Influences of Sea Surface Temperature and Ground Wetness on Asian Summer Monsoon

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ABSTRACT

The authors have conducted a series of experiments with a general circulation model to understand the influences of sea surface temperature (SST) and ground wetness (GW) (measured by snow amount and soil moisture content) on the Asian summer monsoon. The experiments are designed to illustrate the dominant features of monsoon response to SST and GW forcings and to delineate the relative importance of each forcing function in contributing to the variability of the monsoon.

Results indicate that ocean basin-scale SST anomalies exert a stronger control on the interannual variability of the monsoon compared to GW anomalies. The impact of SST anomalies on the monsoon appears nonlinear with respect to warm and cold events. The monsoon is weakened during the warm events but changes less noticeably during the cold events. The diminution of monsoon circulation associated with the warm SST anomalies is accompanied by a broad-scale reduction in water vapor convergence and monsoon rainfall.

Results also indicate that, following wet land surface conditions (enhanced snow and soil moisture) in the Asian continent during previous cold seasons, the summer monsoon becomes moderately weaker. Antecedent land surface processes mainly influence the early part of the monsoon. Wetter and colder conditions occur in the Asian continent during warm SST events. This results in reduced land–sea thermal contrast, which reinforces the weak monsoon anomalies produced initially by warm SST forcing. These interactive forcings are also responsible for the changes in the winter–spring westerlies over subtropical Asia, which are key precursory signals for the subsequent summer monsoon.

It should be pointed out that this study is conducted for the climate decade of 1979–88 only. The general robustness of the results needs to be explored by further investigations. In addition, chaotic features may have affected the results because of sampling errors.

1. Introduction

It has long been recognized that both sea surface temperature (SST) and land surface processes are two important factors in governing the interannual variability of Asian summer monsoon. Since the 1970s, experiments with general circulation models (GCMs) have significantly enhanced our knowledge of the physical mechanisms responsible for the influence of SST on the monsoon. Early atmospheric GCM studies aimed mainly at the response of monsoon to idealized SST forcing in either the Pacific Ocean or the Indian Ocean (Shukla 1975; Washington et al. 1977; Keshavamurty 1982). In contrast, recent studies have focused on understanding changes in the monsoon associated with observed SST anomalies (Palmer et al. 1992; Chen and Yen 1994; Ju and Slingo 1995; Yang et al. 1996; Lau and Bua 1998; Soman and Slingo 1997). During the past few years, coupled ocean–atmosphere models have also been used to study the relationship between the monsoon and SST anomalies of the Pacific and Indian Oceans (Latif et al. 1994; Nagai et al. 1995). In brief, most of these studies have shown that the tropical Pacific SST is important in modulating the interannual variability of Asian summer monsoon while the role of the Indian Ocean SST remains uncertain. More specifically, the warming of the central-eastern Pacific Ocean is generally accompanied by a weaker-than-normal Asian summer monsoon.

The influence of land surface processes on the Asian summer monsoon has also been studied extensively (Webster 1983; Yeh et al. 1984; Barnett et al. 1989; Yasunari et al. 1991; Meehl 1994a; Vernekar et al. 1995; Zwiers 1993; Douville and Royer 1996; Li and Yanai 1996; Lau and Bua 1998). Most GCM studies have focused on the relationship between the summer monsoon and the Eurasian snow amount in the previous cold season. It has been suggested that more (less) wintertime snow amount is generally followed by a weaker (stronger) Asian summer monsoon. In most of these experi-
ments, models are forced by prescribed snowmass. Hydrologic and radiative processes are usually involved in explaining the snow–monsoon relationship that has been simulated.

Despite our recognition of the influences of SST and land surface processes, the relative contribution of each forcing function to the variability of Asian monsoon has not been assessed. Unlike SST, forcing functions associated with land surface processes are very complex because of the interaction among ground temperature, soil moisture, and all components of surface energy budget. In GCM experiments, it is difficult to simply treat land surface processes in a universal way and results are likely different depending on what land surface parameters are specified. Furthermore, in the real system, SST and land surface process themselves are likely to be mutually interactive (Meehl 1994b, 1997; Yang 1996). This interactive nature gives rise to an additional difficulty in understanding the physical processes that are responsible for the variability of the monsoon.

Another important issue that is related to SST and/or land surface forcing is the cause of precursory signals of the Asian summer monsoon. For seasonal-to-interannual predictions of the broad-scale Asian monsoon, useful precursory signals have been found in the variability of subtropical upper-level westerlies during the previous winter–spring seasons. The observational study by Webster and Yang (1992) shows that prior to anomalously strong (weak) summer monsoon circulation, the 200-mb westerlies over subtropical Asia are weaker (stronger) than normal. These signals are potentially important for predicting monsoon variations because they occur over a broad region and last persistently for several seasons. Similar features of the precursory signals emerge also from GCM simulations (Ju and Slingo 1995; Yang et al. 1996).

Clearly, atmospheric processes themselves are unable to provide such a long-term (a few seasons) memory for the precursory signals, so they must have their roots in the slowly varying boundary forcing at the earth’s surface. Yang et al. (1996) have attributed the signals to both the remote influence from tropical SST and the more local impact of Asian land surface processes such as variations in snowmass and soil moisture content. However, the physical processes of the association between monsoon’s precursory signals and the dominant forcing function have not been demonstrated.

In this study, we aim at providing a better understanding of the mechanisms for the interannual variability and precursory signals associated with the Asian summer monsoon. To accomplish our goal, we conduct a series of experiments using the National Aeronautics and Space Administration/Goddard Laboratory for Atmospheres (GLA) GCM. We will depict the dominant features associated with the relative contribution of SST and land surface processes to the variability of the monsoon and its precursory signals. We will focus on the underlying physical processes of the impact of SST and land surface forcings on the variations of the broad-scale Asian monsoon. For land surface forcing, we focus on hydrologic processes measured by the changes in GW that is represented by the fields of snow mass and soil moisture content.

The organization of this paper is as follows. In section 2, we briefly describe the main features of the model, experiment design, and observational data used for validating model results. Presented in section 3 is a comparison between model and observations for the seasonal mean climate, annual cycle, and interannual variability of the Asian monsoon. In sections 4 and 5 we depict the interannual variability of the monsoon and discuss the relative contribution by SST and land surface anomalies. Issues of data sampling and the role of chaotic variability in the monsoon anomalies are addressed in section 6. In this section, we point out the limitation in data samples and the problem that our results may include features due to the impact of internal dynamics. A summary of the main findings by the study is given in section 7.

2. Experimental design and validation data

The model used in this study is the Atmospheric Model Intercomparison Project (AMIP) version of the GLA GCM. The 17-sigma-layer model with a horizontal resolution of 4° lat × 5° long is derived from an earlier 9-layer version of the model and has undergone continuous modification through the years (see Sud and Walker 1993). One of the important model features relevant to this study is that the early slab soil hydrology model has been replaced by the SiB model. According to Yang et al. (1994), in SiB the world vegetation is divided into two morphological groups: trees–shrubs and ground cover. Vegetation parameters are composed of three groups: morphological, physiological, and physical parameters. The soil structure is represented as three layers for which soil wetness is calculated. The change in land surface parameterization has led to a better representation of the influence of vegetation on land surface hydrologic processes.

For this work, three major experiments have been carried out for the climate decade of 1979–88. In the first experiment (AMIP), the model is forced by globally observed monthly SST, with fully interactive model atmosphere and land surface processes. Detailed features of the Asian monsoon simulated by this experiment have been reported in Lau and Yang (1996a) and Yang et al. (1996). In the second experiment (CLSST_A), the model is forced by the climatological annual cycle of the observed AMIP SST. In this experiment, snowmass and soil moisture content are prescribed by the daily values produced by AMIP, that is, the snow amount and soil moisture content of CLSST_A are identical to those of AMIP. Thus, the differences in atmospheric response between the two experiments are due to the impact of SST anomalies. The SST forcing used in the third ex-
Fig. 1. Schematic diagram summarizing the main features of the design for various experiments. The bottom part of the diagram represents the difference between AMIP and CLSST_A (measuring the impact of SST anomalies) and the difference between CLSST_A and CLSST_B (measuring the impact of GW anomalies).

The experiment (CLSST_B) is identical to that used in CLSST_A. The only difference between these two experiments is that the model atmosphere and land surface are fully interactive in CLSST_B. This atmosphere–land interaction should produce different snow and soil moisture from those specified in CLSST_A. Thus, the differences in atmospheric response between CLSST_A and CLSST_B result from the sole effect of different treatments of land surface processes. The above description of the various experiments is shown schematically in Fig. 1.

To help better understand the results of this study, we show in Fig. 2 the magnitude of the surface forcing functions: the time series of Niño-3 (5°S–5°N, 150°–90°W) SST used in AMIP and CLSST_A as well as those of soil moisture content averaged for 20°–50°N, 30°–130°E used in CLSST_A and CLSST_B. It can be seen from Fig. 2a that two major warm events (1982–83 and 1986–87) and two cold events (1984–85 and 1988) occur during the AMIP decade. The warm and cold events chosen for this study refer to these particular years, and the impact of the SST anomalies of these events on the Asian monsoon will be discussed subsequently. Figure 2b shows that soil moisture content changes clearly from year to year in both CLSST_A and CLSST_B. Although the largest differences between the two experiments usually appear during summer and fall seasons due to the effect of monsoon rainfall, values before the Asian summer monsoon season will be focused on in this study.

A number of observational data sets are used in this study to validate results from the control experiment (AMIP). They include the 200- and 850-mb winds from AMIP and NCEP reanalyses, averaged for December–February (DJF) and May–September (MJJAS), respectively. It can be seen that the GLA GCM simulates the upper-tropospheric atmospheric circulation reasonably well. For DJF (Figs. 3a,b), the Asian westerly jet stream and western Pacific anticyclone are captured quite realistically by the model. Both reanalyses and model show that, during MJJAS (Figs. 3c,d), southern Asia and its neighboring oceans are clearly under the influence of a large-scale anticyclonic circulation centered around 25°N, 95°E. To the southern flank of the anticyclone is the easterly monsoon flow, an important component of the monsoon system. Clearly, the model captures successfully these basic features of the Asian monsoon circulation. However, discrepancies between the reanalyses and simulation are also noticeable. For example, the model overestimates the intensity of upper northeasterly flow over the equator and tropical Southern Hemisphere during MJJAS.

Figure 4 shows the comparison between MSU and AMIP rainfall for both DJF and MJJAS. The model

3. Model-observation comparison

Figure 3 shows the 1979–88 climatologies of 200-mb winds from AMIP and NCEP reanalyses, averaged for December–February (DJF) and May–September (MJJAS), respectively. It can be seen that the GLA GCM simulates the upper-tropospheric atmospheric circulation reasonably well. For DJF (Figs. 3a,b), the Asian westerly jet stream and western Pacific anticyclone are captured quite realistically by the model. Both reanalyses and model show that, during MJJAS (Figs. 3c,d), southern Asia and its neighboring oceans are clearly under the influence of a large-scale anticyclonic circulation centered around 25°N, 95°E. To the southern flank of the anticyclone is the easterly monsoon flow, an important component of the monsoon system. Clearly, the model captures successfully these basic features of the Asian monsoon circulation. However, discrepancies between the reanalyses and simulation are also noticeable. For example, the model overestimates the intensity of upper northeasterly flow over the equator and tropical Southern Hemisphere during MJJAS.

Figure 4 shows the comparison between MSU and AMIP rainfall for both DJF and MJJAS. The model

Fig. 2. (a) Niño-3 SST (°C) index used in AMIP and CLSST_A. (b) Monthly values of area (20°–50°N, 30°–130°E) averaged soil moisture content (cm) used in CLSST_A and CLSST_B.
rainfall patterns are similar to the observed in many ways. This is particularly true for DJF (Figs. 4a,b) when many rainfall phenomena including the Asian–Australian winter monsoon, the Northern Hemisphere storm tracks, the southern Pacific convergence zone, and the intertropical convergence zone over Africa and the Indian Ocean are simulated quite realistically. For MJJAS (Figs. 4c,d), the model captures the main, large-scale features of monsoon rainfall such as the maximum centers over tropical Asia and Africa. The heavy rainfall over the western Pacific including the rain belt over the southwestern Pacific is also simulated reasonably well.
However, compared to the observations, the simulated rainfall does not extend sufficiently northward to the Bay of Bengal. Obviously, the model underestimates the rainfall over the equatorial Indian Ocean. As indicated by Lau and Yang (1996a), the model underestimates the Mei-yu rain belt in East Asia. The discrepancy between the observed and simulated rainfall over the equatorial central-eastern Pacific may be because the MSU generally overestimates the local rainfall compared to other satellite estimates (Lau et al. 1996).

To assess the skill of the GLA GCM in Asian monsoon simulation, in the remainder of this section we validate the model’s performance by focusing on the variability of the monsoon. Figure 5 shows the latitude–time cross sections of the climatological annual cycle of monsoon rainfall. The rainfall is averaged between 50°E and 120°E and shown for both MSU and AMIP. The figure indicates a similarity in the phase of the annual evolution between the observed and simulated rainfall. In both panels, monsoon rainfall attains maximum magnitude in June and reaches its northernmost position during July–August. Similarly to the observed, the model monsoon exhibits an abrupt northward jump during a period from late boreal spring to early summer. However, the transition of model monsoon is more gradual and confined to lower latitudes (10°–15°N) compared to the observed (15°–20°N). The sudden monsoon transition is one of the most prominent features associated with the onset of the Asian summer monsoon system (Lau and Li 1984). An updated, detailed description of this feature for the Southeast Asian monsoon has recently been given by Lau and Yang (1997) and Wang and Wu (1997).

Shown in Fig. 6 is the time series of monthly rainfall and 200-mb zonal wind component averaged within 5°–25°N, 40°–120°E. In general, the model simulates the variability of monsoon circulation and rainfall reasonably well. The correlation between the simulated and observed values yields a coefficient ($R$) of 0.95 for the wind and 0.88 for rainfall. The model produces a reasonable annual cycle of the monsoon. Realistic simulation can be found during some years but disappears during other years. In some years, a realistic circulation simulation can be accompanied by a less authentic rainfall production, and vice versa. The model generally underestimates the upper-tropospheric monsoon flow compared to the NCEP reanalyses and overestimates monsoon rainfall compared to MSU.

To have a more direct view of the interannual variability of the Asian summer monsoon, we show in Fig. 7 the yearly values of area-averaged MJJAS wind and rainfall. For circulation simulation (Fig. 7a), the model captures the reanalyses values quite realistically in 1981 and 1986. It simulates the weak monsoon in 1983 and 1987 as well as the strong monsoon in 1981, 1985, and 1988. The correlation between the AMIP and reanalyses flow is significantly above zero at the 95% confidence level. However, the general underestimation of the monsoon flow by the model is obvious, as shown in Fig. 6. It can be seen from Fig. 7b that the model captures the monsoon rainfall realistically in 1979, 1980, 1985, and 1986. However, it overestimates the rainfall especially in 1984 and 1988. The correlation between the simulated and observed rainfall gives a coefficient of 0.61, slightly smaller than 0.63 that is required for significance at the 95% confidence level.

It should be pointed out that the simulation of Asian monsoon by GCMs including the state-of-the-art models is far from perfect or realistic. A number of studies have shown that the models participating in AMIP do not score highly in monsoon simulation. Nevertheless, it has also been shown that the GLA GCM is among the best models in simulating the Asian monsoon system. Lau et al. (1996) intercompare the hydrologic processes simulated by the AMIP GCMs and find that the GLA GCM is among the top tier models in simulating various components of the global hydrologic cycle. Sperber and Palmer (1996) show that among the AMIP models, the GLA GCM performs favorably for all-India monsoon rainfall and Webster and Yang (1992) monsoon index. Gadgil and Sajani (1998) also show that the GLA GCM is one of the best models in capturing the year-to-year variability of all-India monsoon rainfall. According to Slingo et al. (1996) and Sperber et al. (1997), the GLA GCM gives one of the best simulations of atmospheric intraseasonal oscillation.

4. Impact of SST anomalies

In this section, we discuss the interannual variability of Asian summer monsoon and antecedent subtropical westerlies that is accounted for by SST anomalies. Because the change in SST forcing has been made globally in our experiments, it is difficult to link monsoon variability to regional SST anomalies. Although the role of the Indian Ocean in the Asian monsoon is uncertain, it has been known that the change in tropical Pacific SST

![Figure 5](image1.png)

**Fig. 5.** Latitude–time sections of monthly rainfall climatology averaged within 50°–120°E for (a) MSU and (b) AMIP.
is important in governing the variability of the monsoon system (e.g., Palmer et al. 1992; Nigam 1994; Soman and Slingo 1997) and the extratropical oceans play only a minor role (Lau and Nath 1994; Park and Schubert 1997). In this study, our focus will be on the impact of SST forcing from the tropical Pacific Ocean, especially the equatorial central-eastern Pacific.

\( a \). Warm and cold SST events

We define warm and cold SST events using the index of SST in the Niño-3 region (see Fig. 2). Based on the MJJAS values of the index, 1983 and 1987 have been chosen for the warm events, and 1984 and 1988 for the cold events. The composite patterns of SST anomalies for these warm and cold events are shown in Fig. 8. Except for the coastal waters and southeastern Indian Ocean, the sign of SST anomalies is nearly opposite between the two patterns. The most noticeable signal appears in the central-eastern Pacific Ocean where the SST anomalies in warm and cold events are of similar magnitude but opposite sign.

The response of Asian summer monsoon to the SST anomalies during the warm SST events is depicted in
Fig. 8. Composite patterns of MJJAS SST anomalies (°C) for (a) warm (1983 and 1987) and (b) cold (1985 and 1988) SST events.

Fig. 9. Composite patterns of the differences in MJJAS (a) 200-mb winds and (b) rainfall between AMIP and CLSST p for warm SST events. Associated with the warm SST anomalies, the monsoon becomes significantly weaker. The figure indicates that, over a broad region including tropical Asia, the Indian Ocean, and part of the western Pacific, the easterly monsoon flow in AMIP is much weaker than that in CLSST p. Consistently, monsoon rainfall in AMIP diminishes substantially in these regions. Over the equatorial central-eastern Pacific, the upper westerlies are weakened, and anticyclonic and cyclonic circulation patterns appear in the Northern and Southern Hemispheres, respectively. Compared to CLSST p, the rainfall in AMIP increases over the equatorial region but decreases over the subtropics. These features are similar to those observed during El Niño years (e.g., Rasmusson and Carpenter 1982; Yang and Webster 1990; Wang 1992).

In contrast, only a mixed signal can be found in the response of Asian summer monsoon to cold-event SST anomalies. Figure 10a shows that the monsoon circulation intensifies moderately over the western Pacific Ocean and the South China Sea but diminishes slightly in the regions to their west. Indeed, a weak cyclonic pattern appears over tropical Asia and the Indian Ocean. This indicates a weakening rather than strengthening of the monsoon. Figure 10b indicates that the response of monsoon rainfall to the cold SST anomalies exhibits a less organized pattern compared to the response to the warm SST anomalies (Fig. 9b). Although rainfall increases slightly in the southernmost India and part of Southeast Asia, a stronger reversal change can be seen to the north of the enhanced-rainfall band. Over the central-eastern Pacific, the upper westerlies become stronger over equatorial regions and weaker over the subtropics. Although this feature is similar to that observed during La Niña years, it is far from a mirror image of the feature portrayed in Fig. 9a considering the differences in both magnitude and phase. Obviously,
the rainfall response over the Pacific Ocean is not necessarily opposite to each other in Figs. 9b and 10b. Therefore, the response of monsoon circulation and rainfall seems nonlinear with respect to the warm and cold SST anomalies. It should be pointed out that we have also analyzed the year of 1984 (another cold SST year; see Fig. 2) and found that the feature of monsoon response is similar to that shown in Fig. 10.

Analyses of other parameters (not shown) indicate that during the warm events sea level pressure decreases over the central-eastern Pacific and increases over tropical Asian regions, leading to a reduction in the zonal gradient of atmospheric pressure. At the same time, both Walker- and Hadley-type circulations become weaker. Here, we explain the SST–monsoon relationship by focusing on the simultaneous changes in SST-induced tropical circulation or a “fast” forcing-response mode occurring during MJJAS only. For this reason, we refer to this influence of SST on Asian summer monsoon as SST’s direct impact.

The above feature of nonlinear response of Asian summer monsoon to SST is obtained by analyzing only a few data samples. Nonetheless, the result is supported by the results of of Soman and Slingo (1997) who study the response of monsoon to SST using the U.K. Universities’ Global Atmospheric Modelling Programme (UGAMP) GCM. In general, the equatorial Pacific SST anomalies are not linear functions of the Southern Oscillation (Philander 1990). Even in the case of tropical SST forcing with a precisely equal magnitude but opposite phases, the response of atmospheric circulation can still be nonlinear (Hoering et al. 1997). According to Figs. 9 and 10, a nonlinear response of atmospheric circulation and rainfall occurs even over the central-eastern Pacific where SST anomalies are nearly opposite. The above result is also supported to some extent by the observational study of Yang (1996) who shows that the Indian monsoon responds more significantly to El Niño events than to La Niña events. Because of the smaller impact of cold SST anomalies, we will focus only on the influence of warm SST forcing in the following discussion of this section.

b. Regional water vapor budget

Because water vapor provides the fuel for latent heat release in monsoon convection, it is also important to understand the water vapor budget associated with changes of the aforementioned monsoon circulation features. We have calculated the water vapor budget for three monsoon regions: the Indian subcontinent (I), Southeast Asia (II), and the East Asian continent (III). Figure 11a shows the vertically integrated (surface-to-200 mb) water vapor flux across the various boundaries and area-net water vapor budget during the warm SST events for AMIP. It can be seen that there is a net water vapor convergence for all three regions, especially in Southeast Asia. This suggests that in these regions the rainfall rate is usually larger than the surface evaporation rate. For the Indian monsoon (region I), water vapor is transported into the region from all directions except the east. Although most of the incoming water vapor is contributed by the transport from the western boundary, the area net gain (14.8 × 10^7 kg s⁻¹) is accounted for by the net meridional flux (30.2 × 10^7 kg s⁻¹), while the net zonal transport (−15.4 × 10^7 kg s⁻¹) is out of the region. In contrast, the Southeast and East Asian domains obtain water vapor from southern and western boundaries and export it in their northern and eastern boundaries. Both zonal and meridional transports contribute to the water vapor convergence in these two regions, although the gain is relatively small in the East Asian domain.

Figure 11b illustrates the difference in water vapor transport between AMIP and CLSST_A for the warm
events, portraying the impact of warm SST anomalies on the water vapor budget in the Asian monsoon regions. The arrows indicate the relative directions with respect to those shown in Fig. 11a. For a specific boundary, if the arrows are in opposite direction in the two panels, the magnitude of water vapor flux of AMIP is smaller than that of CLSST_A. Otherwise, AMIP flux is larger. It can be seen that, associated with the warm SST anomalies, less water vapor is transported through all the boundaries except some of the East Asian region (domain III). The largest reduction occurs in the western and eastern boundaries of the Indian domain. The weaker AMIP water vapor fluxes lead to smaller area water vapor convergence. The net gain is reduced by about 38% in the Indian monsoon region and 22% in Southeast Asia. It is found that the changes in surface evaporation rate from AMIP to CLSST_A are relatively small in these monsoon regions (not shown). Thus, the decrease in water vapor convergence accounts for most of the reduction in monsoon rainfall shown in Fig. 9b.

c. Precursory signals

As stated in the introduction, strong precursory signals of the Asian summer monsoon exist in the subtropical westerly circulation and they may be linked to tropical SST forcing. In this study, we not only show the reproduction of the signals in our simulations but also demonstrate the role of SST forcing in the signals. Figure 12 shows the composite patterns of the difference in 200-mb winds between AMIP and CLSST_A for the warm SST events. The patterns are presented for summer monsoon season as well as the winter and spring seasons prior to the monsoon. The warm SST anomalies give rise to a weak Asian monsoon in June–August (JJA). Prior to the weak monsoon, strong westerlies appear over subtropical Asia during the previous cold seasons, especially in March–May (MAM). This is manifested by the intensification and equatorward shift of the Asian subtropical westerly jet stream. The feature shown in Fig. 12, especially the MAM pattern, is similar to the result of Yang et al. (1996; see their Fig. 7) in which the roles of SST and land surface processes have not been clearly isolated. Since the changes shown in Fig. 12 have their root in SST only, it can be concluded that a remarkable portion of the antecedent signals is due to the warm SST forcing. Whether or not land surface processes are partially responsible for the precursory signals of the monsoon will be discussed in section 5d.

5. Impact of GW anomalies

We quantify the interannual variability of Asian summer monsoon and its precursory signals as a response to Asian ground wetness anomalies in this section based mainly on the analyses of results from experiments CLSST_A and CLSST_B.

Fig. 12. Composite patterns of the difference in DJF, MAM, and JJA 200-mb winds between AMIP and CLSST_A for warm events (1982–83 and 1986–87).

a. SST-induced changes in GW

Previous studies have suggested that tropical SST anomalies may exert an impact on the wintertime land surface conditions in the Eurasian continent (e.g., Meehl 1994a; Yang 1996). In the following, we will test this conjecture by examining the changes in snow amount and soil moisture during warm and cold SST events. It should be restated that even though the same climatological SST forcing is used in CLSST_A and CLSST_B, the anomalies of snow amount and soil moisture in CLSST_A are those induced by SST variability because they are identical to those of AMIP.

To estimate the impact of SST anomalies on GW variability, we show in Fig. 13 the December–April (DJFMA) averaged patterns of the differences in snowmass and soil moisture content between CLSST_A and
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**Fig. 13.** Changes in DJFMA (a)–(b) snowmass (kg m\(^{-2}\)) (c)–(d) and soil moisture content (cm). Shown are differences between warm SST winters of CLSST\(_A\) and winter climatology of CLSST\(_B\) (left panels), as well as those between cold SST winters of CLSST\(_A\) and winter climatology of CLSST\(_B\) (right panels).

CLSST\(_B\). The differences shown in the figure are computed between the values of warm SST winters (1982–83 and 1986–87), or cold SST winters (1980–81 and 1984–85) and the wintertime climatological values of CLSST\(_B\). This method will be justified shortly. It should be pointed out that we choose the warm and cold SST events here based on wintertime Niño-3 SST index (see Fig. 2). This means that the classification is not identical to that discussed in the last section where warm and cold events are determined by MJJAS SST anomalies. It can be seen from Fig. 13 that the warm SST anomalies produce more snowmass in subtropical Asia, especially near 30\(^°\)N. In contrast, associated with the cold SST anomalies, no organized pattern of snow change can be found in the subtropics (e.g., 25\(^°\)–35\(^°\)N) but more snow appears in higher latitudes instead. Associated with the increased snow amount produced by the warm SST anomalies, enhanced soil moisture content appears. Thus, the warming of central-eastern Pacific seems linked to wet land surface shown in both snow and soil moisture fields. Associated with the cold SST anomalies, soil moisture decreases in the subtropics but increases in the extratropics. Figure 13 suggests that there is a relationship between snowmass and soil moisture fields. However, this relationship is only moderate and may vary, depending on many factors including the condition of frozen land and the rate of snowmelt.

As pointed out above, for CLSST\(_B\), the values used in Fig. 13 are the DJFMA climatologies of the entire 10 years. This is done here and in the following to reduce the possible bias due to data sampling. It can be assumed that because of the nature of experiment design the interannual boundary forcing for CLSST\(_B\) is relatively smaller than that for other experiments. In fact, replacing those climatologies by the corresponding values of CLSST\(_B\) during the same individual winters chosen for CLSST\(_A\) yields similar results.

**b. Influence of GW on monsoon: SST’s indirect impact**

Figure 14 shows the changes in wintertime land surface temperature and sea level pressure that are accounted for by the snow and soil moisture forcings given in Fig. 13. The greater snow amount and soil moisture over the subtropical Asian land leads to a strong large-scale cooling in the continent (Fig. 14a). As a result, sea level pressure increases significantly (Fig. 14c). It is evident from Fig. 14 that the decrease in surface temperature and increase in atmospheric pressure over the Asian continent enhance the winter-type thermal contrast and pressure gradient between landmass and the neighboring oceans.

Figure 14 also indicates that the anomalies of snow and soil moisture that are first produced by cold SST anomalies do not cause well-defined changes in surface temperature and sea level pressure (right panels). Neither of these fields exhibits a pattern that is opposite to the warm SST-related response shown in the left panels of the figure. Although the decreased temperature and increased pressure can be seen in the extratropics associated with the enhanced snow and soil moisture, a large-scale warming is missing from the subtropics where soil moisture content is reduced.

The notion that a premonsoon moistening and cooling of the Asian landmass alone can weaken the following
**Fig. 14.** Changes in DJFMA (a)–(b) surface temperature (°C) and (c)–(d) sea level pressure (mb). Shown are differences between warm SST winters of CLSST_A and winter climatology of CLSST_B (left panels), as well as those between cold SST winters of CLSST_A and winter climatology of CLSST_B (right panels).

**Fig. 15.** (a) Composite patterns of the difference in MJJAS 200-mb winds between CLSST_A (1983 and 1987) and CLSST_B (climatology). (b) Same as (a), but for May–June only. The Asian summer monsoon circulation is demonstrated in Fig. 15. Figure 15a shows the difference in 200-mb winds between CLSST_A and CLSST_B for the MJJAS following the wet DJFMA as seen from Fig. 13. It can be seen that the Asian summer monsoon is noticeably weaker in CLSST_A than in CLSST_B, with the strongest signal over the latitude band of 15°–25°N. Clearly, the weakening of the monsoon is consistent with the decrease in surface temperature of the Asian continent (Fig. 14a). On the other hand, because of the smaller change in the DJFMA surface temperature during the 1980–81 and 1984–85 winters (Fig. 14b), the response of the Asian monsoon during their following summers is insignificant (not shown).

Figure 15 reveals another interesting feature: land surface processes mainly affect the early stage of the Asian summer monsoon. It can be seen that the May–September feature depicted in Fig. 15a is largely accounted for by the change in monsoon circulation shown in Fig. 15b, a corresponding pattern but for May–June only. Indeed, both panels of the figure show a similar pattern including the reduction of monsoon easterlies over subtropical Asia, although the feature of Fig. 15b appears more pronounced due partially to the smaller number of months used to calculate the average. We also examined the corresponding pattern for July–September only (not shown) and find no noticeable, similar change in the monsoon circulation. Recently, Dong and Valdes (1998) obtained a similar result from Asian monsoon simulations using the UGAMP GCM.

We have attributed the above-described weakening of monsoon to the enhanced snow amount and soil moisture in the Asian continent. Because of the design of
our experiments, however, the changes in ground wetness (prescribed in CLSST\textsubscript{A}) are rooted in the premonsoon warming of tropical central-eastern Pacific. Thus, we can consider that the winter–spring warming of the ocean exerts first a “simultaneous” influence on the land surface processes through which modulates then the intensity of the following summer monsoon. To distinguish this from SST’s direct impact discussed in section 4a, we refer to this influence as SST’s indirect impact on the Asian summer monsoon.

c. Influence of GW on monsoon: General impact

It has been shown in the above subsection that the warm SST anomalies reduce the intensity of Asian monsoon indirectly but cold SST seems to cause little change in the monsoon. Although the cold SST anomalies are accompanied by reduced soil moisture in subtropical Asia (Fig. 13d), the response of the monsoon is quite insensitive. Why are the reduced soil moisture and cold SST anomalies not followed by a strong summer monsoon? One explanation is that a vigorous monsoon tends to maintain its internal robustness and is less vulnerable to the effect of external forcing. This is similar to what has been shown in section 4: the direct response of the monsoon to SST anomalies is more impressive when the system is in a weak but not vigorous state. The conjecture is also supported by the studies of Yasunari (1990), Lau and Yang (1996b), and Wainer and Webster (1996). The other explanation resides in the possible error due to data sampling. To reduce the possibility of data sampling bias, we reexamine the relationship between soil moisture and monsoon using more samples. Figure 16 shows the year-to-year values of difference in DJFMA soil moisture content between CLSST\textsubscript{A} and CLSST\textsubscript{B} averaged within the region of 20\degree–50\degree N, 30\degree–130\degree E. It can be seen that relatively large changes in soil moisture do not necessarily have one-to-one correspondence with the warm or cold SST events. From the figure, we first classify 1982–83, 1985–86, and 1986–87 into a more soil moisture group and 1979–80, 1980–81, 1983–84, 1984–85, and 1987–88 into a less soil moisture category. Then, we examine the changes in the subsequent summer monsoon between these two groups.

Figure 17 displays the composite features of 200-mb circulation for the more and less soil moisture groups identified above. Obviously, the Asian summer monsoon becomes weaker following the increased soil moisture seasons, especially over the subtropics (10\degree–30\degree N). However, no strong monsoon can be found associated with reduced soil moisture. In fact, the monsoon changes much less noticeably in this category. Obviously, the anticyclonic pattern centered in the northern Arabian Sea (Fig. 17b) does not suggest intensification of the broad-scale Asian monsoon circulation. These results confirm those shown in the last subsection with fewer samples.

d. Precursory signals

Since previous studies have linked the variability of Asian summer monsoon to the antecedent change in subtropical westerlies, it is interesting to understand the influence of ground wetness anomalies on the precursory signals of the monsoon. Figure 18 shows the
changes in DJFMA 200-mb winds and 500-mb geopotential height associated with the more snow, more soil moisture, and colder land illustrated in Figs. 13–14. During the wet–cold seasons, the Asian westerly jet stream intensifies and shifts southward. Changes in the 500-mb geopotential height exhibit a consistent pattern: the East Asian trough becomes stronger. The anomalous pattern of the pressure systems is accompanied by stronger cold-temperature advection (not shown), which is consistent with the strong cooling of the subtropical Asian landmass. In brief, an increase in snow and soil moisture alone can cause a stronger Asian westerly jet and a deeper East Asian trough, the two important features that precede a weaker summer monsoon.

An alternative way to demonstrate the direct impact of GW anomalies on the following Asian summer monsoon is to follow a similar approach used by Yang et al. (1996) by analyzing results from CLSST_A alone. In this experiment, the interannual variability of SST forcing has been excluded but yearly GW anomalies remain. Figure 19 shows the DJFMA anomalies of snowmass and soil moisture prior to the summers of weakest monsoon (1982 and 1983) in CLSST_A. That stronger-than-normal subtropical westerlies appear during the previous cold seasons before the monsoon.

6. Further discussions

Up to now, we have discussed the influence of the earth’s boundary forcings on the variability of Asian summer monsoon. Obviously, our results are obtained based on the analyses of only a few data samples, especially for the analysis of SST’s impact. Undoubtedly, the robustness of our findings would be more credible if they were reproduced by larger samples. In addition, the internal dynamics associated with the variability of the monsoon without changes in boundary forcing has not been discussed so far. In this section, we address briefly the data sample issue and the role of internal dynamics.

a. Data sampling

The limitation in data sampling of this study occurs due to the short model simulation in each experiment.
In the previous sections, we have taken several approaches to strengthen the results found. To confirm the feature regarding the weak response of monsoon to cold SST anomalies shown in Fig. 10, we have analyzed an additional cold event (1984) and obtained similar results. In verifying SST's direct impact on land surface processes and indirect influence on the monsoon (Figs. 13–15 and 18), we have analyzed the climatology of CLSST\(^p_B\) to reduce the limitation in data sampling. It should be restated that, in CLSST\(^p_B\), model circulation is driven by climatological annual cycle of SST and small land surface forcing. Nevertheless, we have also tested our result using different data samples (from individual years) of the experiment. To verify the contribution of land surface processes to the precursory signals of Asian summer monsoon, we have examined not only the differences between CLSST\(^p_A\) and CLSST\(^p_B\) but also the anomalies of CLSST\(^p_A\) alone in different ways. To further bolster our results, we introduce here an additional experiment (CLGW; see Yang and Lau 1996 for more about this experiment) using the same GCM. In CLGW, the model is forced by monthly observed SST used in AMIP. However, snowmass and soil moisture content have been specified to the climatological cycles computed from the output of AMIP. Thus, the influence of SST anomalies on the Asian monsoon can be demonstrated alternatively by analyzing the results from this independent experiment.

Figure 20 shows the MJJAS anomalies of CLGW 200-mb winds for the warm and cold SST events that are identical to those analyzed in section 4. During the warm SST events, the Asian summer monsoon diminishes remarkably. On the other hand, the monsoon strengthens only slightly during the cold events. The small monsoon anomalies shown in the lower panel of the figure are due to the influence of warm-event monsoon in computing the mean climate because these weak signals disappear when 1983 and 1987 are excluded from the calculation of climatology. Over the central-eastern Pacific, El Niño– and La Niña–like changes in atmospheric circulation occur respectively during the warm and cold events. Because of the nature of experiment design for CLGW, the features shown in Fig. 20 are accounted for by SST forcing only. Thus, these results support those shown in Figs. 9 and 10 regarding monsoon's nonlinear response to warm and cold SST anomalies.

b. The role of internal dynamics

Our results so far have confirmed that external forcing such as SST anomaly plays a crucial role in the interannual variability of the Asian monsoon (e.g., Shukla 1987; Palmer et al. 1992; Chen and Yen 1994; Soman and Slingo 1997). While it has been claimed that chaotic features are prominent for the high-frequency modes such as the intraseasonal variations of the monsoon (Palmer 1994; Sperber and Palmer 1996; Webster 1996; Ferranti et al. 1997), how important internal dynamics are for the interannual variability of broad-scale monsoon is unknown yet because it is very difficult to distinguish the effect of internal dynamics from that of external forcing.

We provide here some evidence of the relative contributions of external boundary forcing and internal dynamics to the interannual anomalies of broad-scale Asian monsoon. Figure 21 shows the anomalies of vertical shear of the zonal wind component (averaged within 5°–20°N, 50°–120°E) for AMIP, CLSST\(^A\), and CLSST\(^B\). Shown in the figure are the monthly area-averaged values for each year of the AMIP decade (1979–88). For the summer months (see the shaded areas), the index is equivalent to the monsoon index defined by Webster and Yang (1992). It can be seen that the Asian monsoon varies from year to year in all experiments. The interannual variability of summer monsoon is most significant in AMIP in which interannual forcing of both SST and land surface processes is included. On the other hand, the smallest year-to-year change in summer monsoon occurs in CLSST\(^B\) in which there are no interannual SST anomalies and thus anomalous land surface forcing is presumably small. To quantify the above feature, we show in Table 1 the seasonally averaged variance of the wind shear (zonal wind, rather than anomalies) averaged for the area defined in Fig. 21. The table indicates clearly that except for DJF the variances in experiments CLSST\(^A\) and CLSST\(^B\) are much smaller than those of AMIP. For spring and summer seasons, the monsoon of CLSST\(^A\)
and CLSST_B changes only about half as much as that of AMIP. Both Fig. 21 and Table 1 demonstrate that SST forcing is crucial for the interannual variability of the broad-scale Asian summer monsoon. Without these SST anomalies, the changes of monsoon are relatively small regardless of the ways of specifying ground wetness anomalies. These boundary forcings explain a remarkable part of the interannual variability of the Asian summer monsoon. On the other hand, as shown in CLSST_B, a chaotic effect may account for a noticeable part of monsoon variations and thus cannot be ignored in describing even the low-frequency variability of the broad-scale Asian monsoon. In CLSST_B, the monsoon variance is caused by both internal dynamics and other boundary forcing (e.g., albedo effect) that has not been considered in our experiments.

7. Summary

Using the GLA GCM, we have performed a series of experiments to isolate the influences of SST and ground wetness (GW) on the Asian summer monsoon. We have identified the dominant signals associated with these influences and delineated the relative importance of each forcing function on the variability of the monsoon.

The main results of this study are summarized schematically in Fig. 22. Compared to GW variations, SST anomalies cause a more significant change in the Asian summer monsoon. For example, the reduction in MJJAS monsoon rainfall averaged within 10°–25°N, 50°–120°E, is about 1 mm day⁻¹ due to warm SST anomalies and 0.2 mm day⁻¹ because of increased GW. The response of monsoon circulation seems nonlinear with respect to warm and cold SST anomalies. The monsoon becomes significantly weaker during the warm events but changes relatively little during the cold events. Associated with warm SST anomalies, both the Walker circulation and local Hadley circulation become weaker. Consistently, the amount of atmospheric water vapor transported into tropical Asia is reduced. We refer to this influence of summertime warm SST anomalies as SST's direct impact on the monsoon.

In the absence of SST’s direct impact, the Asian summer monsoon becomes moderately weaker following wetter and colder conditions in the Asian continent. Land surface processes mainly affect the early stage of the monsoon. Results also indicate that winter–spring warm SST anomalies lead to increased soil moisture in the Asian continent and weaken indirectly the Asian monsoon during the following summer. Therefore, the influences of SST and GW on the Asian summer monsoon clearly involve mutually interactive processes. The GW variations reinforce the monsoon anomalies produced by warm SST forcing.

<table>
<thead>
<tr>
<th>Season</th>
<th>AMIP</th>
<th>CLSST-A</th>
<th>CLSST-B</th>
</tr>
</thead>
<tbody>
<tr>
<td>DIF</td>
<td>9.2</td>
<td>15.0</td>
<td>10.6</td>
</tr>
<tr>
<td>MAM</td>
<td>12.7</td>
<td>5.5</td>
<td>5.3</td>
</tr>
<tr>
<td>JJA</td>
<td>7.7</td>
<td>5.7</td>
<td>4.1</td>
</tr>
<tr>
<td>SON</td>
<td>21.0</td>
<td>10.0</td>
<td>19.2</td>
</tr>
</tbody>
</table>

Fig. 22. Schematic diagram summarizing the direct and indirect influences of SST and impact of ground wetness on the Asian summer monsoon and its precursory signals.
This study also shows clearly that both tropical SST and land surface processes are responsible for the variations of the winter–spring westerlies over subtropical Asia, precursory signals for the Asian summer monsoon. Finally, it should be pointed out that the results obtained by this study are based on analyses of the climate decade of 1979–88 only. The conclusions may not be generalized simply to other time periods. Another precaution is that the results may be influenced by errors due to the limited data samples used.

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