Model Study on the Interannual Variability of Asian Winter Monsoon and Its Influence

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ABSTRACT

The interannual variation of Asian winter (NE) monsoon and its influence is studied using the long-term integration of Max-Plank Institute ECHAM3(T42 L19) model.

The simulation well reproduces the main features of the climatological mean Asian winter monsoon and shows pronounced difference of atmospheric circulation between strong and weak winter monsoon and for the consecutive seasons to follow. Most striking is the appearance and persistence of an anomalous cyclonic flow over the western Pacific and enhanced Walker circulation for strong winter monsoon in agreement with the observation. The contrast in summer rainfall patterns of both East China and India can also be discerned in the simulation.

Comparison of three sets of experiments with different SST shows that the forcing from the anomalies of global SST makes a major contribution to the interannual variability of Asian winter monsoon and, in particular, to the interseasonal persistence of the salient features of circulation. The SSTA over the tropical western Pacific also plays an important part of its own in modulating the Walker circulation and the extratropical flow patterns.

The apparent effect of strong NE monsoon is to enhance the convection over the tropical western Pacific. This effect, on the one hand, leads to a strengthening of SE trades to the east and extra westerly flow to the west, thus favorable to maintaining a specific pattern of SSTA. On the other hand, the thermal forcing associated with the SSTA acts to strengthen the extratropical flow pattern which is, in turn, conducive to stronger monsoon activity.

The result seems to suggest a certain self-sustained regime in the air-sea system, which is characterized by two related interactions, namely the air-sea and tropical-extratropical interactions with intermittent outburst of NE cold surge as linkage.

There is a connection between the strength of the Asian winter monsoon and the precipitation over China in the following summer. Links between these two variabilities are mainly through SST anomalies but snow over Asia is a contributing factor as well.

Key words: Winter monsoon, Interannual variability,Interseasonal connection, SST anomaly

I. INTRODUCTION

Monsoon circulation is an important component of global general circulation. The seasonal activity and interannual variation of both SW summer monsoon and NE winter monsoon are related to the variation or anomaly of the global general circulation and boundary conditions. There has been substantial progress in the study of summer monsoon. In addition to exploring the mechanism responsible for the year-to-year variation of both Indian and eastern Asian monsoon and the way of forecast, a prominent feature of recent studies is to focus on the active role played by Asian monsoon in the variation of global circulation, taking it as one of the most active components of the global climate system. For example, starting from anomaly of summer monsoon, Yasunari addressed how anomaly of summer monsoon could lead to a chain of events in the air–land–sea system and posed a concept of "monsoon year" (Yasunari and Seki, 1991). However, the research on the winter monsoon, which dominates a vast area of the extratropics and the tropics for nearly half a year, has been relatively limited. An abnormal strong winter monsoon does not only bring about disaster weather of strong wind and low temperature, but also links to the variation of global circulation system. The winter monsoon over the eastern Asia (hereafter, AWM for brevity) is the most active circulation of Northern Hemisphere winter. Strong northeasterly flows starting from the front edge of the Siberian High, periodically push towards lower latitudes to become a winter monsoon surge. It tends to cross the equator and merge into the summer monsoon of the Southern Hemisphere. It has been realized that a strong or weak winter monsoon surge could be a direct and major factor that influences the convergence of flow, hence the convection, over the tropical western Pacific, the area of maximum precipitation in the winter hemisphere (Lau and Chang, 1987; Chen et al., 1991). So, the activity of AWM could be considered as a medium or even a drive which gives an impetus to the interaction between tropics and extratropics, and between the two hemispheres.

Moreover, meteorologists in their long–term forecasting practice have noticed that the circulation pattern in preceding seasons to a flood or drought summer in the monsoon region bears different features. Recently, Liu et al. (1994) have shown a good correlation between the dominant flow regime in Northern winter and the rainfall of eastern China in next spring. Sun and Sun (1994) have found that a severe drought at the Meiyu region of eastern China, in general, is preceded by an active or strong winter monsoon. The reverse is also true for flooding summer. And the pronounced difference in large scale winter circulation can be found not only in extratropical but also in tropical areas.

From the above–mentioned interseasonal connection of the circulation, one may naturally think of the role that AWM may play in air–land–sea interaction. As a matter of fact, Namias, in the early seventies, suggested that a weakening of SE trades over the equatorial Pacific might lead to the occurrence of El Nino (Numias, 1976). More recent and widely cited models of ENSO formation (e.g. Wyrtke 1976; Philander, 1984), stressed the lead of the strengthening rather than weakening of equatorial trade wind. And Li et al.(1988) directly attributed the initiation of ENSO to the frequent outburst of cold air over the eastern Asia. In other wards, the anomaly AWM may also be a medium and trigger of air–sea interactions.

Numerical simulation can be a powerful approach towards understanding the variability and role of AWM. It may partly make up the lack of high quality data in monsoon research and help explore the physical mechanism behind empirical findings. There have been extensive studies on monsoon activity (WMO, 1992). And more recently, a number of numerical simulations suggested by MONEG have been carried out (e.g. Duemenil, 1995), focusing on the influence of SST and land surface process on the monsoon in the Northern Hemisphere. Substantial progress has also been made in the eastern Asian monsoon (e.g. Zeng et al., 1994). However, most of the past studies are confined to summer monsoon (Huang and Wu, 1987). Studies of the impact of SSTA over the tropical Pacific on the monsoon have been concentrated on the Indian monsoon.

Based on the long–term integration of ECHAM3 model of Max–Plank Institute of Meteorology, starting from different initial states with different boundary forcings, this paper will address the following three related problems.

(1)The capability of the model to simulate the mean feature and interannual variation of

AWM and the lagged relation of anomalous AWM with the circulation in following seasons. The result is expected to provide a test to the performance of the model and, in turn, a support to the empirical findings.

(2)The possible reason or factors that might lead to the observed interannual variation and lagged influence of AWM, in particular, the role of SST anomalies.

(3)The role of AWM in the interaction between tropical and extratropical circulation and between air and sea.

II. METHOD

Investigation in the following sections is based on the results from long-term simulation by ECHAM3 model. For the description of the model, the readers are referred to Roeckner et al.(1992).

Three sets of model experiments were used. The control experiment set with climatological SST consists of a 30-year integration. For brevity, it will be referred to as CLE. The second set of experiment includes five 14-year simulations starting from different Januaries of the CLE, but with the same observed global SST from 1979 to 1992. This accumulates to an ensemble of 70 year realizations of the model atmosphere. It will be called OGE. The third set of experiment adopted is the one where the observed SST was used at the central and eastern tropical Pacific. However, only one 14-year run has been carried out so far (hereafter called OPE).

In addition to the model simulation data, analysis data from ECMWF for 1979-1993 have been used.

As basic approaches, the correlation coefficient between AWM and different parameters was calculated for the same and lagged seasons, and composite charts made to characterize the evolution of strong or weak AWM events. It is expected that intercomparison between these groups of experiments could provide some insight to the role of AWM activity, and believed that the results could be more general and typical than those by one single case sensitivity study.

To distinguish strong AWM from weak AWM years and classify them into groups, some objective criteria are used. We have defined and compared several indices which describe the intensity of AWM from different aspects, such as the strength of NE surface wind at the coastal area of the eastern Asia, the surface temperature of Southeast China, or combination of these parameters. Careful examination of the results using the various indices shows that the basic features hence, the conclusion that may be reached are, by and large, similar. Consequently, only the results using the "v index" will be presented in the following sections, i.e. the mean v component of 1000 hPa wind velocity for the coastal eastern Asia (10°-30°N, 115°-130°E), including land and sea. This index will be referred to as VI in the following text.

III. THE RELATION BETWEEN AWM AND THE GLOBAL CIRCULATION

1. *Simulation of Winter Monsoon*

First we describe the ECHAM3 model's simulation of the winter monsoon mainly referring to the wind fields at lower level. The main wind field at 1000 hPa in the winter (DJF) averaged from 70 samples of OGE (with global observed SST) is given in Fig.1a. Compared to the analysis (Fig.1b) from ECMWF, the main features are in good agreement. The three branches of winter monsoon with strong northeasterly wind in the Eastern Hemisphere, located respectively over East Asia, the Indian peninsula and East Africa are all well reproduced. The simulated circulation systems related to the AWM are also successful. Among them are the anticyclone over the Asian continent, connected to the Siberia High, which is the major system characteristic of the cold air activity at high latitudes and the monsoon surge over the coastal area of the eastern Asia linked with the northerly wind in the east flank of the high, which is the main sign of the winter monsoon, the northerly winds across the equator joining the westerly summer monsoon in the Southern Hemisphere constructing the equatorial trough and convergence region there. These systems are all important for the winter monsoon activity and well simulated.

There is a realistic inter-annual variability of winter monsoon in the model. The standard deviation of the v index (VI) of the OGE run is 0.67 while the standard deviation of the analysis is 0.38. The standard deviation of the CLE run is 0.50 which is lower than that of the

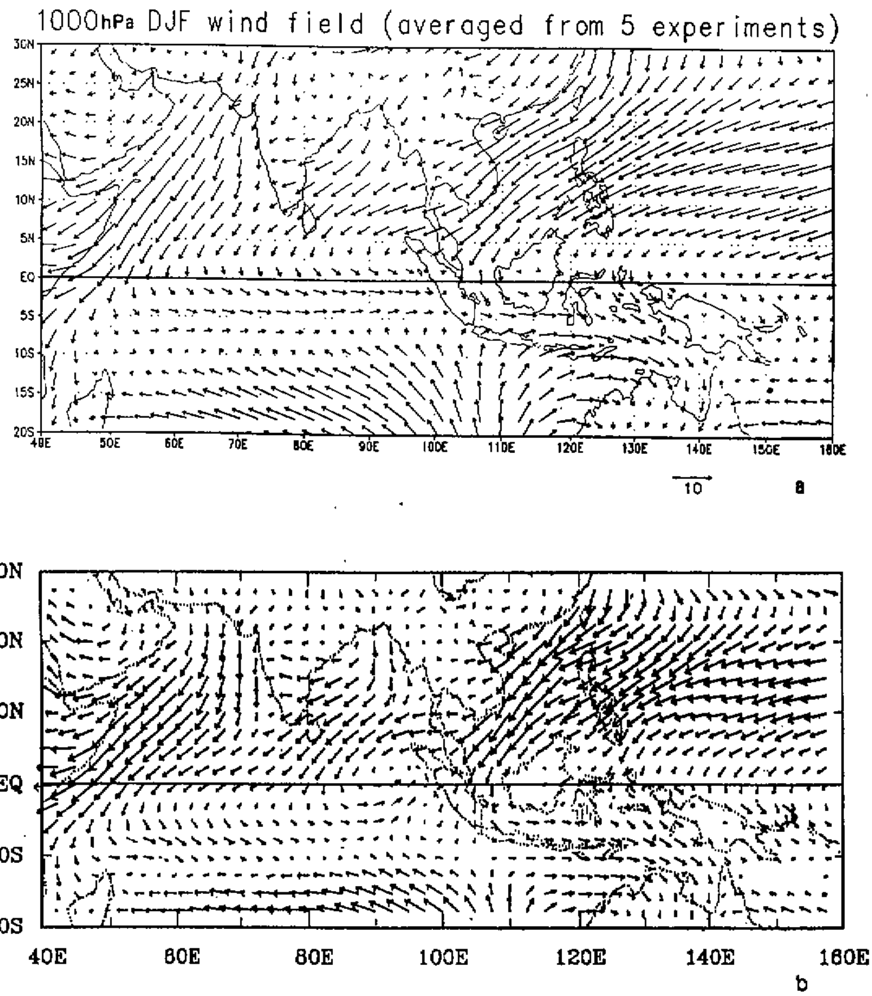


Fig.1. Mean 1000 hPa wind fields for winter(DJF). (a) averaged for five experiments of OGE (see text). (b) the analysis averaged for 1979–1991.

OGE (just statistically significant difference from OGE). The correlation coefficient between the mean of five OGE experiments and the single OGE experiment is 0.68, suggesting a very strong forcing from the observed SST on the low level wind of the coastal eastern Asia (VI). The correlation between the mean OGE run and the analysis for 1979–1993 is 0.59. This is reasonably high and suggests that the model gives realistic responses to the observed SST. The high standard deviation of the CLE run (higher than that of the analysis) suggests however that forcings other than the SST and internal variability are large as well in this area.

The good model reproducibility of AWM, which will be illustrated further in the following, provides a sound basis to investigate the relation between AWM and global circulation.

2. *Connection between Anomalies of AWM and Global Atmospheric Circulation Systems*

(1) Anomaly AWM and global wind fields

Using the normalized VI of the 70 samples of OGE, 9 strong and 9 weak AWM samples can be identified, the criterion being a departure of one standard deviation from the averaged. Making the composite of the two groups, we get the mean difference wind field for the two as shown in Fig.2. The implication of these difference wind fields is further studied by calculating the zero–lag correlation coefficient between VI and other variables. Figure 3 shows three of them, i.e. with precipitation, surface pressure and 500 hPa geopotential height fields. The minimum correlation coefficient is 0.24 to be significant at the level 0.05 and 0.31 at the level 0.01. Therefore, most of the major centres in Fig.3 meet the requirement of significance.

As can be seen in Fig.2, over the Asian continent there is a pronounced anticyclonic circulation associated with a cold strong high dominating that area typical of strong winter monsoon year. Another area of marked wind difference is located over the central North Pacific. Cyclonic flow appears to the west of the winter mean position of the Aleutian Low. This means the Aleutian Low shifts to the southwest in strong AWM situation. Over the tropical western Pacific, a cyclonic flow with convergence of wind can be seen. To the south, there are westerlies which is characteristic of the strengthening of summer westerly monsoon in the Southern Hemisphere caused by AWM, except positioning slightly to the north of the observed. The extra strong convergence at low levels brings more convective activity and precipitation.

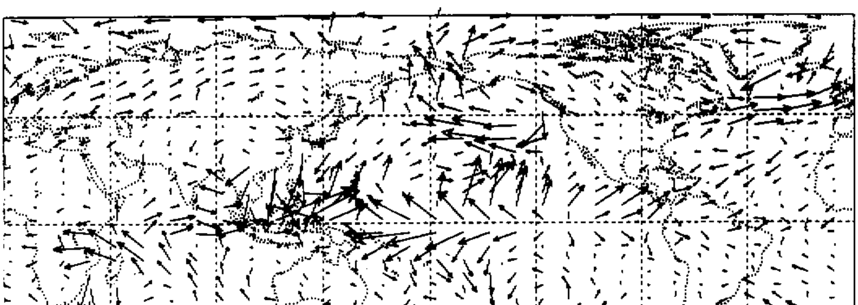


Fig.2. Composite difference of 1000 hPa wind between strong and weak AWM for winter.

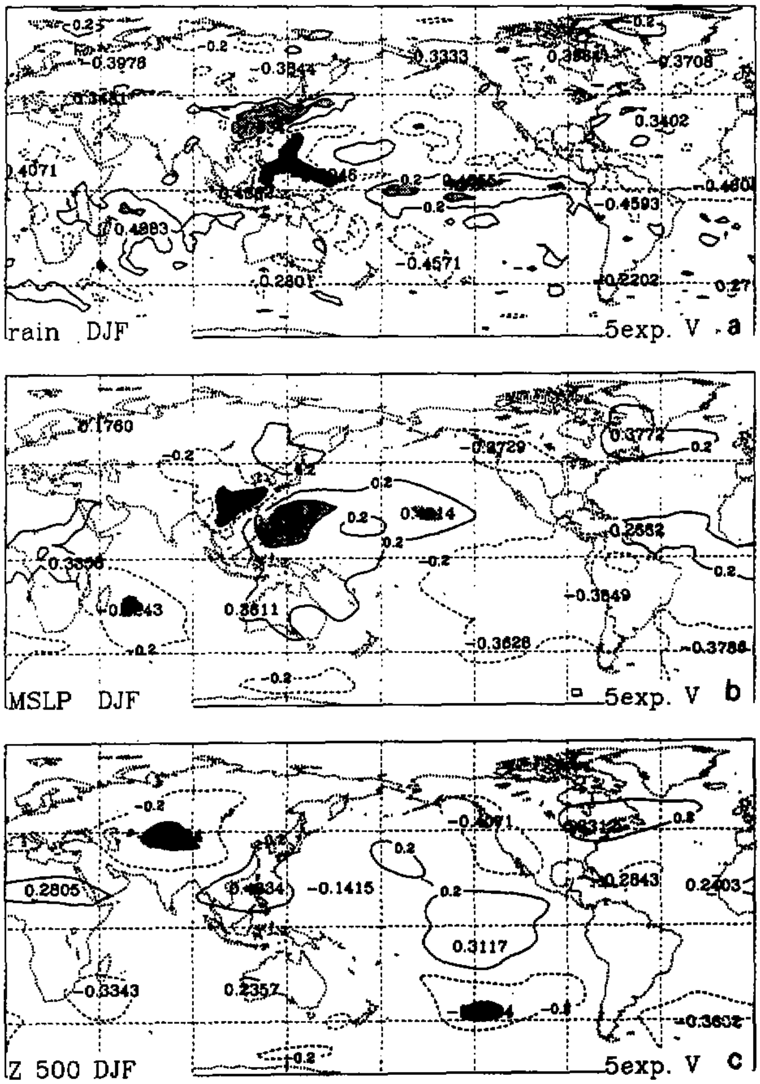


Fig.3. Distributions of zero-lag correlation coefficient in winter (DJF) for (a) VI and precipitation, (b) VI and surface pressure and (c) VI and 500 hPa geopotential height.

As shown in Fig.3a, correlation with precipitation, there is an area of negative correlation over the maritime continent, the extreme reaching -0.60 as much, which means an increase of precipitation with strong AWM ($VI < 0$). In contrast, there is a decrease of precipitation in the area of positive correlation over the eastern Asian continent where extra northerly wind dominated. The equatorial central Pacific is an area of positive correlation, i.e., a decrease of precipitation related to the divergence of differential wind field.

The dominant feature in the MSLP is a dipole near the China Sea which can be explained by the geostrophic relation. One can find further that a positive correlation over the western Pacific is connected with a negative correlation over the eastern Pacific. Negative correlation coefficients in the MSLP over the Asian continent mean that there is a pronounced anticyclonic circulation associated with northerly winds along the coast of China, typical of strong winter monsoon years (see Fig.3b).

(2) 200 hPa χ and vertical circulation

The differential velocity potential at 200 hPa between strong and weak AWM years is also calculated, which characterizes large scale vertical circulation and the position and intensity of convective heating (Fig. 4). For validation the corresponding fields from the analysis are shown as well (Fig. 5). The model captures the main features of χ fields at high level well. Both the simulation and analysis clearly indicate that in strong AWM years the centre of the large scale χ field is located slightly to the NW with a stronger extreme. The difference is mainly concentrated over the tropical western Pacific, the difference in the analysis being even stronger. These findings agree with the correlation maps of VI with precipitation discussed above.

The strong upper divergence area coincides with the area of lower convergence and strong precipitation as can be expected. These results suggest that strong winter monsoon processes cause not only the strong low level convergence and convections over the tropical western Pacific, but also the strengthening and northward shift of large scale upper divergence field, which, no doubt, would lead to the change of the local Hadley cell and Walker circulation. Fig.6 shows the composite N–S cross–section of vertical velocity omega for strong and weak monsoon year respectively. The difference is remarkable. For weak monsoon year, the ascending branch of Hadley cell located at the southern equatorial area is largely similar to the climatological mean for winter. However, for strong AWM year, the ascending branch apparently northward displaces with a second centre of ascending due to enhanced convection. The zero line shifts from 12°N for weak AWM to 18°N. Stronger and well organized descending motion also appears at mid–latitudes, which features the outburst and southward intrusion of cold air.

Over the eastern equatorial Pacific (Figs. 4 and 5) there is a positive χ area, which is consistent with the low level divergence of southeast trade wind (Fig. 2). The upper equatorial westerlies at 200 hPa is also clearly strengthened (not shown). This indicates that the Walker circulation over tropical area is greatly enhanced for strong AWM situation, in good agreement with observations.

(3) Difference in extratropical flow pattern

The strength of winter monsoon over the eastern Asia is directly related to the outburst of cold waves at extratropics. It is well established that when a meridional flow pattern is dominating Eurasian region, with the decay and eastward moving of blocking high over Ural area, there will be an intrusion of strong cold wave in the eastern Asia. Therefore, the strength of meridional flow and its variation directly leads to the outburst of cold air.

Fig.3c is the correlation between the VI and 500 hPa geopotential height field where positive value means a negative height departure under strong AWM (VI$<$0) situation. So there would be a stronger long–wave ridge to the southeast of Ural area and a deeper and southward stretching major trough over the eastern Asia, leading to an enhanced meridional flow pattern over that area. The simulation well features a typical flow pattern for outburst of winter monsoon. This feature of extratropical circulation on seasonal time scale is an integral part of air–sea interaction. We will come back to this point in Section four.

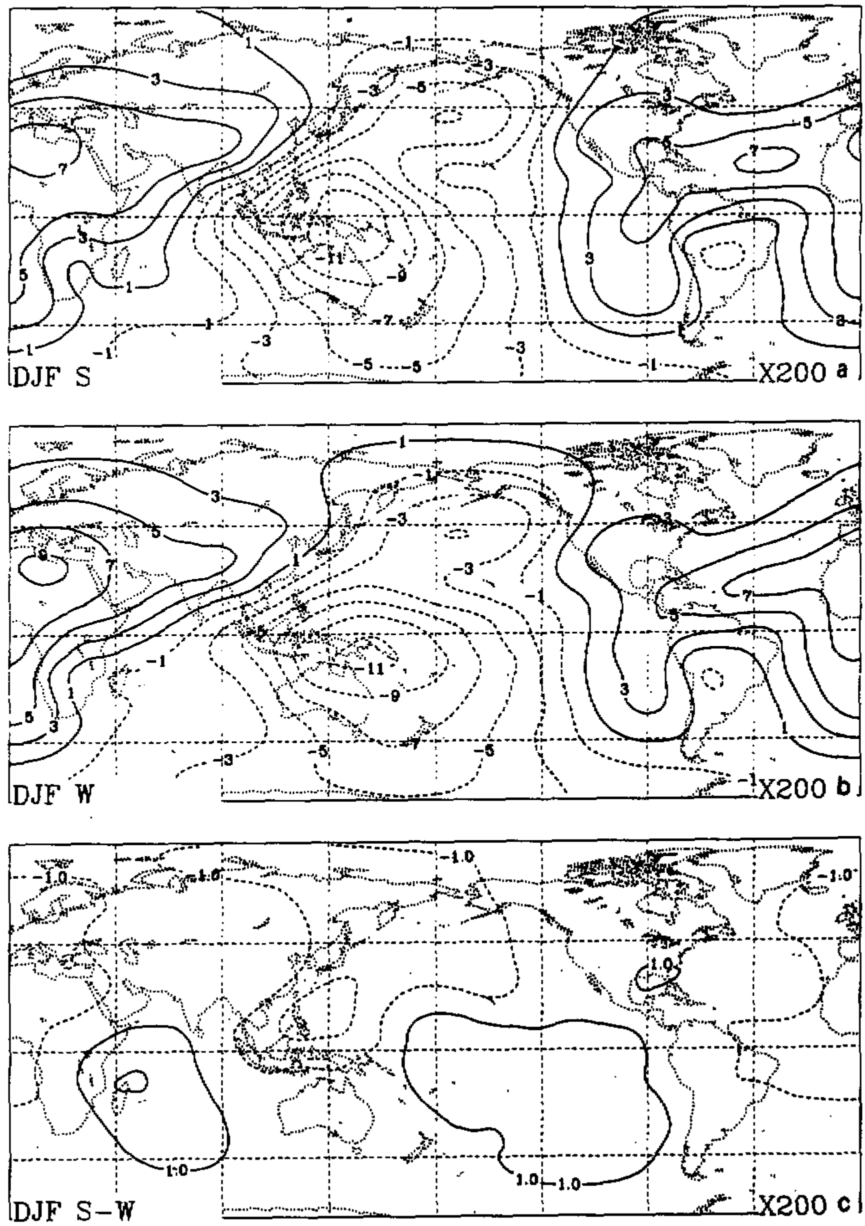


Fig.4. Composite of velocity potential at 200 hPa in winter (DJF). (a) strong AWM years (S), (b) weak AWM years (W) and (c) difference of (S) and (W). Contour interval: $5 \times 10^6 m^2 / s$.

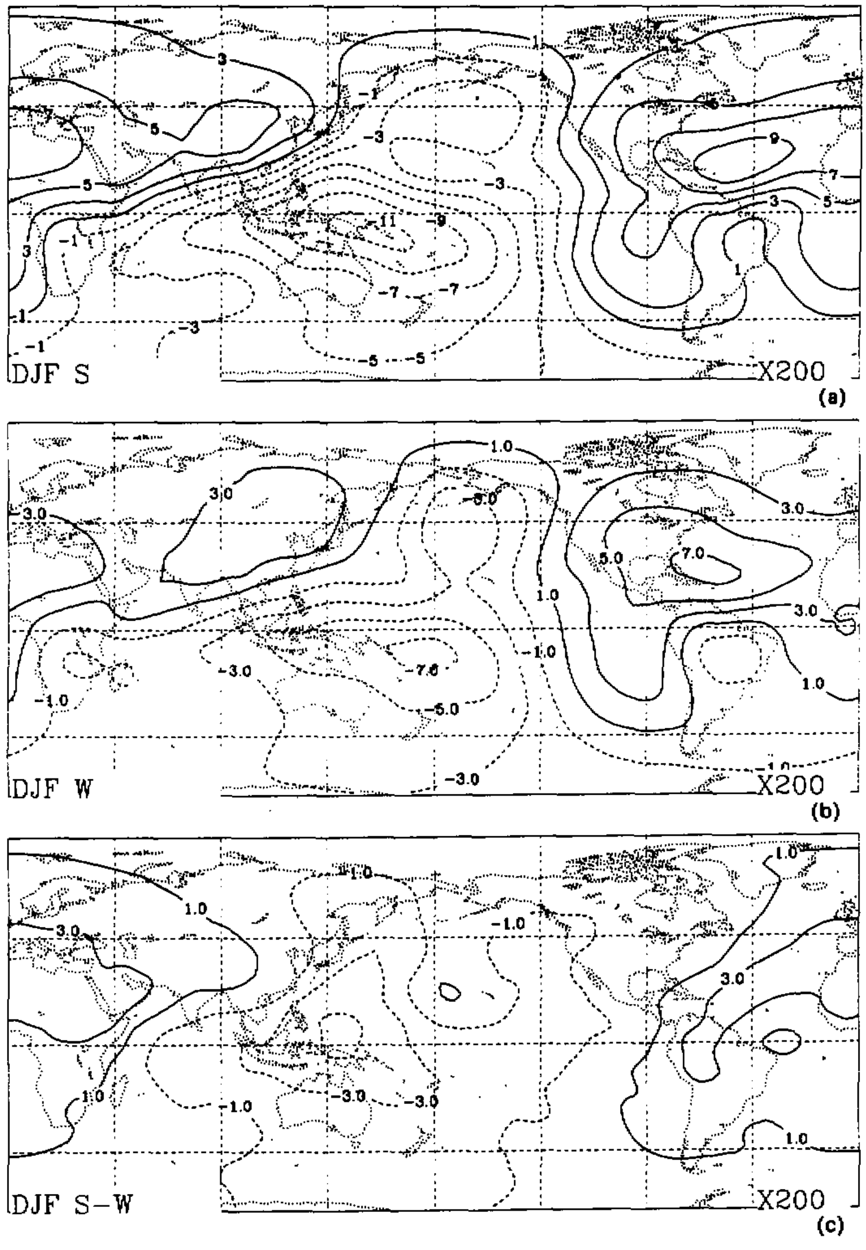


Fig.5. Same as Fig. 4 except for the analysis.

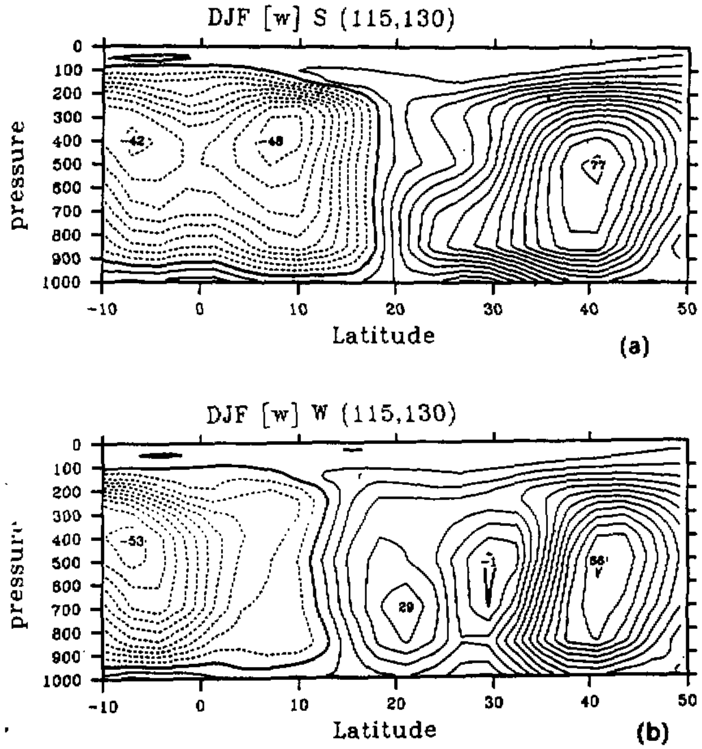


Fig.6. Composite N–S cross-section of vertical velocity omega for (a) strong and (b) weak AWM year. Contour interval: 5×10^{-5}hPa / s.

3. *Connection between AWM and Circulation of Following Seasons*

Observational studies have shown that anomalous AWM presents a good connection with the anomalies of following summer. In general, summer floodings in the central region of the eastern China correspond mostly to preceding weak AWM, drought years mostly are preceded by strong AWM (Sun and Sun, 1994).

As has been shown, the ECHAM3 model demonstrates a reasonably good performance in simulating the Indian summer monsoon rainfall, while the simulated rainfall for the eastern Asia, in particular, for the eastern China, is much less than the observed (Duemenil, 1995). Keeping this in mind we have compared the difference of simulated summer rainfall following the strong or weak AWM. Fig.7a shows the composite difference of total rainfall for summer (JJA). A band of negative value is ranging from the central eastern Asian continent to the southern Japan, largely corresponding to the so called "Meiyu" belt. This implies that in a strong AWM year, deficient summer rainfall in the "Meiyu" area is likely, and vise versa.

This relation has also been shown in the lagged correlation maps between the AWM and 500 hPa geopotential height (not shown). In June and July, there are positive correlation over Ural mountains region and negative over the eastern Asian continent, implying a stronger ridge over Ural and deeper trough over the eastern Asia for weak AWM year (VI > 0). This is just the typical Meiyu pattern of wet year. This result is in line with the observational study, the finding of Sun and Sun (1994).

It is more interesting that a pattern of +,−,+ is presented in Fig.7a ranging from the south to the north of the eastern China, which implies that an excess of rainfall at the Meiyu region is likely to be associated with a drought both to the south (South China) and to the north (North China). This delicate feature also coincides well with established empirical findings.

From Fig.7a, the connection between Indian monsoon rainfall and AWM can also be seen. In contrast to the Meiyu area, there is basically an area of positive departure around Indian Peninsula, which means a strong AWM would indicate an excessive Indian monsoon rainfall in the following summer. A number of earlier studies have shown that there is strong summer monsoon and more rainfall in India, when the Meiyu tends to be less in the eastern Asia (Chen et al., 1991). This connection between the summer rainfall of the two regions is well simulated by the model.

Fig.7b, the distribution of correlation between AWM index and summer precipitation also well illustrates the above rain pattern and connections.

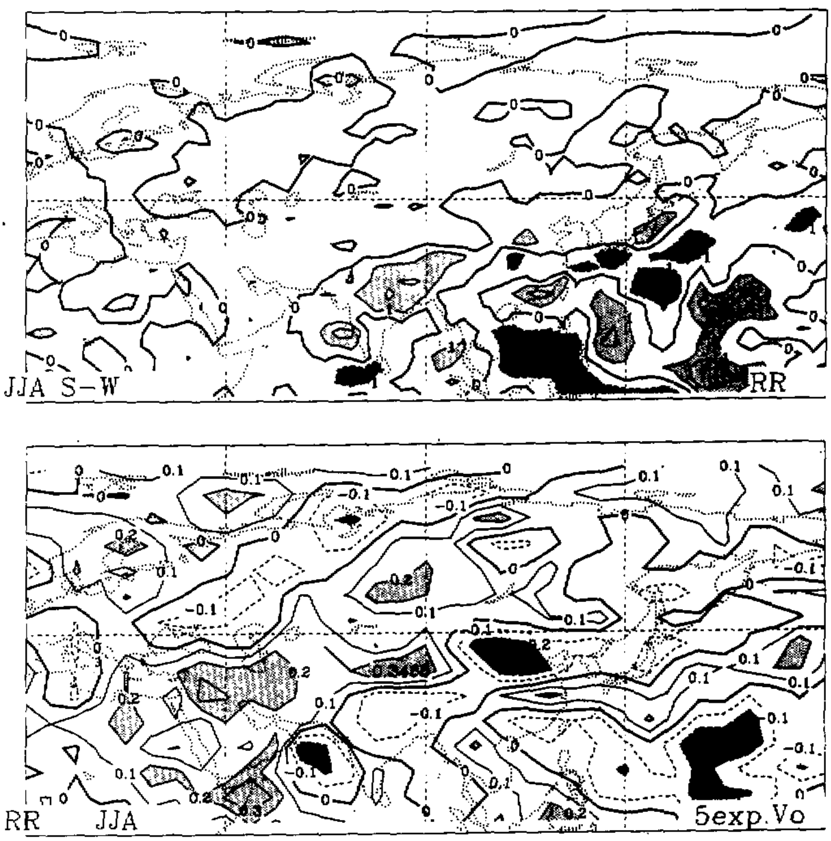


Fig.7. (a) Composite difference of summer (JJA) rainfall (mm / d) and (b) distribution of correlation between VI and summer (JJA) rainfall.

The difference wind field (S−W) for three consecutive seasons is shown in Figs. 8 a,b,c. Since the vertical extension of summer westerly monsoon is higher than that of winter northerly flow, here the wind field at 850 hPa has been adopted. As can be seen in the following summer, there is a belt of SW wind stretching from the Arabian Sea to the Indian Peninsula. An anticyclone is located right over the area of Mascarene High and the associated southeasterly winds cross the equator, turning into southerly flow along East Africa. Therefore, it seems that related to a strong AWM year, not only the Indian summer monsoon, but also those monsoon systems appear to be stronger. Also noticeable is the extra cyclonic flow over the subtropical western Pacific which is a typical anomalous circulation leading to dry Meiyu as identified by Song et al. (1993).

The above simulated results demonstrate that there may exist certain kind of temporal teleconnection of circulation on a seasonal time scale. To further explore the lag effect of AWM, we examine the seasonal evolution of circulations. In Fig.8c, the conspicuous features in summer, as stated above, are enhanced cyclonic convergence over the western tropical Pacific and the anticyclonic circulation over the northern Pacific. These features are clearly present in spring and even in summer.

The persistence also can be seen at upper large−scale divergence field. The differential velocity potential at 200 hPa for the consecutive seasons is displayed in Fig.9. The persistence of large−scale upper χ field over the tropical Pacific is apparent, indicating that the enhanced Walker circulation and local Hadley cell ssociated with strong AWM might maintain beyond the winter, even up to the summer.

The anomalies and persistence of large scale flow must be influenced by, and exert impacts on the state of the oceans and land surface. This will be discussed in the next section.

IV. THE ROLE OF WINTER MONSOON IN AIR−SEA INTERACTION

1. *Relationship between AWM and SST*

The variation of global atmospheric circulation on a seasonal time scale associated with winter monsoon anomaly should presumably be accompanied with SST change in a wide region. In order to investigate the interaction involved, a set of SST difference between the strong and weak AWM years is given (Fig.10), which starts from the preceding autumn.

We examine the winter situation first (Fig.10b). There is a negative SST departure belt with an extreme of −0.5°C along the southeast coast of Asian continent, roughly covering the Kuroshio current region. This negative belt maintains through the spring and summer (Figs.10 c and d), but is not very clear in the preceding autumn (Fig.10 a). One may consider that the drop of SST at that area might be an outcome of cold surge. Statistical studies indicate that there exists a positive correlation between the SST in Kuroshio area in winter and the summer precipitation in East China. So a weak winter monsoon may bring about more rainfall in that area consistent with observational results.

Another marked negative region is around the southern tropical and subtropical part of the Indian Ocean. The value of the centre near the Australia Continent also reaches −0.5°C. In the map for the preceding autumn, we find that there are only weak and sporadic negative areas and even positive ones at the eastern part of the Indian Ocean. It indicates that the appearance of negative region over there does accompany the strong AWM, which maintains until the coming summer (see Fig. 10d).

The most striking feature is over the tropical central and eastern Pacific. There is a wide and strong negative belt strengthening from preceding autumn to winter and persisting to the

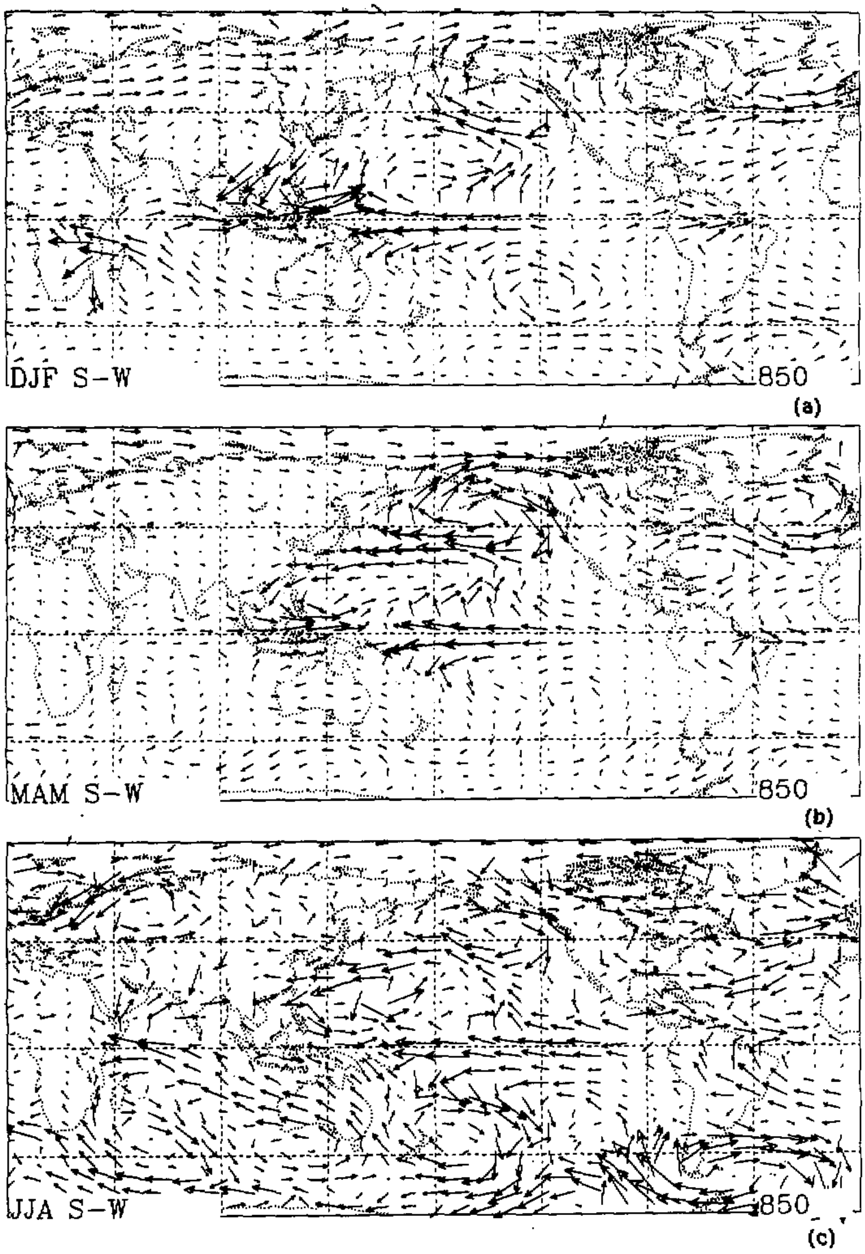


Fig.8. Difference of wind (S−W) at 850 hPa for (a) winter (DJF), (b) spring (MAM) and (c) summer (JJA).

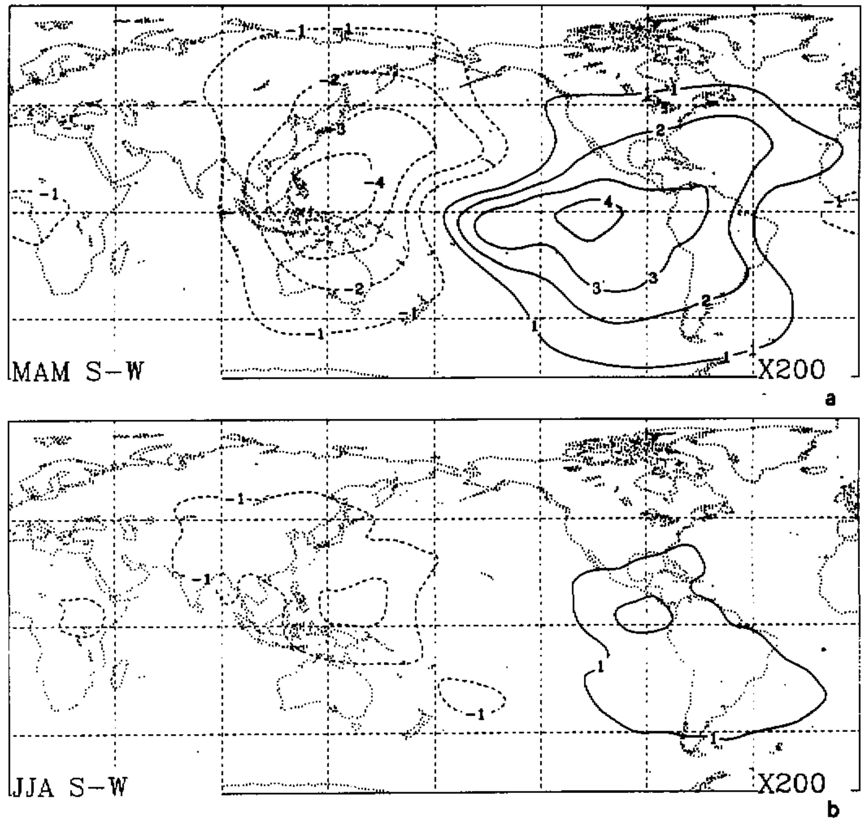
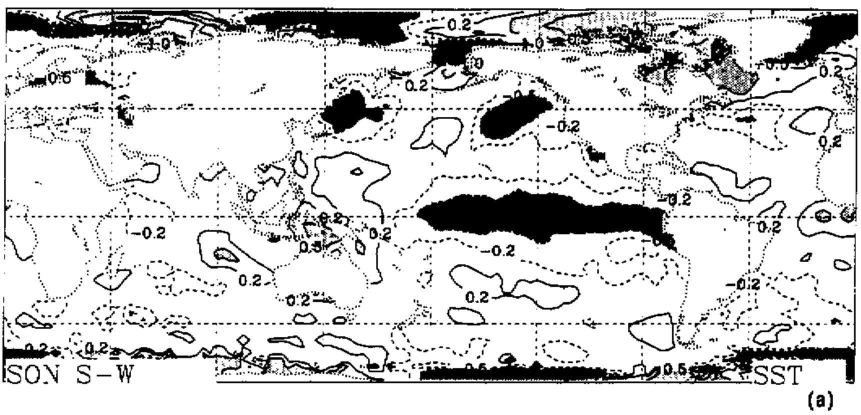


Fig.9. Difference of velocity potential at 200 hPa for (a) spring (MAM), (b) summer (JJA). Contour interval: $1 \times 10^6 m^2 / s$.



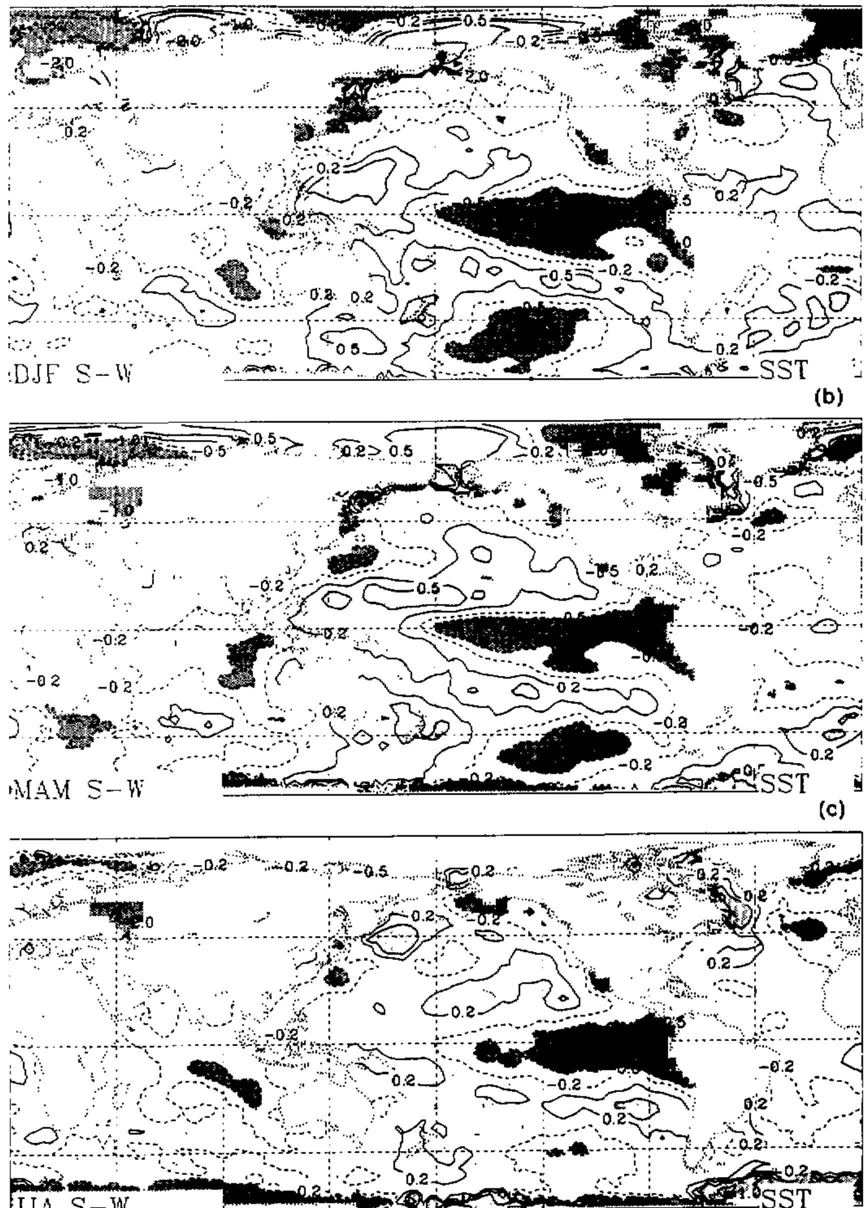


Fig.10. SST difference between strong (S) and weak (W) AWM years for (a) preceding autumn (SON), (b) winter (DJF), (c) spring (MAM) and (d) summer (JJA).

summer. A positive centre appears over the western Pacific, i.e. warm pool area, which is strengthened from autumn to winter, and moves northeastward in summer. Obviously the warmer sea surface temperature here is linked to the development of the equatorial trough and the convergence caused by stronger AWM as described in previous section. This positive area extends eastward. It appears at the date–line in spring and to the east of date–line in summer. It should be pointed out that the total wind field over that area is easterly though the departure being westerly, so the eastward expansion of SSTA cannot simply be explained by advection. The belt of cooler SST over the eastern Pacific persists consistently to the summer with the intensity slightly reduced.

These results clearly display a very close relationship between the activity of AWM and the SST at several main tropical regions. The SST distribution for stronger or weaker AWM years is different. More important is that this kind of difference would maintain or even develop in the seasons to come.

2. *Influence of SST on AWM*

To further explore the influence of SST, the results from three sets of experiments with different boundary forcing are compared., namely with climatological SST (CLE), observed global SST (OGE) and with observed SST only over the central and eastern tropical Pacific (OPE), respectively.

Fig.11 shows the interannual variation of VI of the three sets and from analysis. Only one 14–year simulation with observed SST over the tropical Pacific (OPE) has been carried out while there are 5 experiments with observed SST everywhere (OGE) and for the later also the ensemble mean is shown and copied into the middle panel for comparison with OPE results. The simulation with climatological SST was run over 29 years and for convenience we have divided it into two runs in one plot for the sake of using the same time axis. Apparently, there exists a dramatic variation of AWM for all these three sets. It is not surprising that the AWM of the model atmosphere presents clear year–to–year variation even without anomaly SST. However, the amplitudes of the fluctuation are different. We have calculated the standard deviation of the VI for 70 samples of OGE, 29 samples of CLE and 14 years of the analysis, being σ(OGE) = 0.67, σ(OPE) = 0.67, σ(CLE) = 0.50 and σ(ana) = 0.38 respectively, which implies that the difference of variance for the two sets of experiments using observed SST and climatological SST is significant at a 0.05 level. This suggests that the impact from the SSTA is significant to the AWM activities, and is likely to enhance the amplitude of interannual variation. Since we are dealing with AWM on a seasonal time scale, we cannot tell at this stage whether the global SSTA acts to enhance the strength of individual cold waves or increase the frequency of its occurrence. The comparison with the analysis shows that the model is overestimating the variability.

The comparison between strong and weak AWM cases for CLE run has also been carried out by composite charts of V at 1000 hPa, 200 hPa χ etc. (not shown). One can also find a cyclonic flow to the east of the Philippines and easterly flow along the equator further east on the 1000 (or 850) hPa differential wind field of DJF, similar to Fig. 8 for OGE run, though weaker. The enhancement of convection in that area can also be discerned in 200 hPa χ map for DJF. But no persistence of these features can be found in the following spring (MAM). This indicates an active role that AWM can play in the tropical air–sea interaction.

Now we will turn our attention to OPE (only the observed SST at the eastern Pacific is used). Comparisons between OPE and OGE (with observed global SST) will help clarify the

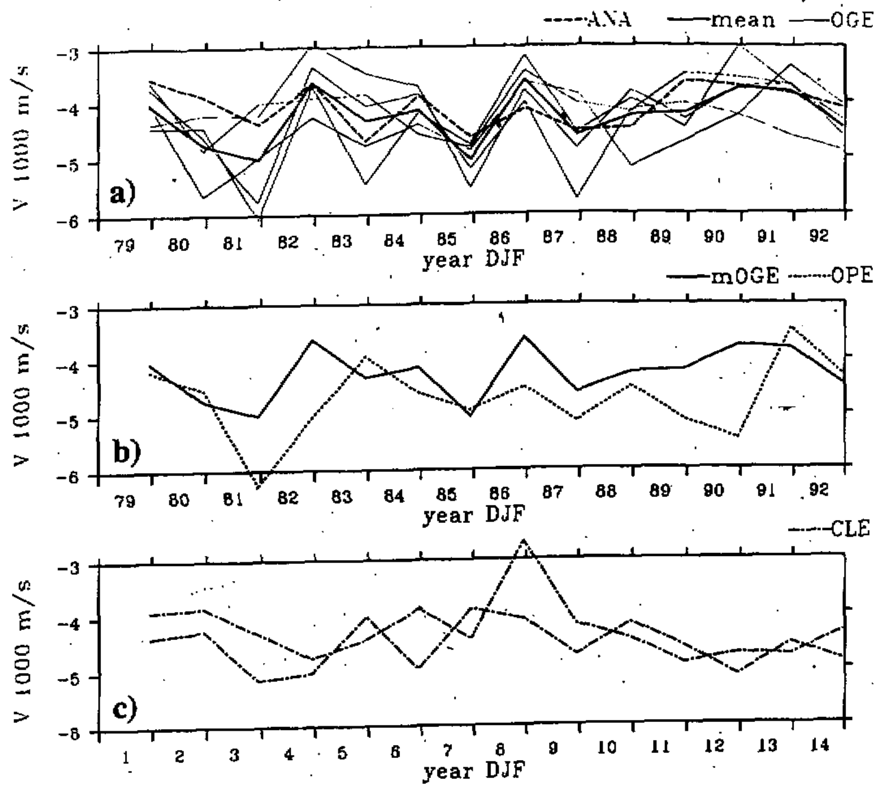


Fig. 11. Interannual variation of the 1000 hPa v component in the area (10°–30°N, 115°–130°E) during DJF for three sets of experiments. Solid line: OGE (thin for single experiments and heavy for the mean of 5 experiments), dashed line: analysis, dotted line: OPE, and dashed-dotted line: CLE. In the graph for CLE the whole experiment has been split into two so that the same time axis increment can be used as in the other graphs.

role played by SSTA of the eastern Pacific and other oceanic area, mainly the SSTA in the western Pacific. As can be seen in Fig.11, the variation of VI in the OPE and OGE runs is very similar until 1986 / 87. After that the OPE run continues to have large interannual variations while the variability in the OGE simulations stays small as in the analysis data. This indicates a major role of the SST over the eastern Pacific but leaves the possibility of other important influences.

Fig.12 shows the difference of composite 500 hPa geopotential height field for the two sets which is constructed by the same strong (or weak) monsoon years. It clearly shows that large scale features of the two are similar, which implies that the forcing of SSTA from other sea domain only acts to enhance the amplitude of the extratropical flow pattern. However, by a careful inspection, one may find that the pattern of OGE shifts to the west compared with OPE run, which is likely linked to the forcing over the western Pacific. As already noted in Section 3, the westward shift of the major trough over the eastern Asia is one of the prominent features associated with strong AWM year. The difference of composite precipitation during JJA following strong or weak AWMs (not shown) gives larger amplitudes than OGE runs with a shifted pattern.

The difference between OPE and OGE can be attributed to the ensemble impact of the SST over other ocean areas than the eastern Pacific. However if we notice the most pronounced thermal forcing, particularly, the anomaly heating (see Fig.7 of precipitation), we may well say that the difference of the two experiments mainly comes from the contribution of the western Pacific forcing, in turn linked to strong AWM.

The snow cover over the Eurasian continent might be another medium of the lag connection of AWM with the following seasons. Many studies have been focused on this problem, both by observational approach (e.g. Hahn and Shukla, 1976) and model simulation (e.g. Barnett et al., 1989; Vernekar and Zhou, 1995; Duemenil et al., 1995). Most of the early studies reached a same conclusion that an excessive Eurasian snow cover leads to a weaker Indian monsoon and less monsoon rainfall, while some of the studies show different results (Zwiers, 1993). There have been a number of investigations dealing specifically with the relation between the snow cover over the Tibetan Plateau and eastern Asian monsoon rainfall (Yang,1994). However, due to the lack of adequate data and probably the complexity of the problem, the conclusion so far has not been definite.

Now, we examine the relation of winter monsoon and Eurasian snow cover in the simulation. Fig.13 shows the composite difference between strong and weak AWM for OGE run. As expected, there would be more snow cover over the Eurasian continent, the features being largely similar for both winter and spring. Also noticeable is the vast area of excessive snow cover in the north of North America for strong AWM. However, as can be seen in Fig.13, the positive area is mainly located over the northern Eurasian continent, roughly to the north of 55°N, while there is basically negative area southward. It should be pointed out that this feature is also apparent in the composite difference between strong and weak AWM for CLE run, i.e. without SST anomalies. We have noticed that Hahn and Shukla (1976) adopted the snow cover south of 52°N in their study and most of the numerical studies treat the Eurasian snow as a whole (e.g. Vernekar and Zhou, 1995). So the result shown in Fig.13 is not inconsistent with their findings. A strong AWM may be related to a low snow pack over South Siberia and the Tibetan Plateau as well, which, in turn, links to a stronger and earlier arrival of Asian summer monsoon. It seems therefore that the snow cover associated with AWM, as a contributing factor, can also carry information from winter to summer.

Vernekar (1995) pointed out that an excess of Eurasian snow cover may lead to a weakening of trade wind over the equatorial eastern Pacific. This supports interactions between tropics and extratropics and the air and sea as suggested in next section.

3. *Role of AWM in Air–sea Interaction*

Since we are dealing with a SST forced GCM rather than a coupled model, the following discussion is more of reasoning than demonstration. Based on the results in previous sections, the role of AWM in the operation of air–sea interaction can be synthesized as follows. An anomaly strong AWM may lead to an extra cyclonic flow and enhanced convection over the warm pool area, in particular, to the east of the Philippines. From vorticity equation, one may deduce that this is due to the combined effect of $-\beta v$ and ensuing convergence ($-\zeta D$) in the tropics. In turn, it tends to strengthen the Walker circulation as a whole to the east of the region, and to the west , to simulate a more regional anti–Walker cell. This feature, as already shown in both the observed and model simulations, could present and persist through winter and following seasons.

This consequence may bring about other two effects. The first, as noted above, is that the

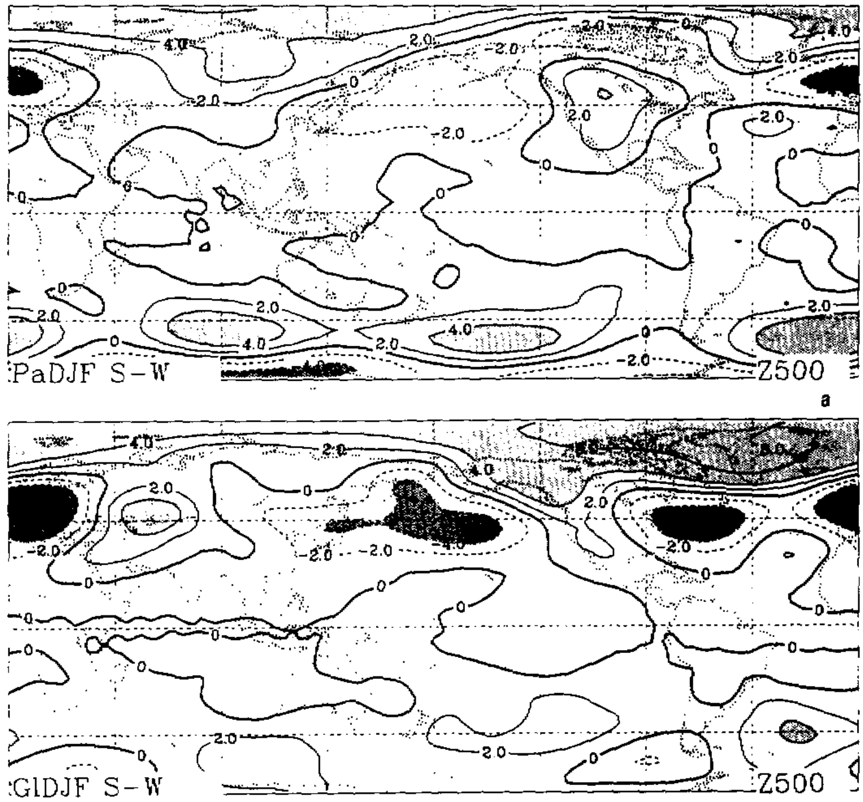


Fig.12. Composite difference of 500 hPa geopotential height (dam) between strong (S) and weak (W) AWM year (DJF) for (a) OPE run and (b) OGE run.

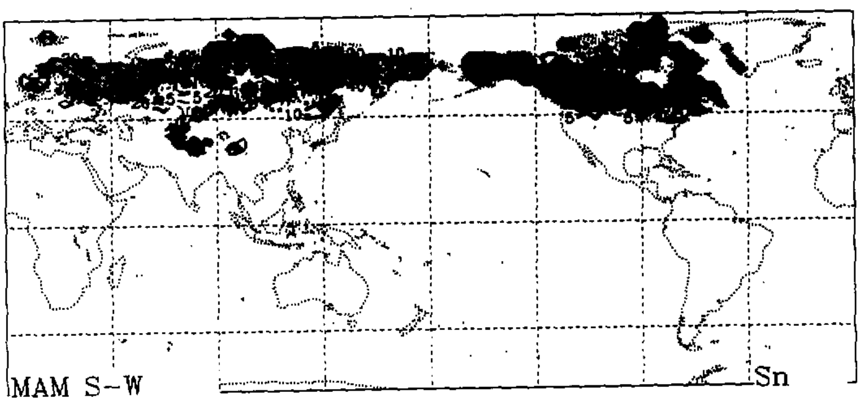


Fig.13. Composite difference of snow cover between strong (S) and weak (W) AWM year in spring (MAM). Units: mm (water equivalent).

enhanced large scale convection, as a forcing or Rossby wave source, works in concert with the forcing over the central and eastern Pacific to maintain or strengthen the extratropical flow pattern. This pattern is favourable to the southward intrusion of NE monsoon surge. Molteni et al. (1993) have suggested the positive feedback in tropical–extratropical interaction in their study on the effect of the eastern Pacific SST in La Niña years. Our results suggest that the AWM might be not only the medium of, but also a drive to this interaction.

The second effect is that to the east of the enhanced convection area, anomalous easterly flow would be induced, as seen in Fig.8. The enhanced equatorial easterly trades, in turn, might result in the cooling of SST in that area, through upwelling and evaporation. And below the convergence area of flow, there would be a maintenance or strengthening of warm SST.

It should be pointed out that the typical pattern of SST difference between strong and weak AWM (see Fig.10) already can be discerned in the preceding autumn. However, all three major centres are intensified dramatically in winter with the southward intrusion of AWM and weakened with the withdrawal of AWM. This consistency seems not incidental.

With the typical pattern of SSTA as forcing, the response of the atmosphere, both in tropics and extratropics on a seasonal scale, has been shown in previous sections. Therefore we may see that the AWM, again, plays the role of a drive and medium in the specific episode of air–sea interaction.

We may give a schematic diagram (Fig.14) to summarize the air–sea interaction of the specific episode which could be conceived as two related processes.

(i) Accumulative effect of stronger (or more frequent) outburst of AWM acts as a contributing factor to the development or maintenance of a La Niña type SSTA distribution.

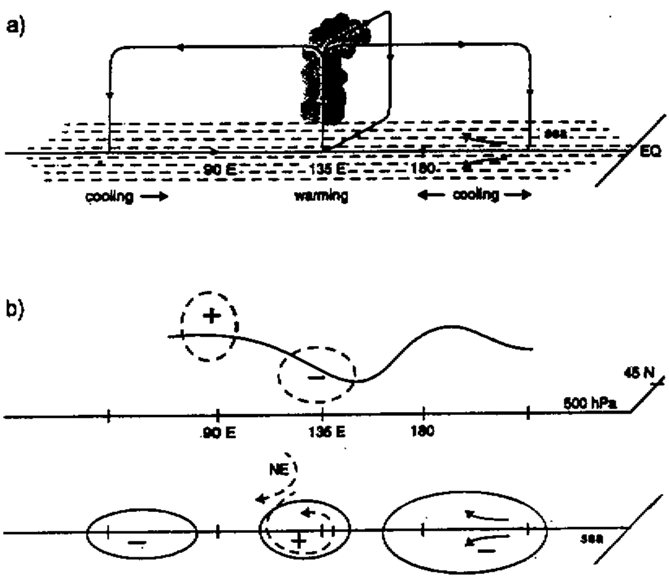


Fig.14. Schematic diagram to show the air–sea interaction in a strong Asian winter monsoon episode. (a) Southward intrusion of monsoon and variation of SST. (b) Extratropical response to the forcing of SSTA.

(*ii*) *The SST pattern established over the tropical Pacific consistently forces the atmos*pheric circulation, which helps form an extratropical flow pattern favourable for stronger Asian winter monsoon.

V. SUMMARY AND CONCLUDING REMARKS

The purpose of this study is to investigate the interannual variation of Asian winter monsoon (AWM) and the observed evidence of the linkage between a strong (weak) AWM and the circulation in the following seasons and to explore the operation of air–sea interaction in the evolution of the episode with emphasis on the role by NE monsoon. The preliminary results are as follows:

The simulation of ECHAM3 well reproduces the main features of the climatological mean Asian winter monsoon. The simulation shows pronounced difference of atmospheric circulation between strong and weak winter monsoon and for the consecutive seasons to follow. The different is not confined to Asian continent but is of global nature. Most striking is the appearance and persistence of an anomalous cyclonic flow over the western Pacific and enhanced Walker circulation for strong winter monsoon year in agreement with the observation. The contrast in summer rainfall patterns of both East China and India can also be discerned in the simulation.

Comparison of three sets of experiments with respectively climatological SST, global observed SST and observed SST only in the eastern tropical Pacific shows that the forcing from the anomalies of global SST makes a major contribution to the interannual variability of AWM and, in particular, to the interseasonal persistence of the salient features of circulation as seen in both the observation and simulation. The SSTA over the tropical western Pacific also plays an important part of its own in modulating the Walker circulation and the extratropical flow patterns.

The apparent effect of strong northeasterly monsoon is to enhance the convection over the tropical western Pacific. This effect, on the one hand, leads to a strengthening of SE trades to the east and extra westerly flow to the west, thus maintaining or developing a specific pattern of SSTA. On the other hand, the thermal forcing associated with the SSTA acts to strengthen the extratropical flow pattern which is, in turn, conducive to stronger monsoon activity.

Therefore, the NE winter monsoon seems to play a role of a medium or a drive for two related interactions, namely the air–sea and tropical–extratropical interactions. The result may suggest a certain self–sustained regime in the air–sea system, which is characterized by the above two related interactions with intermittent outburst of NE cold surge as linkage.

As a further medium for carrying anomalous information from winter to the following summer we found that the snow pack over Asia is above normal north of 55°N during strong AWM, but below normal to the south. The effect, working in concert with SSTA, may lead to a stronger Indian summer monsoon.

The limitation of the present study primarily lies in that the investigation is based on the result from a GCM with prescribed SST. The discussion on air–sea interaction seems more of reasoning than demonstration. To fully explore the role of Asian winter monsoon in the specific episode, a coupled air–sea model is needed. This will be the next step of our work.

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