Monsoon and ENSO: A Coupled Ocean/ Land/Atmosphere System

Introduction

The El Niño/Southern Oscillation (ENSO) phenomenon is often regarded as primarily being an interannual oscillation of the coupled ocean/atmosphere system in the tropical Pacific, and numerous recent studies using simple atmosphere/ocean coupled models (e.g., Philander *et al.*, 1984; Zebiak and Cane, 1987) have substantiated this idea. On the other hand, recent observational studies also have shown that there exists a tight link between the Asian/Australian monsoon and the atmosphere/ocean system in the tropical Pacific (e.g., Barnett, 1985, 1988; Neehl, 1987; Yasunari, 1987, 1990a). Furthermore, the interannual variability of Asian summer monsoon activity also has proved to be strongly influenced by the winter and spring snow cover over the Eurasian continent (e.g., Hahn and Shukla, 1976; Dickson, 1984). Thus, this theoretical and observational evidence suggests that the coupled ocean/atmosphere system in the tropical Pacific *and* the Asian monsoon system is one of the major climate systems on the earth, i.e., a coupled ocean/land/atmosphere system.

Indian Monsoon and SWT/SST in the Tropical Pacific

Figure 1 shows a highly coherent fluctuation between the all-India monsoon rainfall (IMR; Parthasarathy, 1987) and the sea water temperature (SWT) of the oceanic mixed layer in the tropical western Pacific along 137°E in the following winter (January), basically with a biennial time scale (Yasunari, 1990a).

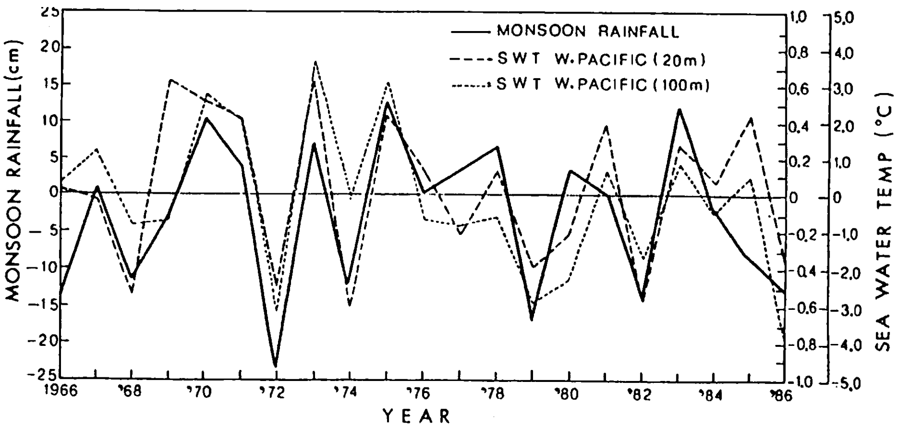


Figure 1. Time series of Indian monsoon rainfall anomaly (thick solid line) and SWT anomaly at 20 m (thick dashed line) and 100 m (thin dashed line) depth averaged for the 137°E line (2°N-10°N) in the succeeding January. (Yasunari, 1990.)

Figure 2 shows lag-correlations between IMR and sea surface temperature (SST) anomalies for the equatorial eastern and western Pacific. It is noteworthy that the correlations of SST to IMR, whether positive or negative, gradually increase through the seasonal march from summer of Y(0) (which denotes the year of summer monsoon referenced) to the following winter, reaching a maximum in January or February of Y(+1) in the two regions. In the western Pacific, the maximum value of the correlation coefficient is particularly high (+0.79), while in the eastern Pacific the negative correlation is apparent at the same time.

The evidence strongly suggests that the Asian summer monsoon activity is greatly responsible for the fluctuation of the heat content anomaly both in the warm water pool of the western Pacific and in the oceanic surface layer of the eastern Pacific.

Indian Monsoon and Wind Field over the Tropical Pacific

The apparent relationship between the IMR and the SWT or SST in the tropical Pacific should be physically based upon anomalies in the surface wind stress field, which is associated with the activity of the Asian summer monsoon as shown in the IMR (Figure 1).

Figure 3 shows, for example, the seasonal evolution of SST and zonal wind anomalies at 700 mb (as a substitute for the surface wind) along the equatorial belt, composited for two years by using the six-pair years starting with the strong monsoon years (i.e., 1971, 1973, 1975, 1978, 1981 and 1983). Because of the strong biennial nature of monsoon variability, the second years are mostly weak monsoon years, including the ENSO years. It is noteworthy that an easterly anomaly starts to develop

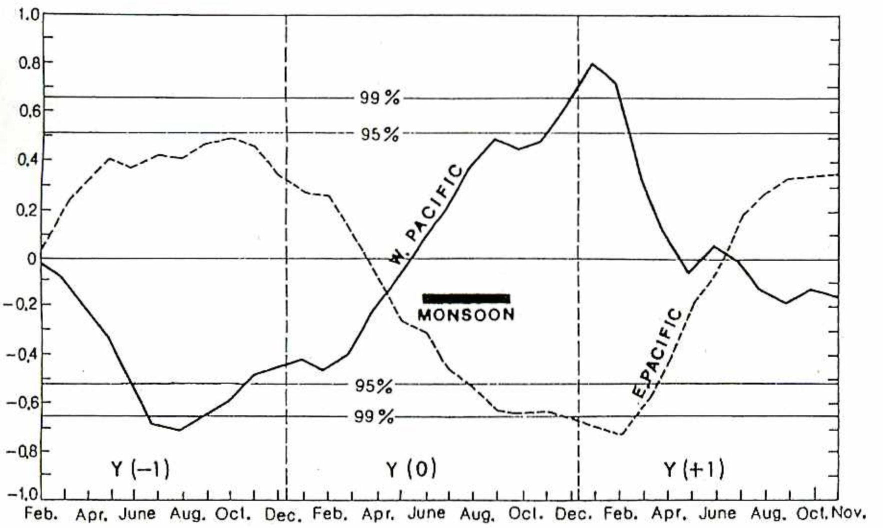


Figure 2. Lag correlations between Indian monsoon rainfall anomaly and SST anomaly in the western (0°-8°N, 130°E-150°E) and the eastern (0°-8°N, 170°W-150°W) Pacific. The reference monsoon season is shown with thick black bar. Y(0) denotes the year of reference monsoon, and Y(-l) (Y(+l)) denotes the year before (after) Y(0). (Yasunari, 1990.)

over the whole Pacific basin at nearly the same time that the strong Indian (Asian) summer monsoon has started. The development of an easterly anomaly in the western Pacific from summer to early winter is followed by a significant negative SST anomaly in the central and eastern Pacific and a positive SST anomaly in the western Pacific. These anomalous states in the SST and wind field for the strong monsoon year are completely reversed in sign at or just before the weak monsoon of the next year starts. This interesting feature of interannual anomalies over this region has prompted us to propose a concept of the "monsoon year" as a unit climatic year in the tropics (Yasunari, 1990b).

It should be pointed out here that the anomalous wind field over the western Pacific to the west of the dateline (150°E to 170°E) seems to be very important for producing the SST anomaly over the whole equatorial Pacific, which is apparently shown in the lag correlation diagrams between the zonal wind at particular longitudes from the west to the east and the SST at all longitudes in the Pacific basin (Figure 4). The zonal wind (stress) anomaly precedes as a whole the SST anomaly, particularly in the longitudes of 150°E to 170°E. The zonal wind anomaly in these longitudes during summer through autumn also shows the highest correlation to the IMR.

Atmospheric Circulation Associated with the Wind Field over the Equatorial Pacific

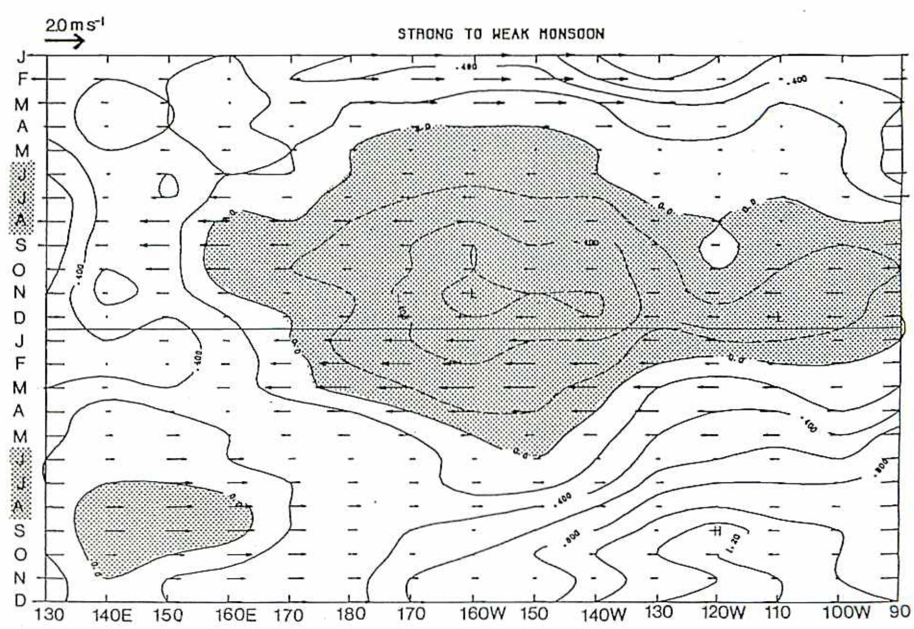
The important role of the zonal wind (stress) field along the equatorial western/central Pacific in the formation of SWT and SST anomalies has been stressed in the previous sections. This zonal wind anomaly should be associated with anomalous atmospheric circulation of larger scale.

Within the tropics, this anomalous zonal wind could be interpreted as part of the anomalous tropical east-west (or Walker) circulation, associated with the stronger (or weaker) convective activity over the Asian monsoon through the western Pacific region (e.g., Meehl, 1987; Yasunari, 1990a). However, this anomalous eastwest circulation seems to be closely associated with the preceding or concurrent anomalous circulation in the extratropics.

Figure 5, for example, shows the anomalous wind stress field over the northern Pacific, composited for April of weak monsoon years minus strong monsoon years. One remarkable feature for the weak monsoon years is the overall weakening of the north Pacific (subtropical) high, as shown with the persisting cyclonic circulation anomaly over the northeast Pacific. This anomalous state of the north Pacific high seems to occur from the preceding late winter to the following autumn of the weak Asian summer monsoon.

Another prominent but more localized feature is the (north)easterly anomalies over

Figure 3. Longitude-time section of SST anomaly and zonal wind anomaly vector at 700 mb along the equator composited for the two years from active to weak Indian monsoon year. Positive (negative) values in SST anomaly are shown with solid (dashed) lines (unit: 2.0°C) and negative values shaded. (Yasunari, 1990.)



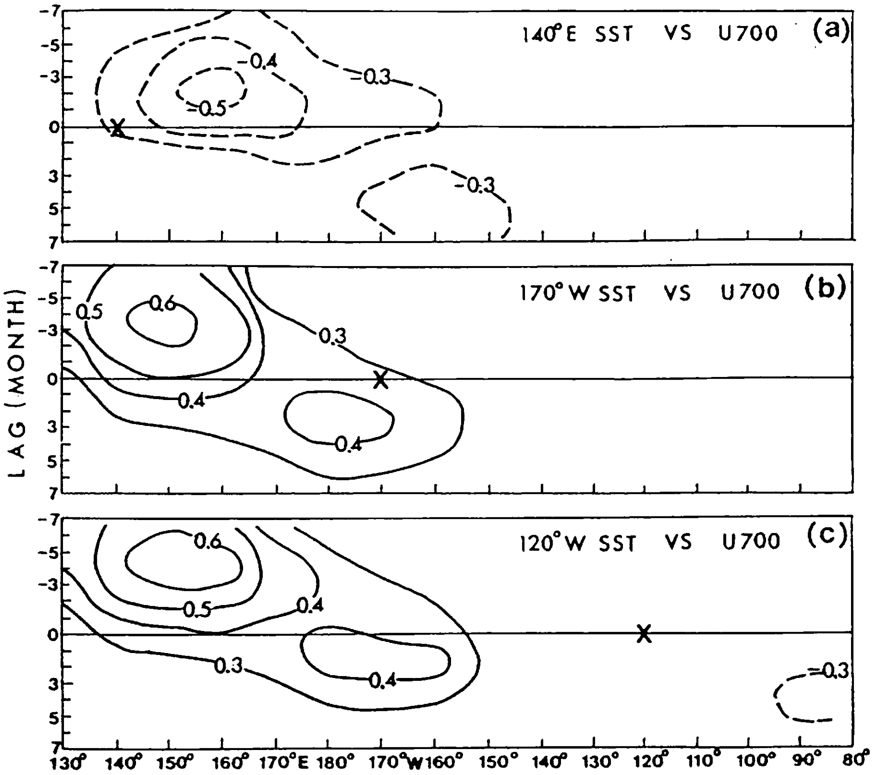


Figure 4. Lag correlations between 700 mb zonal wind anomalies along the equator and SST anomalies at (a) 140°E, (b) 170°W and (c) 120°W at the equator. Contours of more (less) than 0.3 (-0.3) are shown with solid (dashed) lines. (Unit is 0.1.) Negative lags imply that zonal wind anomaly leads SST anomaly. (Yasunari, 1990.)

the western Pacific to the south of Japan, combined with the westerly anomalies along the equatorial western Pacific from the beginning of the year (January) through late summer (August or September). These wind stress anomalies form the localized anomalous cyclonic circulation over the western Pacific to the east of the Philippine Islands. Very recently, Masumoto and Yamagata (1990) pointed out the importance of this localized circulation anomaly for the destruction of the warm water pool, which may, in turn, be a precondition for the warm event (i.e., El Niño) in the eastern Pacific. In the strong monsoon years, entirely the opposite conditions prevail, which favors producing and maintaining a warm water pool in the western Pacific and cool water in the eastern Pacific (i.e., La Niña condition).

Role of the Land/Ocean Heating Contrast

The anomalous circulation over the north Pacific discussed above should be part of global, or at least hemispheric, circulation anomalies associated with the Asian monsoon activity. Yasunari (1988) found some dominant circulation patterns in the northern extratropics which precede or follow the Asian summer monsoon. For example, the weakened north Pacific high in the preceding winter/spring of weak monsoon seems to be coupled with the weakened Aleutian low, associated with the weakened PNA pattern (Wallace and Gutzler, 1981). This weakened PNA seems to be closely linked to the circulation anomaly upstream, i.e., over the Eurasian continent. It is now well known that the IMR shows a very high negative correlation with the preceding winter/spring snow cover over the Eurasian continent (e.g., Hahn and Shukla, 1976). Morinaga and Yasunari (1987) noted that the variation of this Eurasian snow cover is closely associated with the local circulation pattern called the Eurasian (EU) pattern.

One of the most important physical processes accounting for the anomalous circulation over Eurasia through the Pacific basin should be the anomalous differential heating between the Eurasian continent and the surrounding oceans. Indeed, Yasunari and Kodera (1990) noticed a very high linear correlation between the Eurasian snow cover in spring and the SWT in the tropical western Pacific the following winter, as shown in Figure 6. It is noteworthy that this correlation (R=-0.71) is as high as, or even higher than, those of the IMR vs. snow cover (r=0.54) and the SWT vs. the IMR (R=0.69) during the same data period of 22 years. Very recently, Nikaidou (personal communication) found through a GCM experiment that the weak (strong) land heating (or land/ocean heating contrast) during spring to summer is largely responsible for the weak (strong) subtropical high over the ocean. This GCM result may give us a very plausible physical explanation for the observational evidence shown here.

Monsoon as a Transmitter of Climatic Signals

We may emphasize now that the coupled ocean/land/atmosphere system over Eurasia through the tropical Pacific acts as a transmitter of climatic signals from the extratropics (or the continents) to the tropics (or the oceans), by giving renewed boundary con-

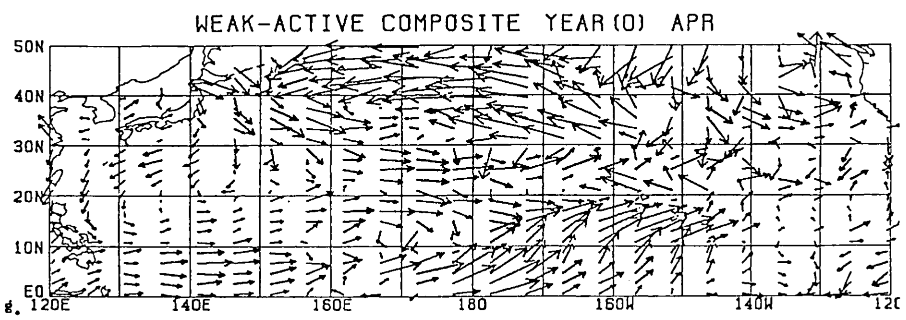


Figure 5. Differences of surface wind stres: vectors of April between the composite: for the weak Indian monsoon years anc those of the strong Indian monsooı years. Wind stress data are kindl: supplied by Dr. K. Kutsuwada.

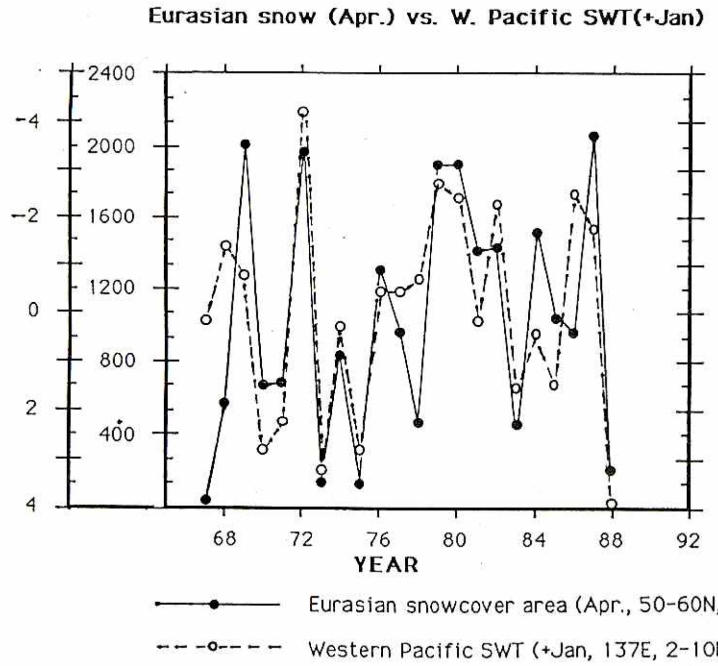


Figure 6. Time series of Eurasian snow cover in April (black circle: solid line) and sea water temperature (white circle: dashed line) of the western tropical Pacific in the following January. Units are km² and °C. (Yasunari and Kodera, 1990.)

ditions (e.g., snow cover, soil moisture) for the more or less deterministic climate system in the tropics. On the other hand, this system transmits climatic signals from the tropics to the extratropics, by means of Rossby wave propagation forced, for example, by SST anomalies (e.g., Horel and Wallace, 1981). In this case, the system in the tropics may give a renewed initial condition for the more or less non-deterministic(or intransitive) system in the extratropics.

The processes of formation and transmission of climatic signals mentioned above are schematically shown in Figure 7. It is noteworthy that these processes seem to have a strong seasonal dependence both in the tropics and in the extratropics. The "monsoon year" in the tropics (Yasunari, 1990b) may be derived from the characteristic physical nature of the system in the tropics.

Thus, the monsoon variability and the ENSO should be considered as part of the oscillation in the coupled ocean/land/atmosphere system, which involves the largest continent and the largest ocean on earth.

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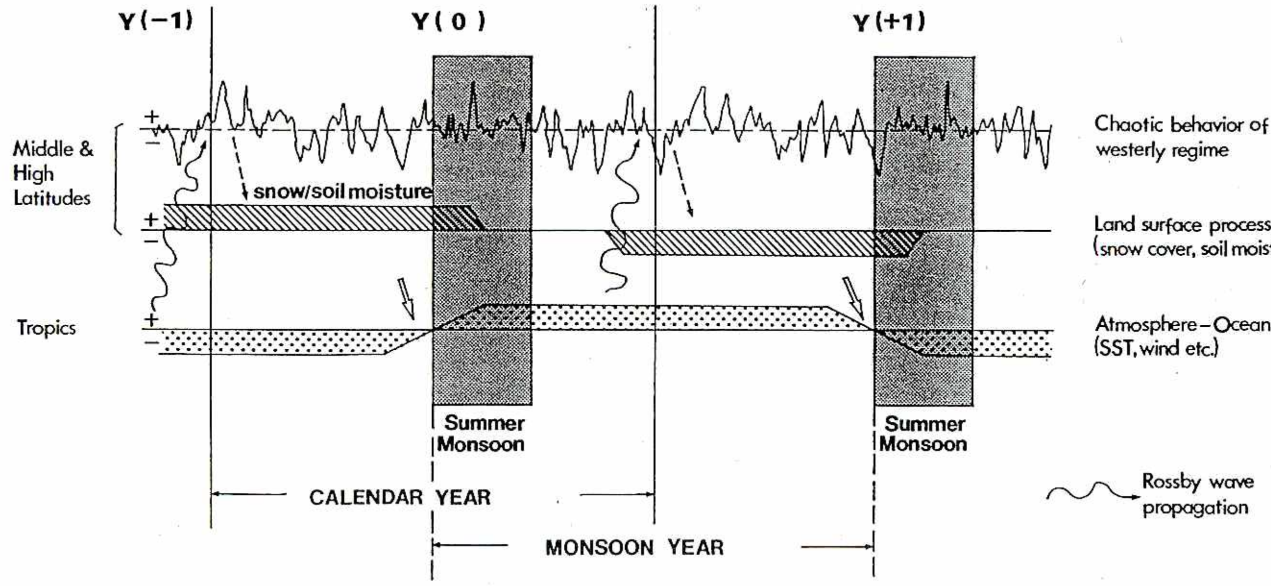

Figure 7. Schematic diagram of formations and transmissions of climatic signals between the tropics and the extratropics in the seasonal cycles. Dashed arrow indicates weather processes (snowfall, etc.), which produce anomalous surface boundary conditions. White arrow indicates the effect of anomalous surface boundary conditions to the tropical atmosphere.

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