Role of the Asian Monsoon on the Interannual Variability of the Global Climate System

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Abstract

The role of the Asian summer monsoon on the interannual variability of the global climate system particularly relevant to the ENSO time scales is discussed, by examining the statistical and dynamical links between the Asian summer monsoon, the atmosphere/ocean system in the tropics and the westerly flow régimes in the extratropics.

The Asian monsoon, the ocean and the atmosphere in the tropical Pacific are tightly linked together as one climate system, named here as the MAOS (Monsoon and the Coupled Atmosphere/Ocean System). The MAOS prominently shows the biennial oscillatory nature which tends to have anomalous states starting in the northern summer monsoon season and persisting for about one year (Yasunari, 1990a: 1991).

The anomalous state of the MAOS produces the anomalous atmospheric circulation over the subtropics and the extratropics of the north Pacific during summer through the early winter, through the modulation of the subtropical high and the stationary Rossby wave propagation mechanism. In the mid winter, this anomalous circulation over the north Pacific is evolved to the hemispheric winter anomalous circulation with wavenumber-one and/or-two structure.

The anomalous circulation over Eurasia associated with this hemispheric anomalous flow régime seems to provide a favorable condition for the extensive (or diminished) snow cover area over central Asia, which in turn is responsible for the reversed anomalous state of the next Asian summer monsoon and the MAOS. That is, the biennial nature of the climate system in the northern hemisphere may be due, at least partly, to this two-way interactions between the tropics and the extratropics. In these processes, the Asian monsoon plays a key role as a transmitter of climate signals between the tropics and the extratropics through the land/atmosphere/ocean interaction in the seasonal cycle.

In addition, it is strongly suggested that the North Atlantic Oscillation (NAO), in reality, plays a crucial role in the timing of the occurrence of the ENSO event, by stochastically amplifying or damping the biennial oscillation of this coupled climate system. That is, the more or less irregular ENSO cycle may result from this interaction between the MAOS and the NAO, where the former seems to have the nature of an almost-intransitive climate system, while the latter seems to represent the more chaotic nature of the westerly flow régime.

1. Introduction

The Asian summer and winter monsoon form huge circulation systems in the general circulation of the global atmosphere (Krishnamurti, 1971; 1973). Simply speaking, this system is characterized as a zonal asymmetry as well as a meridional asymmetry in the tropics. Figure 1 shows the monthly mean sea level pressure and surface wind vectors (a) and the monthly mean velocity potential and divergent wind vector at 200 mb (b) in July 1990 (JMA, 1990). This figure apparently shows a predominant system with the zonally-oriented east-west (or Walker) circulation over south Asia through the southern subtropical Pacific, as well as the meridionally oriented monsoon circulation over the south Asia through the Indian Ocean. This remarkable zonal asymmetry of the circulation has proved to be maintained and modulated through the interaction between the Asian monsoon and the coupled atmosphere/ocean system in the equatorial Pacific (Yasunari, 1990a). This interaction seems to play a key role on the mechanism of the El Niño/Southern Oscillation (ENSO), as has been discussed by Barnett (1985, 1988) and Yasunari (1987). That is, a weaker (stronger) than normal Asian summer monsoon is very favorable for triggering the El Niño (anti-El Niño or La Niña (Philander, 1985)) state of the equatorial Pacific through the weaker (stronger) east-west circulation in the tropics.

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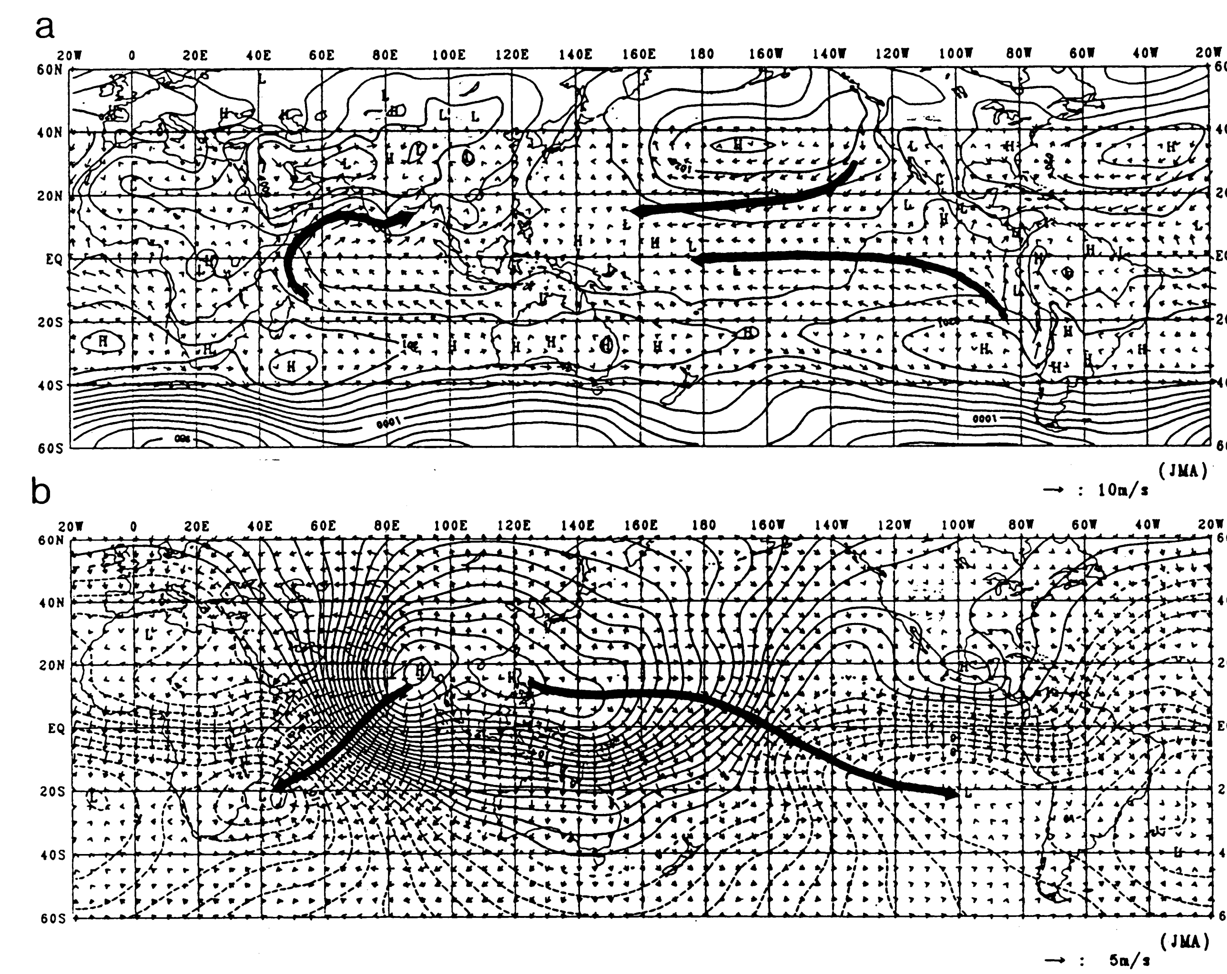

Fig. 1. Monthly mean sea level pressure (a) and velocity potential and divergent wind vector at 200 mb (b) for July 1990. Contour interval in (b) is 1.06×10 m^2/s. (JMA,1990)

In addition, Yasunari (1990b) and Kodera and Chiba (1989) pointed out a possible link of the Eurasian snow cover the preceding spring to the anomalous state of the tropical and subtropical Pacific, presumably through the anomalous state of heating contrast between the continent and the oceans. The observational evidence of the association between the Eurasian snow cover extent and the succeeding summer monsoon shown firstly by Hahn and Shukla (1976) was confirmed by the general circulation model (GCM) studies by Barnett *et al.* (1989) and Yasunari *et al.* (1991). These results strongly suggest that the extra-tropical forcing through the Asian summer monsoon should be incorporated into the physical process of the ENSO.

On the other hand, recent numerous studies have shown that the anomalous state of the equatorial Pacific, either El Niño or La Niña, greatly influences the anomalous state of the circulation in the extra-tropics through stationary Rossby wave propagations (*e.g.*, Hoskins and Karoly, 1981; Webster, 1981; Horel and Wallace, 1981; Palmer and Mansfield, 1984). These results have even brought the idea that the potential predictability of the long-range forecasting of the extra-tropical climate owes its major part to the anomalous state of the tropical climate system.

How, then, may the ENSO cycle, or more generally, the interannual variability of the global climate system, occur through the interaction between the tropics and the extra-tropics? This paper attempts to solve this problem, by reviewing and re-examining each process of the interactions in more detail. Particularly, the role of the Asian summer monsoon will be focused on as a transmitter of climatic signals between the tropics and the extra-tropics in the seasonal cycle.

2. Asian monsoon and the coupled ocean/atmosphere system in the tropical Pacific

Since the early part of this century (Walker and Bliss, 1932) the relationship between the Asian summer monsoon and the ENSO has been manifested by numerous studies, showing that the occurrence of the ENSO is closely associated with the preceding weak Asian summer monsoon as shown in Fig. 2.

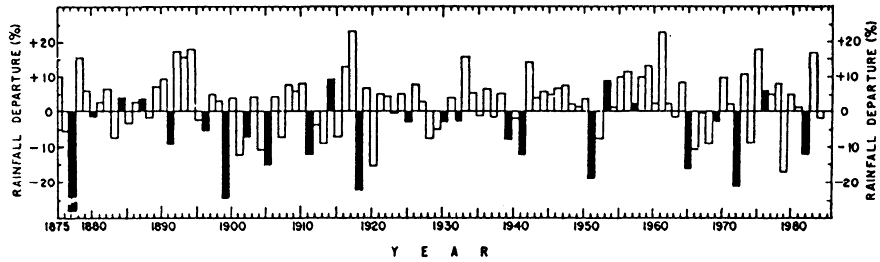

Fig. 2. Interannual variation of all-India monsoon rainfall. The ENSO years are indicated with black bars. (Mooley and Shukla, 1987)

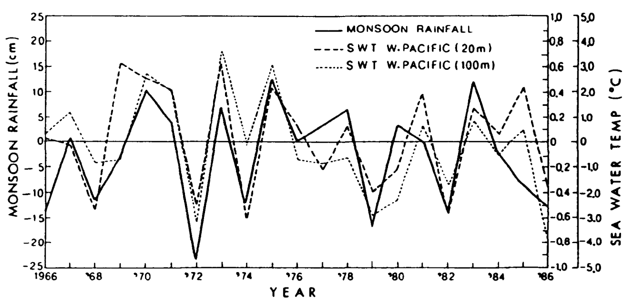

Fig. 3. Time series of Indian monsoon rainfall anomaly (thick solid line) and sea water temperature anomaly at 20 m (thick dashed line) and 100 m (thin dashed line) depth averaged for 137E line (2N–10N) in the succeeding January. (Yasunari, 1990a)

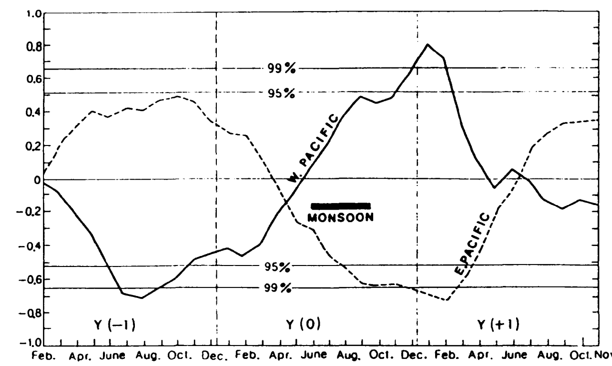

Fig. 4. Lag-correlations between the Indian monsoon rainfall anomaly and the sea surface temperature anomaly in the western (0–8N, 130E–150W) and the eastern (0–8N, 170W–150W) Pacific. The reference monsoon season is shown with a thick black bar. Y(0) denotes the year of reference monsoon year and Y(−1) (Y(+1)) denotes the year before (after) Y(0). (Yasunari, 1990a)

However, Meehl (1987) and Yasunari (1990a) noted that the Indian monsoon activity seems to play an active rather than a passive role in determining the anomalous state of the warm water pool in the western Pacific through the anomalous east-west circulation in the following autumn and winter seasons, as shown in Fig. 3. These studies further substantiated the tight link between the Asian summer monsoon fluctuation and the anomalous state of the coupled ocean/atmosphere system in the tropical Pacific, and suggested that the Asian summer monsoon and the coupled system in the tropical Pacific should be understood as a climate system combined together in the tropics. Hereafter, we refer to this climate system by the abbreviation MAOS (*i.e.*, Monsoon/Coupled Atmosphere Ocean System).

A prominent feature of the MAOS is its oscillatory nature on the quasi-biennial time scale. In addition, an anomalous state of this MAOS shows a strong phase lock to the seasonal cycle, as is apparently seen in Fig. 4. This strong seasonality of the anomalous state of the MAOS has prompted us to introduce a concept of the "monsoon year" (Yasunari, 1991) as a unit climatic year in the tropics. In this context, an ENSO event may be understood as a considerably amplified phase of the oscillation in the MAOS.

3. Influence of the extra-tropics on the MAOS

The Asian summer monsoon is forced primarily by the differential heating between the Eurasian continent and the surrounding oceans, though the moist process in the atmosphere is essential for maintenance of the gigantic monsoon circulation. However, this monsoon circulation is supposed to be very sensitive to the surface heating condition (Charney and Shukla, 1981). Blanford (1884) first noted this aspect, by examining the relationship between the Himalayan snow cover in winter and the following Indian summer monsoon. Hahn and Shukla (1976), Dickson (1984), Dey and Bhanu Kumar (1984) and others have re-examined this Eurasian snow cover-Indian monsoon connection by using the data of satellite-derived snow cover extent. Morinaga (1992) pointed out that the snow cover over central Asia near the Caspian Sea in late winter through spring shows significant negative correlation with

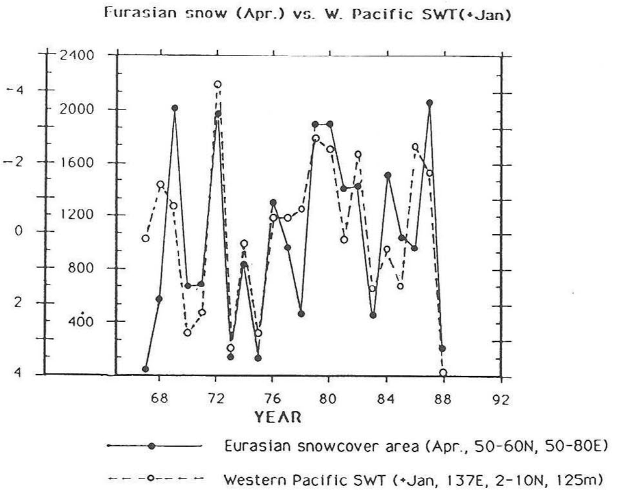

Fig. 5. Time series of the snow cover extent anomaly over central Asia in April (Kodera and Chiba, 1989) and the sea water temperature anomaly at 125 m depth in the western tropical Pacific in the following Janurary for the period of 1967 to 1988.

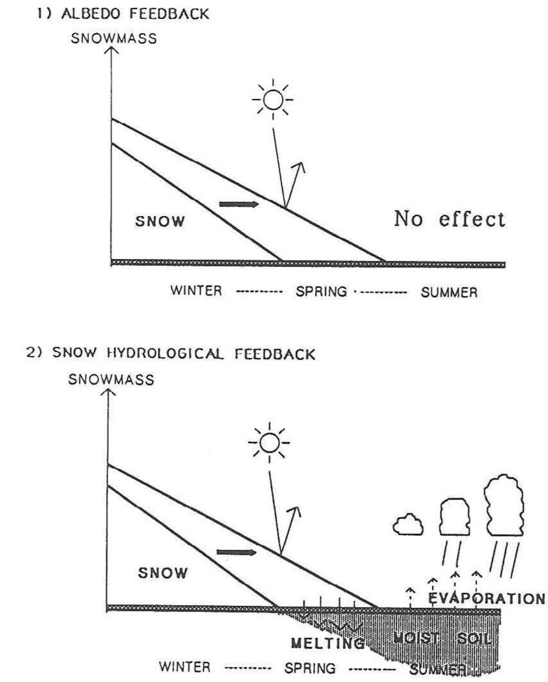

Fig. 6. Schematic diagram for the albedo feedback and the hydrological feedback of snow cover during the seasonal march from winter to summer. (Yasunari *et al.*, 1991)

the Indian monsoon rainfall (IMR) (Parthasarathy, 1987).

The anomalous state of the surface conditions (*e.g.*, snow cover, soil moisture) of the continent in winter and spring, therefore, should affect the anomalous state of the MAOS for each monsoon year. The apparent time-lag correlation between the snow cover extent in central Asia in April (Kodera and Chiba, 1989) and the oceanic mixed layer temperature of the tropical western Pacific in the succeeding January as shown in Fig. 5, strongly supports this idea. Undoubtedly, this time-lag correlation should physically be based upon the anomalous state of Asian monsoon and the associated coupled ocean/atmosphere system in the tropical Pacific persisting from summer through winter, which may be induced by the anomalous snow cover over the continent.

There seem to be two physical processes responsible for the time-lag correlation between the winter snow cover and the summer monsoon. One is the albedo effect and another is the snow-hydrological effect, as shown schematically in Fig. 6. The latter effect involves the melting of an anomalous snow mass, moistening of the soil and an anomalous evaporation, which substantially reduces the heating of the atmosphere from the ground surface in the warmer seasons. The importance of this process was noted by Yeh *et al.* (1983), and was further substantiated in the GCM experiments by Yamazaki (1989), Barnett *et al.* (1989) and Yasunari *et al.* (1991). These recent GCM studies also noted that the combined albedo/snow hydrological effect is actually important for this "time-lagged" effect.

On the other hand, Morinaga and Yasunari (1987) deduced the atmospheric circulation pattern in winter responsible for the maximum (minimum) snow cover extent over central Asia. Figure 7 shows the lag-correlation pattern between the snow cover extent over central Asia in February and the 500 mb height anomaly of the northern hemisphere in the preceding December. This figure indicates that the heavy snowfall (and the large snow cover extent) over central Asia is associated with the anomalous deep trough over there combined with the anomalous ridges over Europe and east Asia. This pattern may be identified as a sort of "Eurasian (EU)" pattern (Wallace and Gutzler, 1981). That is to say, the anomalous winter circulation pattern over the Eurasian continent shown here seems to be a precursor signal affecting the anomalous state of the MAOS in the following monsoon year.

4. Subtropical and extratropical responses to the MAOS

Since Bjerknes (1969), it has been noted that the El Niño event, or more exactly, the anomalous state of the atmosphere over the equatorial central/eastern Pacific associated with the sea surface temperature (SST) anomaly there, produces the anomalous circulation over the north Pacific through the north American sector. Horel and Wallace (1981) demonstrated the characteristic circula-

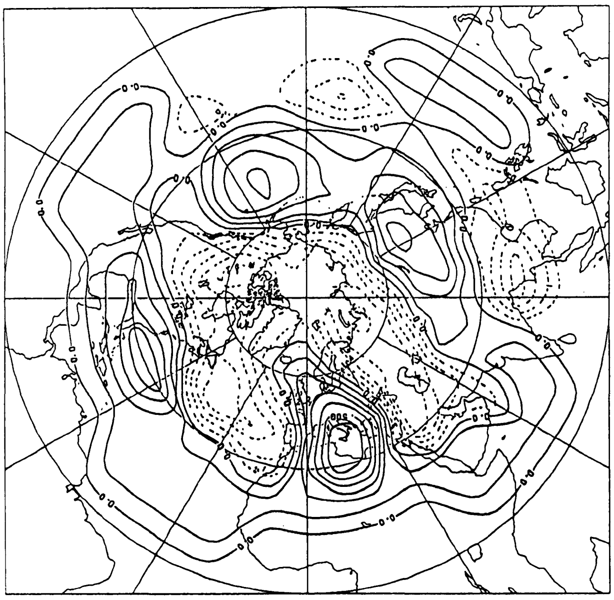

Fig. 7. Lag-correlation map of the 500 mb geopotential height anomaly in December and the snow cover extend anomaly, over central Asia in the following February. (Morinaga and Yasunari, 1987)

tion patterns (*e.g.*, PNA, NAO, WP, Eu *etc.*) associated with the ENSO events. Hoskins and Karoly (1981) and Webster (1981) gave a theoretical basis for these observations as a forced Rossby wave propagation from the tropics. Many GCM studies (*e.g.*, Blackmon *et al.*, 1976 *etc.*) have verified the extra-tropical responses to the SST anomalies in the equatorial Pacific with the El Niño condition.

However, the extra-tropical responses to the SST anomalies in the equatorial Pacific should be examined more generally with reference to each anomalous state of the MAOS, since the extra-tropical response is not only limited to the El Niño conditions. Palmer and Mansfield (1984), for example, showed that the atmospheric response over the north Pacific is as strong as, and even stronger during the La Niña conditions (namely, higher SST anomalies in the western Pacific) than the El Niño conditions.

Figure 8 shows the monthly wind stress anomaly in the northern Pacific subtropics and mid-latitudes, composited for weak summer monsoon years minus strong summer monsoon years, to show more clearly the features for the weak monsoon year. It is apparent that cyclonic circulation anomalies are dominant in the northeastern Pacific (40°N–50°N, 170°W–140°W) from July through October, indicating the weaker subtropical high during summer through the following autumn. This result is well correlated with the westerly anomalies along the equatorial belt, which implies a weaker state of the MAOS with the weaker east-west circulation (*i.e.*, weaker trade wind and upper tropospheric wind sys-

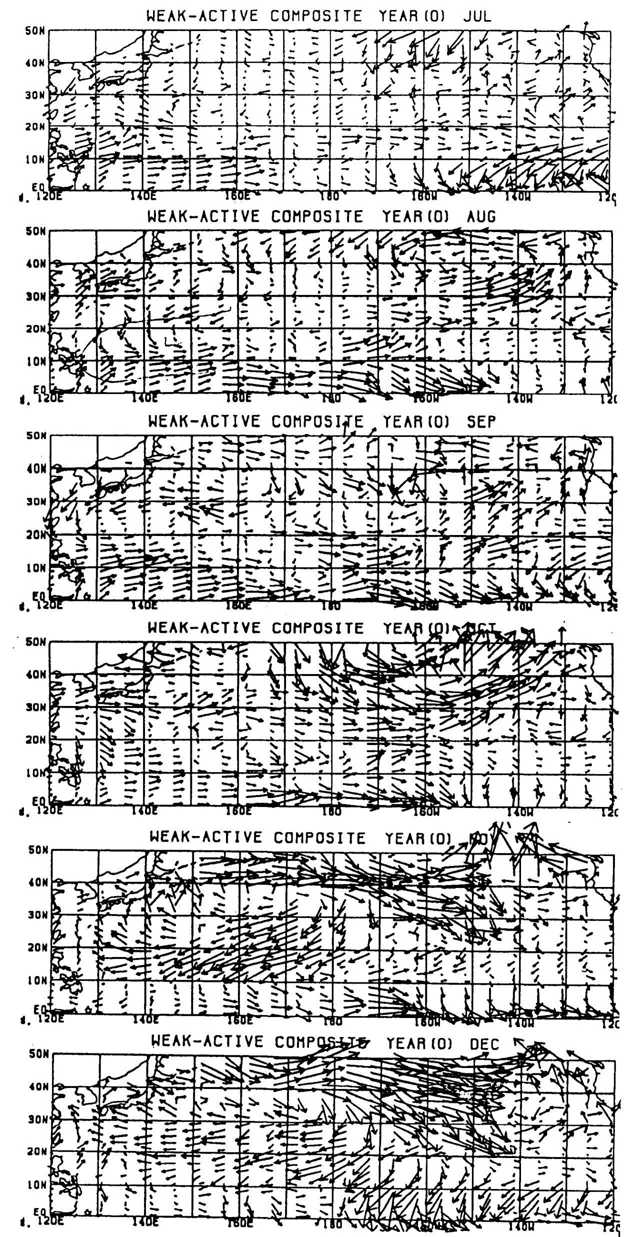

Fig. 8. Anomaly surface wind stress vectors composited for July through December for weak Indian monsoon years minus strong Indian monsoon years. (Yasunari, 1990b)

tem) in the tropics. The persistent cyclonic circulation anomaly in the same region in November and December, on the other hand, corresponds to a stronger than normal, or more equatorward shift of the Aleutian low.

To deduce the dominant regional circulation patterns in the northern mid-latitudes associated with the anomalous state of the MAOS, the time coefficients of some localized dominant circulation (or teleconnection) patterns are correlated to the Indian summer monsoon index. These patterns are obtained by applying the varimax-rotated empirical or-

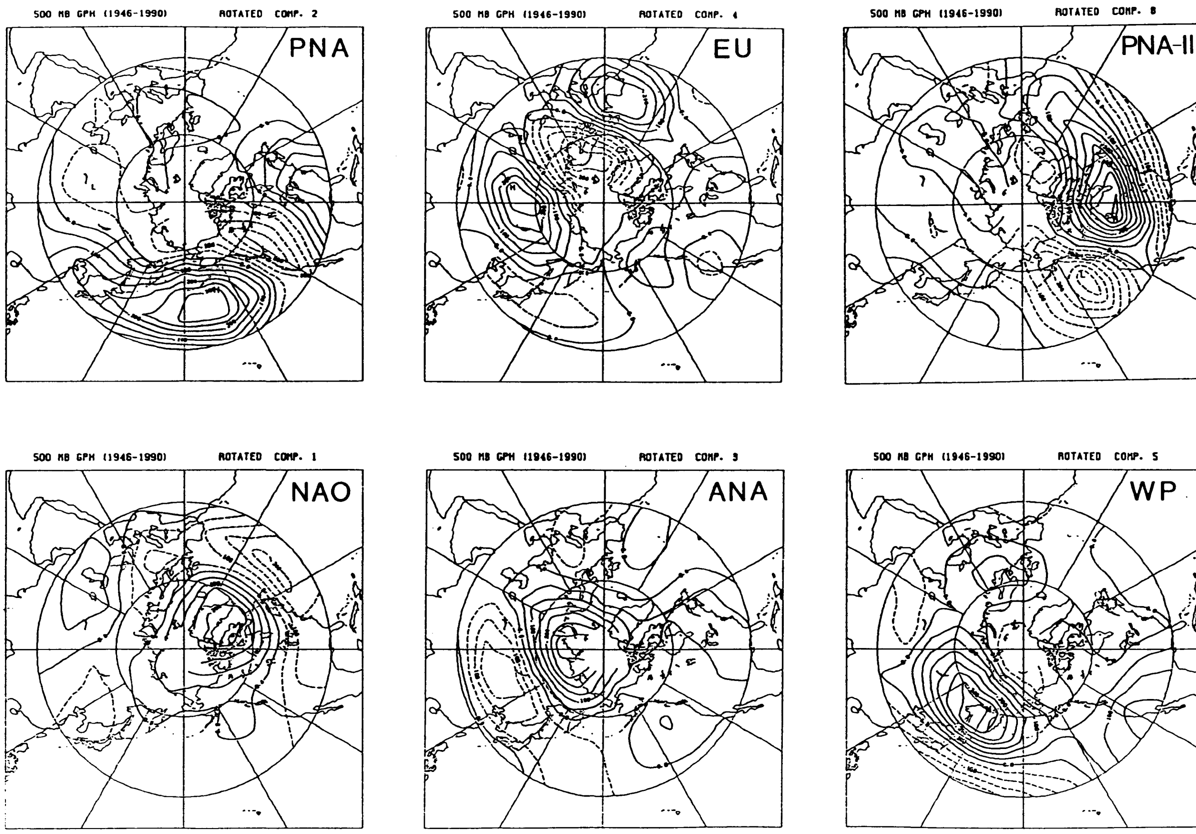

Fig. 9. Examples of dominant teleconnection patterns deduced by varimax rotated EOFs of monthly 500 mb geopotential height anomalies for 45 years from 1946 to 1990.

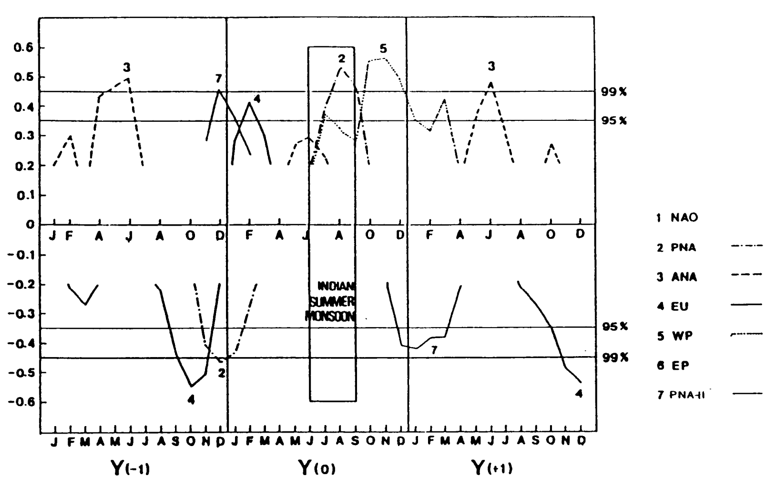

Fig. 10. Lag-correlations between the Indian monsoon rainfall anomaly and time coefficients of the principal teleconnection patterns shown in Fig. 9. Correlations with values less than 0.2 are not shown.

thogonal functions (EOFs) analysis to the monthly 500 mb height anomaly data for the period of 1946 to 1990 (Yasunari and Ueno, 1992). Figure 9 shows the spatial patterns of six dominant circulation patterns. This objective method could successfully deduce most of the well-known teleconnection patterns such as NAO, PNA, WP, EU (Wallace and Gutzler, 1981). Figure 10 shows the time sequence of lag correlation between the Indian monsoon rainfall index and the time coefficients of some dominant teleconnection patterns for each month before through after the reference monsoon year. It is noteworthy to state that the significant correlations (above the 95 % level) related to the Indian summer monsoon (or

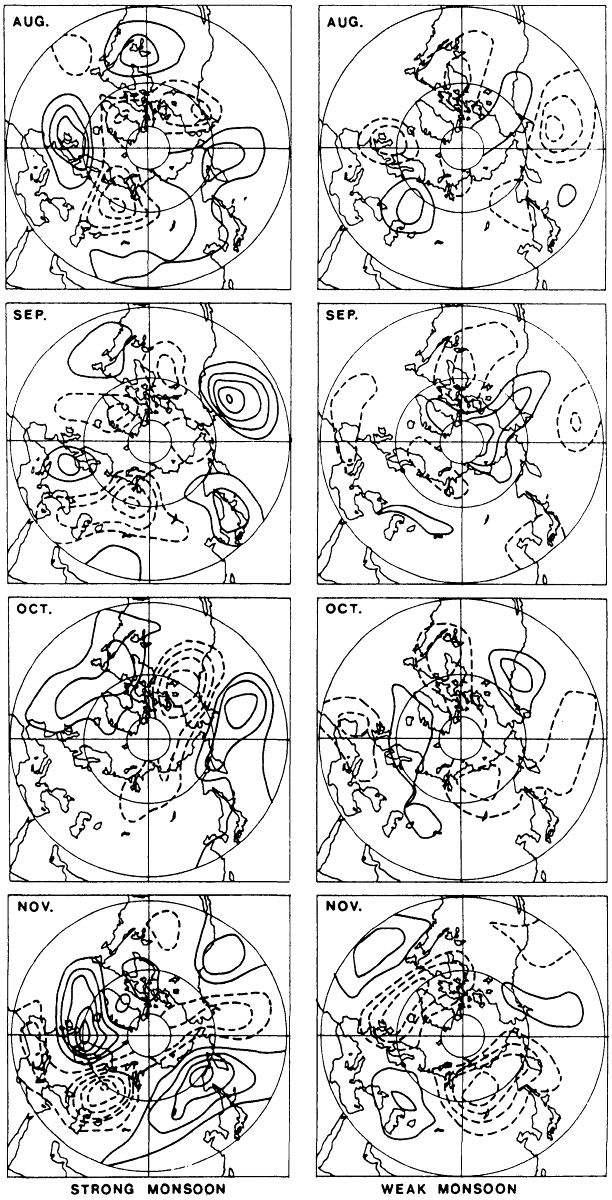

Fig. 11. Composite anomalies of 500 mb geopotential height from August to November of strong (left) and weak (right) Indian monsoon years. Contours are 10 gpm and negative values are shown with dashed lines.

the MAOS) persistently appear during and after the summer monsoon seasons (*i.e.*, autumn and winter) in the area of the north Pacific and the north American sector, as shown in the series of lag-correlations for the PNA, WP and PNA-II patterns. This strong seasonality of the persistent correlations seems to be consistent with the concept of the corresponding monsoon year in the topics. Other significant correlation peaks of the EU, PNA and PNA-II patterns appear in the previous autumn through early winter of Y(−1), which seems to be closely associated

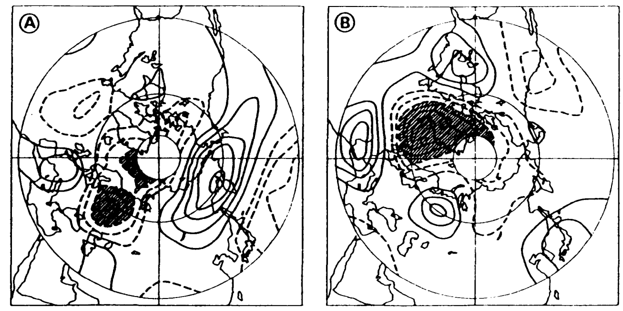

Fig. 12. Same as Fig. 11 except for December of (a) strong monsoon year and (b) weak monsoon year.

with the pattern for the snow cover anomaly over Eurasia, as shown in Fig. 7.

To find out the hemispheric anomalous circulation patterns associated with the anomalous state of the MAOS, the monthly 500 mb height anomalies are composited for each month during and after the weak and strong summer monsoon years, respectively, as shown in Fig. 11. As is readily deduced from this figure, in the weak monsoon years (Fig. 11 right), for example, the PNA pattern (of negative height anomaly over the north Pacific and the northeastern part of north America with positive anomaly over the Alaska/Rockies) is significant in August through October, implying an intensified Aleutian low and the meandering flow on the downstream side over the north American sector. In the strong monsoon years, in contrast, the reversed PNA pattern is dominant in the same period, implying the weakened Aleutian low with more zonal flow over this region. These (reversed) PNA patterns in this season with the intensified (or weakened) Aleutian low may not necessarily be due to the Rossby wave propagation mechanism, but may be more directly related to the weakened (or intensified) north Pacific high in the subtropics, as shown in Fig. 8. This weakened (or intensified) high may facilitate (or obstruct) the southward intrusion of the westerly flow over the north Pacific.

5. Feedback processes from the extratropics to the MAOS

Interestingly, the anomalous circulation with the (reversed) PNA pattern in late autumn seems to develop into the anomalous planetary-scale flow régime with wavenumber one plus two through the course of the seasonal evolution of the polar air mass in early winter (November and December), as shown in Figs. 11 and 12. That is, in the weak monsoon year, the PNA pattern over the north American sector tends to develop the strong seasonal trough over the northeast north America (and that over east Asia) while more zonal flow develops over the Eurasian continent. In the strong monsoon years, by con-

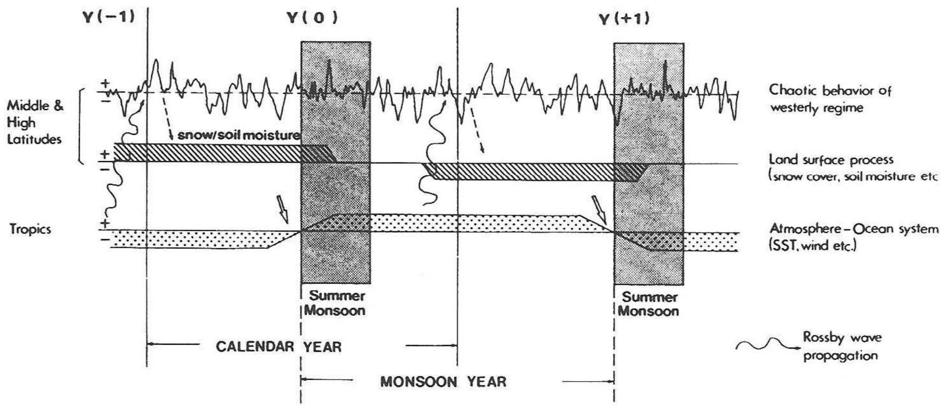

Fig. 13. Schematic diagram of the interannual variation of the climate system through the interactions between the MAOS in the tropics, and the westerly flow régime in the extratropics, by means of the Rossby-wave propagation and land-surface processes over Eurasian continent in the seasonal cycle.

trast, more zonal flow develops over the north Pacific through the north American sector, while an anomalous trough develops over the central part of the Eurasian continent. These anomalous flow régimes in winter may be due partly to the linear stationary Rossby-wave response to the persistent forcing of the anomalous MAOS, but also to the non-linear interaction between the changing basic-state flow and transient eddies through the seasonal march.

Now, it should be noted that this anomalous pattern in early winter (Fig. 12 left) is nearly identical to the pattern favorable for a large snow cover extent over central Asia, as shown in Fig. 7. Namely, the strong (or weak) summer monsoon condition tends to produce the weak (or strong) summer monsoon condition for the next year, through the anomalous westerly flow régime and the associated anomalous snow cover condition in the following winter. In other words, the biennial nature of the MAOS may owe its mechanism at least partly to the two-way interactions between the tropics and the extra-tropics in different seasons of the year, in which a negative feedback exists in the propagation of climatic signals.

The interactions of the climate system between the MAOS in the tropics and the extra-tropics, and between the seasons with a biennial oscillatory nature are summarized schematically in Fig. 13. One persistent anomaly in the MAOS from summer to winter seems to be responsible for the corresponding winter anomaly in the westerly flow régime in the northern extra-tropics, which provides the anomalous land-surface condition (*i.e.*, snow cover and possibly soil moisture) over the Eurasian continent. This anomalous land-surface condition, in turn, produces the opposite anomaly of the following northern summer monsoon (and the MAOS) to the previous year.

6. Role of the NAO on the MAOS and the ENSO

In reality, however, the year-to-year variabilities of the MAOS as well as the westerly flow régimes do not necessarily show the biennial nature exactly, as is seen, *e.g.*, in Fig. 5. The ENSO event (or the ENSO cycle), in fact, shows a preferred periodicity of 4 to 6 years, rather than 2 to 3 years (Rasmusson and Carpenter, 1982 *etc.*), though it seems to basically contain the nature of the biennial oscillation. We will further examine this problem.

Here, to focus on the difference of the circulation patterns between the ENSO years and the non-ENSO years among the weak summer monsoon years, the time coefficients of some dominant patterns (Fig. 9) are composited for three years centered by the weak monsoon year (Y(0)) with the ENSO event (a) and without the ENSO event (b) (Yasunari, 1988), as is reproduced here in Fig. 14. The SOI is also composited in the same manner and included in this diagram. It should be noted that the SOI apparently shows a biennial cycle in the period of Y(−1) through the beginning of Y(+1) particularly in the case of the ENSO. Associated with this feature, a remarkable difference is also seen in the SOI tendency from the winter of Y(−1)/Y(0) to the summer of Y(0) between the two cases.

To examine the teleconnection patterns associated with this remarkable feature, we focus on the difference of the dominant patterns appearing in the winter of Y(−1)/Y(0). The negative phase of PNA-II pattern, with negative height anomaly over the northeast north America and positive anomaly over the northeast Pacific seems to be a common feature in both cases. This may presumably be related to the strong or non-weak summer monsoon of Y(−1), as is suggested in Figs. 11 and 12. A remarkable difference between the two cases is the phase or polarity of the NAO pattern, as noticed

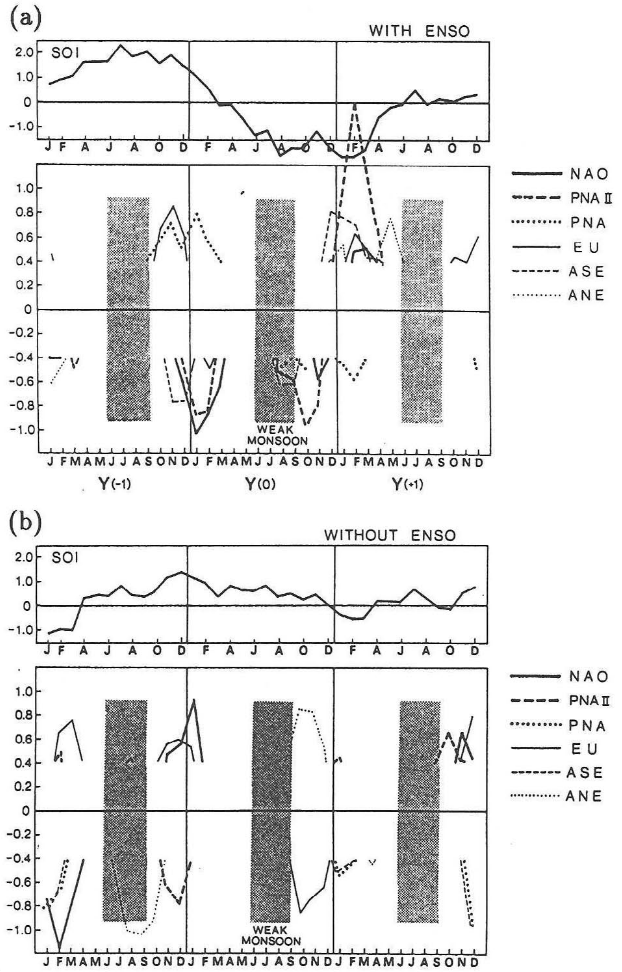

Fig. 14. Composite time coefficients of some dominant teleconnection patterns associated with (a) the ENSO years and (b) non-ENSO years among the years of weak Asian summer monsoon. Time sequence of composite SOI for each case is also plotted. (Yasunari, 1988)

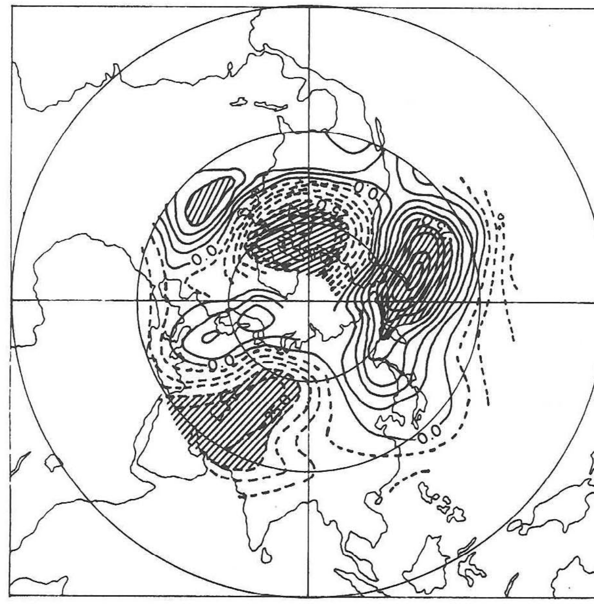

Fig. 15. Composite anomalies for 500 mb height (north of 20N) for the preceding late winter (January and February) of the ENSO event among the weak Indian monsoon years. The contour interval is 5 gpm and negative values are dashed. 5 % (or less) significant area is shaded.

in the sign of the time coefficients in the same period. In the case of the ENSO years, negative values are significant, which implies the phase of the NAO with north/negative and south/positive anomalies. In the case of the non-ENSO years, in contrast, large positive values are noticeable.

It is noteworthy to state that since Walker and Bliss (1932) the NAO has been noted as an oscillation independent from the Southern Oscillation (SO). A more sophisticated complex EOF analysis of the near-global sea level pressure date made by Barnett (1985) also shows the apparent orthogonality of the NAO to the SO. Certainly Fig. 10 shows no significant linear correlation between the Indian monsoon rainfall index and the NAO, though the NAO is one of the most dominant teleconnection patterns in the northern hemisphere. However, this does not necessarily imply that the NAO is not physically linked with the SO. Figure 14 suggests, on the contrary, that the polarity of the NAO, coupled with the negative phase of the PNA-II may play a crucial role in the occurrence or non-occurrence of the ENSO even later on. The seasonal 500 mb height anomaly composited for the late winter (Jan. to Feb.) of Y (0) with the ENSO event (Fig. 15) apparently shows a large positive anomaly over the Aleutian low area and the middle of the north Atlantic and large negative anomalies over the northern part of north America and central Asia. This pattern is, indeed, very similar to the typical anomalous pattern for large snow cover extent over central Asia (Fig. 7), except for the prominent NAO pattern over the north Atlantic and the more southward extent of negative anomaly toward the Arabian Sea. This anomalous pattern strongly suggests that in the winter preceding the ENSO event a more-zonally oriented strong jet is predominant over north America through the north Atlantic while a deep trough with a southward intrusion of the jet is located over central Asia.

A question may arise as to how this slight difference of the preceding winter circulation anomaly over the north Atlantic through the Eurasian continent triggers the ENSO event in later seasons or no, though both conditions are undoubtedly responsible for a weaker summer monsoon condition. One speculative explanation may be a difference of

Rossby-wave train path and associated storm tracks over central and south Asia, which may (or may not) facilitate the cyclogenesis and westerly wind burst over the equatorial Indian Ocean in late winter through the pre-monsoon season, possibly through the mechanism suggested by Hiskins *et al.* (1990). Namely, the NAO seems to play a dynamical role of modulating the intensity and distribution of the jet stream over the north Atlantic, which in turn changes the Rossby wave propagation characteristics on the downstream-side, *i.e.*, over Eurasia and the Indian Ocean.

We could hypothesize, in any event, that if only one anomalous state of the NAO with the polarity of negative/north and positive/south is additional in the preceding winter to the anomalous state of the MAOS with the weak Asian summer monsoon, a very favorable condition for triggering the ENSO event may be produced.

7. Concluding remarks

The role of the Asian monsoon on the global-scale climatic variability with the biennial and ENSO time scales is discussed, by examining the two-way interactions between the tropics and the extratropics in the seasonal cycle. In this process, the Asian monsoon functions as part of one climate system referred to as the MAOS (Monsoon and the coupled Atmosphere/Ocean System) over southeast Asia through the tropical Pacific.

The MAOS prominently shows a biennial oscillatory nature which tends to have one anomalous state starting in the northern summer monsoon season and persisting for about one year (*i.e.*, one "monsoon year") with the mature phase in winter (Yasunari, 1990a; 1991). This anomalous state of the MAOS is combined with the anomalous atmospheric circulation over the subtropics and the extratropics (*i.e.*, PNA or reversed PNA) of the north Pacific from late summer through winter, through the modulation of the subtropical high and the stationary Rossby wave propagation mechanism.

The anomalous circulation over the north Pacific thus produced in the early winter seems to evolve to the hemispheric anomalous circulation régime with wavenumber-one and/or-two structure in late winter. That is, the PNA pattern (negative height anomalies over the north Pacific and the northeast north America with positive anomaly over Alaska/Rockies), followed by the weak summer monsoon, tends to produce on anomalous cold trough over northeast north America with more zonal flow over Eurasia. By contrast, the reversed PNA pattern, followed by the active monsoon, tends to produce more zonal flow over the north Pacific through north Atlantic sector with an anomalous cold trough over Eurasia.

The anomalous circulation thus produced over

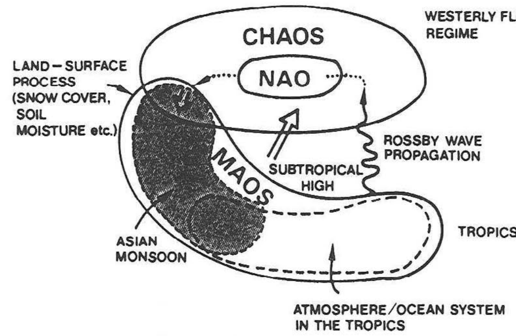

Fig. 16. Schematic diagram of the interactions between the MAOS and the chaotic westerly flow régime (the Chaos), similar to Fig. 13 but focusing more on the spatial structure. The role of the NAO is also specified.

Eurasia seems to provide a favorable condition for the extensive (or diminished) snow cover area and mass over central Asia, which in turn is responsible for the reversed anomalous state of the following Asian summer monsoon and the MAOS, presumably through the albedo and snow-hydrological effect of anomalous snow cover. That is, the biennial nature of the MAOS and the climate system in the northern hemisphere may be due, at least partly, to this two-way interaction between the MAOS in the tropics and the westerly flow régime in the extratropics in the seasonal cycle.

In addition, the significant polarity change of the NAO in the preceding winter depending on the ENSO or the non-ENSO event suggests that the NAO over the north Atlantic seems to play a role in the climatic chaos, by stochastically amplifying or damping the biennial nature of the MAOS. The timing of the ENSO event or the ENSO cycle may, therefore, be modulated by this interaction between the MAOS and NAO, both of which are basically independent of each other. The MAOS seems to behave more or less as a quasi-transitive or almost-intransitive climate system (Lorenz, 1968), while the NAO may represent the chaotic behavior of the westerly régime. The whole view of the MAOS, the chaos in the westerly régime and their interactions through the Rossby wave propagation and the land-surface processes (snow cover, soil moisture *etc.*) over the Eurasian continent is schematically shown in Fig. 16.

The mechanism of the interannual variability of the climate system discussed here should be compared to the recent model result by Lorenz (1990). He argued that the interannual variability of the climate system could result from an interaction between the chaotic winter circulation and the intransitive summer circulation but shows a totally chaotic behavior because of a randomly-renewed initial condition for summer by the chaotic winter condition. This conceptual model for the propagation of climatic signals in the seasonal cycle may be partly compatible with the observational evidence presented here. However, in reality, the anomalous state of the MAOS seems to give persistent anomalous forcing to the wésterly régime from the northern summer to the following winter, to produce a particular planetary flow pattern as a "climatic signal", though it is more or less disturbed by the chaotic nature of the westerly flow régime. This climatic signal in winter is, in turn, transmitted to the following summer, through the memory of land-surface (and possibly ocean-surface) processes during winter and spring. As a consequence, the variability of the real climate system seems to have one maximum power spectra in the biennial through the ENSO time scale, rather than the red or white noise spectra.

These results suggest strongly that the Asian monsoon is manifested as a global-scale land/atmosphere/ocean interaction between the largest continent and the largest ocean on the earth, and plays a key role as a transmitter of climate signals between the tropics and the extratropics in the seasonal cycle. There are, however, still many problems to be solved. Particularly, the interactive physical processes of the Asian monsoon with the coupled ocean/atmosphere system in the Pacific/Indian ocean sector and with the land-surface condition (*e.g.*, snow cover, soil moisture *etc.*) over the Eurasian continent may be a central issue to be intensively studied both observationally and theoretically. These problems should be undoubtedly one of the main targets for the ongoing TOGA and the forthcoming GEWEX.

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地球気候システムの年々変動におけるアジアモンスーンの役割

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ENSO の時間スケールの地球気候システムの年々変動に果たすアジアモンスーンの役割を、アジアの夏のモンスーン、熱帯太平洋の大気・海洋結合系及び中緯度の偏西風レジームの間の統計的、力学的な関係を調べることにより、考察した。

アジアモンスーンは、熱帯太平洋域での大気・海洋結合系と密接にリンクしており、モンスーン/大気・海洋結合系（略して、MAOS と仮称）とも呼べる一つのシステムをなしていることが明らかとなっている (Yasunari, 1990a)。この MAOS は、準２年周期の振動特性を持っており、ある偏差状態は、アジアの夏のモンスーン頃から始まり、約１年持続するという季節性を示す (Yasunari, 1991)。

この MAOS の偏差状態は、亜熱帯高気圧の強弱やロスビー波の伝播という機構を通して、北太平洋の亜熱帯・中緯度の夏から秋にかけての大気循環に、大きな影響を与えていることがわかった。即ち、モンスーンの弱い（強い）年には、（逆）PNA パターンが卓越する。そして、引き続く冬の半球スケールの偏西風循環は、この秋の大気循環の偏差が初期条件となったような波数１または２のパターンが卓越する。即ち、PNA パターンにより、北米東岸あるいは極東のトラフが発達し、ユーラシア大陸上はより帯状流的な流れのパターンとなる。反対に、逆 PNA パターンでは、北太平洋から北米域がより帯状流的となる一方、ユーラシア大陸上のトラフが発達しやすくなる。

ユーラシア大陸上のトラフの発達・未発達は、さらに、そこでの冬から春の積雪面積の偏差の形成とい

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う物理過程を通して、次の夏のアジアモンスーンの偏差に影響することが示された。即ち、MAOS と偏西風レジームが結合したこの気候システムでは、弱い（強い）夏のモンスーンの後の秋には、（逆）PNA パターンが持続し、続く冬にはユーラシア大陸上に少（多）雪をもたらす循環場が卓越することにより、次の夏のモンスーンは、強い（弱い）状態になるという、2 年振動的傾向を持つことがしめされた。このように、MAOS と中・高緯度の偏西風レジームを含む気候システムの準 2 年振動的変動の機構は、アジアモンスーンを媒介とした、熱帯と中・高緯度のあいだの、季節を違えた相互作用によることが強く示唆される。

さらに、現実のより非定常的なシステムの振る舞いと、ENSO のように上記の準 2 年振動が増幅された状態の物理的な理解には、ENSO とは全く独立な振動系として指摘されている北大西洋振動（NAO）の、このシステムへのストカスティックな強制が非常に重要であることを示唆する観測的事実も提示された。