. The conclusions obtained in this study are shown below.

(1) Observational analysis suggests that the northern summer SCS SSTA do not control the summer Asian monsoon on the global scale by regulating convective precipitation on the spot. The summer SCS SSTA seem to be controlled by *in-situ* low-level wind anomalies, although the detailed physical mechanism on a regional scale is not clear.

(2) The weak South Asia monsoon jet is related to a weaker SCS low-level wind speed in summertime and, furthermore, a warm SCS SSTA. Westerly wind anomalies in East Asia and easterly anomalies in the

tropical western Pacific are also found at 850 hPa in the case of warm SCS SSTA.

(3) The summer SCS SSTA are statistically related to the winter Niño4 SSTA. This fact indicates the close relationship between the summer Asian monsoon and the equatorial Pacific SSTA in the previous winter.

(4) All years can be classified into 'the BO-type years' and 'the LF-type years' by combinations of the polarities of the winter SCS SSTA and the winter Niño4 SSTA. The two year types are considered as the in-phase and out-of-phase composition of the biennial mode and the low-frequency mode. The features of the biennial mode tend to be found in the BO-type years.

(5) The upper and lower-level wind anomalies for the LF-type years seem to be shifted westward as compared with those for the BO-type years. Large 850 hPa vorticity anomalies are located east of the SCS for the BO-type years. The LF-type years are characterized by large 850 hPa vorticity anomalies west of the SCS and small ones east of the SCS. The biennially oscillating SSTA of the SCS and its neighbors correspond to where the low-level wind directions are quite sensitive to the location of the vorticity anomalies.

(6) Tropical eastern Pacific SSTA change significantly during spring for the BO-type years. On the other hand, they tend not to change for the LF-type years. The low-level zonal wind anomalies over the equatorial central Pacific in the preceding summer and over the equatorial western Pacific in winter are observed for the BO-type years. The above wind anomalies are not found for the LF-type years. The results may be consistent with the delay-oscillator mechanism for the El Niño events.

The winter SCS SSTA tend to persist until the following summer. But we suggest that the summer SCS SSTA are determined by the summer Asian monsoon circulation anomalies. The winter and summer SCS SSTA are basically controlled by a different atmospheric circulation system. The SCS SSTA persistency from winter to summer comes mainly from the close relationship between the summer Asian monsoon and the preceding winter circulations. ENSO is involved in that relationship.

We suppose that the possibility is quite low that the SCS SSTA plays the key role in the biennial oscillation of the Asian monsoon and ENSO system, at least through summertime. However, the winter SCS SSTA is an available index for the biennial oscillation. This is because the SCS occupies the special geographical location for the tropical low-level circulations in winter.

The summer SCS SSTA are associated with the three low-level wind anomaly branches of South Asia, the western Pacific, and East Asia. This fact

is also due to the special location of the SCS, off the southeast corner of the Continent. The summer SCS SSTA may be said to represent the summer Asian monsoon over the extended regions.

The concern on the study is that the period of the atmospheric data (1980–94) to hand is not necessarily long enough to make a statistically stable analysis although a local significance test is performed. All years including the nearly normal years are classified into either of the two year types, and used for the composite analysis. In order to confirm the statistical results during the short period, the interannual variabilities are checked for some fields.

The explanation of the obtained results is not a full one. The major reason is that, though the El Niño events probably play an central role, the studies were made on the analysis of atmospheric data. We have neither analyzed the tropical ocean data nor used the atmosphere-ocean coupled model in order to explore the oceanic state. This work may be done in the future.

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南シナ海の海面水温 —アジアモンスーンと ENSO 系のひとつの指標—

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南シナ海の海面水温の年々変動は、その地理的位置のため、アジアモンスーンと ENSO のシステムの指標と認識できる。観測値の統計的解析から、次のような結果が得られた。

(1) 北半球冬において、南シナ海の海面水温偏差は、赤道太平洋の海面水温偏差と関係した大規模な風偏差の経度方向の変化に対して、極めて敏感である。この事実は、南シナ海およびその周辺の海面水温偏差が強い 2 年振動を示すことと関係している。

(2) この大規模な風偏差が東に寄っているとき (BO タイプ年)、熱帯東太平洋海面水温偏差は、次の春に大きな変化を示す傾向にある。一方、大規模な風偏差が西に寄っているとき (LF タイプ年)、熱帯東太平洋海面水温偏差は、その年をとおして維持される傾向にある。これに関連した BO タイプ年と LF タイプ年の違いが、前年の夏から冬にかけての熱帯の下層風偏差の中に見られた。

(3) 北半球夏の南シナ海の海面水温偏差は、そこでの下層風偏差によって決まっているように見える。さらに、南シナ海の正の海面水温偏差に対して、南アジアおよび熱帯西太平洋で東風偏差、東アジアで西風偏差が下層で見られる。また、夏の南シナ海の海面水温偏差は、その前の冬の赤道中央太平洋の海面水温偏差と統計的に関係があることが示される。このことは、夏のアジアモンスーンと冬の ENSO の位相に関係があることを示す。