Simulation of local air-sea interaction in the great warm pool and its influence on Asian monsoon

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An ocean mixed layer model was coupled to an atmospheric general circulation model (AGCM) with the aim to investigate the effect of local air-sea interaction on Asian summer monsoon (ASM). When local air-sea interaction was allowed in the great warm pool [GWP, i.e., oceans with annual mean sea surface temperature (SST) above 28°C across Indian Ocean and the western Pacific], the simulated ASM climatology presents a substantial improvement in both the precipitation distribution and monsoon onset timing compared to the AGCM control run. To illustrate the main physical processes responsible for such a change, another AGCM sensitivity experiment was conducted, in which SST within the GWP was derived from output of the coupled run whereas SST outside the GWP are identical to the AGCM control run. Intercomparison results indicate that air-sea interaction modulates SST efficiently via wind-evaporation and cloud-radiation processes for most parts of the GWP, which in turn influences atmospheric circulation and precipitation pattern. In the Bay of Bengal, for example, SST reaches its annual peak before the monsoon onset and increases the atmospheric instability and moisture convergence above it, providing a favorable background for the development of deep convection. After the monsoon onset, however, increased surface wind speed and deceased incoming solar radiation flux lead to the colder in situ SST, lower humidity, and weaker convection. Hence properly representation of local air-sea interaction is crucial in mimicking and predicting the ASM in the coupled model.


1. Introduction

The Asian summer monsoon (ASM), including its seasonal characteristics, interannual variability, and interactions with various phenomena, has been investigated by innumerable studies as summarized in two monographs edited respectively by Chang [2004] and Wang [2006]. Since the ASM is a coupled system, for the purpose of climate simulation and prediction, the air-sea interaction must be properly represented in the coupled general circulation models (CGCMs).

As shown by previous studies [e.g., Goswami, 1998; Webster et al., 1998; Kang et al., 2002; Rajendran et al., 2004], simulation of the South ASM and its variability is one of the most challenging problems for atmospheric GCMs (AGCMs) and CGCMs. Similarly, the East ASM has also been found to be difficult to simulate by AGCMs and CGCMs [Kang et al., 2002]. To understand the role of the atmospheric circulation as a “bridge” between sea surface temperature (SST) anomalies in the tropical Pacific and those in the midlatitude northern oceans, Lau and Nath [1996] coupled an AGCM with a motionless, 50-m deep oceanic mixed layer model (OMLM) at individual grid points and the negative feedback of the mixed layer temperature anomalies on the imposed flux forcing is taken into account by introducing a linear damping term with a 5-month dissipative timescale. Alexander et al. [2002] further developed a coupled AGCM-OMLM to investigate the influence of ENSO teleconnections on air-sea interaction over the global oceans. These studies suggest that there is a close link between SST anomalies in the equatorial Pacific with those in the North Pacific, north tropical Atlantic, and Indian Ocean in boreal winter and spring, and surface heat fluxes are the key component of the atmospheric bridge driving SST anomalies. Lau and Nath [2006] further investigate the ENSO modulation of the interannual and intra-seasonal variability of the East ASM by using the same coupled model. Many aspects of their model simulation of the relationships between ENSO and the East Asian monsoon are in agreement with observational findings.

Interactions between surface processes and atmospheric convection are among the most poorly understood aspects of the climate system and its variability [e.g., Wu and Wang, 2001; Johnson et al., 2007; Wang et al., 2007]. Partially, progress in diagnosing these interactions has been...
mean by using an AGCM-OMLM coupled system, and the coupled feedbacks on the simulated monsoon are also investigated.

The outline of this paper is as follows. After a brief introduction of the data, model, and experimental design in section 2, the relative importance of local air-sea interaction throughout the global oceans is estimated by comparing the OMLM simulation results to the observation in section 3. Section 4 describes influence of the air-sea interaction on the precipitation pattern and onset timing of ASM by intercomparison modeling results and the observations. In section 5, mechanism of the local air-sea interaction in driving the coupled monsoon system is further discussed. Finally, conclusions and discussion are given in section 6.

2. Data, Model and Experiments Design

2.1. Data

The 6-hourly NCEP/DOE reanalysis data set [Kanamitsu et al., 2002] in the period of 1979 to 1989 is used to diagnose the evolution of the atmospheric circulation, surface heat fluxes and wind stresses during the ASM period. For all variables in NCEP/DOE, daily means were obtained first, and then climatological daily, pentad, and monthly were calculated as multi-year-means. The CPC Merged Analysis of Precipitation data [CMAP, Xie and Arkin, 1996] with the same period is used to explore the spatial and temporal nature of monsoon precipitation. The horizontal resolution is 1.875° × 1.875° for NCEP/DOE heat fluxes and 2.5° × 2.5° for all other variables.

2.2. Atmospheric Model

The AGCM, i.e., SAMIL2.08, which is developed in the State Key Laboratory of Numerical Modelling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)/Institute of Atmospheric Physics (IAP), is the updated version of the atmospheric component of the climate model system IAP/LASG FGOALS_s [Zhou et al., 2005]. The horizontal direction of SAMIL2.08 is rhomboidally truncated at zonal wave number (R42), roughly equaling to a grid of 2.8125° longitude and 1.67° latitude. In vertical direction 26-layer in hybrid-coordinate is adopted. The dynamical framework uses a “standard atmosphere reduction” scheme [Zeng, 1963; Phillips, 1973], and semi-implicit time integration program is introduced here. The model includes smoothed topography, gravity wave drag, and predicted clouds. The radiation scheme comes from Edwards and Slingo [1996]. The convection and condensation processes are parameterized by using the dry/moist convective adjustment [Tiedtke, 1989]. The land surface process implemented here is the SSiB model [Xue et al., 1991; Liu and Wu, 1997]. A more detailed description and recent improvements of this model are given by Zhou et al. [2005].

Early versions of SAMIL2.08 have been employed in simulating the effects of condensation heating on the formation of the subtropical anticyclone in the Eastern Hemisphere [Liu et al., 2001], seasonal variation of the ASM [Liu et al., 2004; Wang et al., 2004], effects of land-sea distribution on the formation of the ASM [Liang et al., 2005], and thermal forcing of the Tibetan Plateau on Asian climate patterns [Liang et al., 2006; Duan et al., 2008].
GCM Experiments Design in This Study

<table>
<thead>
<tr>
<th>Air-Sea Coupled Area</th>
<th>SST</th>
<th>Goal</th>
</tr>
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<tbody>
<tr>
<td>Exp1 None Provided by the AMIP-II Climatology at all open ocean grids</td>
<td>Examine the general performance of SAMIL2.08</td>
<td></td>
</tr>
<tr>
<td>Exp2 The great warm pool Generated by the air-sea coupled processes within the GWP and is same to Exp1 outside the GWP</td>
<td>Estimate effects of local air-sea interaction in the GWP on ASM</td>
<td></td>
</tr>
<tr>
<td>Exp3 None Same to Exp1 except that SST within the GWP is derived from Exp2</td>
<td>Examine the sensitivity of ASM to SST within the GWP</td>
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Results indicated that it can reproduce the mean large-scale climate pattern such as the precipitation centers located in the western India, Bay of Bengal, as well as the intertropical convergence zone and the climate variability in various timescale to a considerable degree [Zhou et al., 2005].

2.3. Mixed Layer Ocean Model

[9] The OMLM used in this work is a second-order turbulence closure model developed by Noh and Kim [1999]. The surface boundary conditions for turbulent kinetic energy, the parameterization of stratification effects on turbulence, and the design of convective process are improved in this model compared to those in Mellor and Yamada [1982]. It can successfully simulate the evolutions of profiles of both the dissipation rate and temperature and reproduces various important features of the oceanic boundary layer. This OMLM has been coupled to an oceanic general circulation model [Noh et al., 2002] and some CGCMs [Yim et al., 2008] to study climate variability for many aspects.

[10] The model has 50 vertical levels with uniform thickness of 5 m. Hence the ocean depth is deep enough to avoid the surface mixing reaching the bottom. On the basis of the results of a sensitivity experiment, an important empirical parameter α, which determines the dependence of the mixing length on stratification, is set to 5. Considering a fact that the turbulent kinetic energy in the GWP is particularly large after monsoon onset, the empirical constant m determining the turbulent kinetic energy flux at the sea surface is chosen as 400.

2.4. Experiment Description

[11] Totally five experiments (two OMLM runs and three GCM runs, description for GCM runs see Table 1) have been carried out in this work. The two OMLM runs are performed at each ocean grid by imposed surface fluxes forcing derived from NCEP/DOE and the AGCM control run (Exp1), respectively. By evaluating the OMLM simulated SSTs at each ocean grid in the global, relative importance of local air-sea interaction in various ocean basins can be detected. Also, comparison between the simulated SSTs driven by the forcing fields from NCEP/DOE and AGCM can test whether the AGCM can generate a reasonable surface heat fluxes and wind stress distribution.

[12] Exp1 is the AGCM control run without coupling processes, in which the 20-yr averaged monthly SST and sea-ice data as required by Atmospheric Model Intercomparison Project (AMIP) II (see http://www-pcmdi.llnl.gov/projects/amip/AMIP2EXPDSN/BCS_OBS/amip2_bcs.htm for details) are prescribed and interpolated linearly to each integration step. Thus there is no interannual variation in the forcing field. Results in this experiment are used to examine the general performance of the AGCM.

[13] Exp2 is the AGCM-OMLM coupled run, in which the OMLM is coupled to the SAMIL2.08 in the GWP. For the remaining ocean grids, SST is as same as that in Exp1. In the air-sea coupled area, AGCM provides surface heat fluxes (i.e., sensible heating flux, latent heating flux, net short-wave radiation flux, and net long-wave radiation flux) and meridional and zonal wind stresses to the OMLM at each integration step (600 s), the OMLM then feedbacks an updated SST. Thus the short-range physical processes and the effect of SST diurnal cycle are included in this coupling system. Assume that the effect of oceanic salinity variation on SST is not so important, we prescribe the salinity as a constant in the OMLM. Each experiment has been integrated 12 years, and the output of the last 10 years is averaged to compare with a 10-year mean (1980–1989) in the NCEP/DOE or CMAP. These procedures exclude the impacts of the interannual ocean variability such as El Niño.

[14] Exp3 is an AGCM sensitivity run, which is similar to Exp1 except that SST forcing field within the GWP is derived from the Exp2. The purpose of this experiment is to examine the response of ASM to the SST alone in GWP. Because of the absence of oceanic dynamics, we need add a nudging term in the OMLM to restore the long-term integrated SST toward the climatological SST, which can be regarded as the simplest assimilation scheme. It is incorporated into the thermodynamic equation in the first layer as

\[
\frac{\partial T}{\partial t} = \frac{Q^*}{\rho C_p \Delta \tau_1} + \frac{(T^* - T)}{\tau}
\]

Here \(Q^*\) is the net heat flux generated by AGCM, \(\rho\) the seawater density, \(C_p\) the specific heat, \(\Delta \tau_1\) the thickness of the first layer of the OMLM, \(\tau\) the restoring timescale, \(T^*\) the temperature of the first model layer, and \(T^*_o\) the observed SST. Clearly, a longer restoring timescale represents a weaker nudging effect. This nudging method is similar to the “combined boundary condition” of Noh et al. [2002]. After a systematic test, Manda et al. [2005] found that the restoring timescale of 1 day seems to be the most appropriate in Noh’s one-dimensional OMLM and when it embedded into a three-dimensional primitive equation model of the Japan Sea. In this study, \(\tau\) is chosen as 5 day in the OMLM and coupled runs.

3. Effects of Local Air-Sea Interaction on SST

[15] We first use the climatological 6-hourly forcing fields from NCEP/DOE and the AGCM control run (Exp1) respectively, to drive the OMLM for one year integration within 40°S to 40°N. Figure 1 displays the
spatial distribution of annual mean SST in both observation and simulations. A huge warm pool with the annual mean SST above 28°C covers the Indian Ocean and the western tropical Pacific, connected by South China Sea (SCS) and the Indonesian Through flow. We call it as the GWP so as to distinguish it from the traditional western Pacific warm pool, where the annual mean SST is above 29°C and the surface isotherms are oriented chiefly southeast-northwest.

The basic pattern of the global SST, together with the GWP and the cold tongue in eastern equatorial Pacific can be reproduced by the OMLM with forcing fields from either the NCEP/DOE or Exp1. Differences among the observation and simulations vary significantly with latitude change. In tropics, the simulated GWP seems to be larger and warmer than that in observation, with the western border extending to the coast of Africa. Meanwhile, the simulated cold tongue in the equatorial Pacific retreats eastward. The maximal error (roughly 2°C) occurs in the western boundary currents in each ocean, the Southern Hemispheric westerly belt, and the Niño1+2 and Niño3 areas, where the SST depends mainly on the ocean dynamics. In Niño1+2 region, for instance, the Ekman/Bjerknes feedback induces upwelling of the cold water and discharges the original surface warm water poleward in each hemisphere. Absence of this physics leads to a warmer SST in the eastern equatorial Pacific but colder to its north and south sides.

Since we focus on the local air-sea interaction in the GWP, Figure 2 shows the SST differences between two OMLM runs and observation only in the tropical oceans for each season. Although the error magnitude is generally larger in Exp1 forced case than that in the NCEP/DOE forced case, the seasonal distributions of spatial difference patterns are very similar between those two runs, denoting an acceptable quality of heat and momentum fluxes produced by Exp1. For seasonal variation, the differences are smaller in MAM and DJF, and the maximal errors usually appear in the Arabian Sea, Bay of Bengal (BOB), Kuroshio extension, western part of the south Indian Ocean, and the Maritime continent. The OMLM tends to generate a cooler (warmer) SST in Northern Hemisphere (Southern Hemisphere and equator). However, in JJA and SON, the error magnitude reaches roughly 3°C in some particular locations (e.g., Kuroshio extension), whereas it is still negligible away from coastlines. The deficiency in OMLM reflects need for a more reasonable treatment of ocean transport.
Nevertheless, the similarity between these two simulated SST annual cycles suggests that the current model design is suitable for studying the effect of local air-sea interaction on monsoon evolution through an analysis of the following coupling experiments.

4. Impacts of Local Air-Sea Interaction on Asian Summer Monsoon

[19] Deep convection experiences strong spatial coherence with the GWP. Previous studies have shown that tropical convection and surface winds undergo robust intra-seasonal variations accompanied by multi-scales high-frequency disturbances [Madden and Julian, 1994]. Associated with these variations are significant perturbations in surface heat fluxes that play an active role in modifying both the atmosphere and ocean. In this work, the difference between Exp2 and Exp1 represents the influence of the local air-sea feedbacks in the GWP and the difference between Exp2 and Exp3 denotes the response of the ASM to the SST alone induced by Exp2.

4.1. SST in the GWP and Its Linkage With Heat Flux and Wind Stress

[20] As we expected, SST produced by Exp2 is more similar to the observation compared to the OMLM runs in most seasons (Figure 3, note the scale is half of that in Figure 2). In DJF, Exp2 generated SST is about 2°C colder in the Northern Hemisphere and 1–1.5°C warmer in equator than that in NCEP/DOE, reflecting a larger amplitude of the annual cycle. Why the local air-sea coupling leads to the largest error occur in winter needs further investigation, but

Figure 3. SST difference between Exp2 and NCEP/DOE in °C. From top to bottom are MAM, JJA, SON, and DJF, respectively.
it is likely related to a fact that mid-latitude ocean mixed layer produces lower variability in response to high-frequency forcing [Deser et al., 2003]. Moreover, SST in NCEP/DOE during 1980–1989 is given by monthly and interpolated linearly to daily and 6-hourly, thus it excludes the short timescale variation.

[21] The tropics are unique because maximum incoming solar irradiance occurs twice per year so semiannual cycle can exceed annual cycle in places. For the domain average of the GWP, the SST in Exp2 presents the larger amplitude than that in Exp1 with the maximal error appearing in January (about 1°C), the minima of SST in annual cycle (Figure 4a). As mentioned in section 2.4, SST in Exp1 is monthly mean data and then interpolated linearly to each integration step, thus short-range variation is filtered. In Exp2, however, diurnal cycle of SST is generated by the air-sea coupled processes and high-frequency variation contributes to such a larger annual cycle in SST. The net heat flux in Exp2 is similar to the NCEP/DOE before and during the monsoon season (MJJ), whereas in the Exp1 is too large and Exp3 is intervenient (Figure 4b). Further compare each component of heat flux in experiment and the observation, it is shown that the net heat flux in tropical oceans depends mainly on solar radiation and latent heat fluxes (figures omitted here). In addition, wind stresses in Exp2 and Exp3 are generally stronger than those in NCEP/DOE and Exp1, featured by a greater annual cycle. This should be partially responsible for the larger annual cycle and salient SST difference during winter season.

4.2. Effects of Air-Sea Interaction on ASM

[22] From May to July, there are four precipitation centers located in the western coast of the Indian (IDO), BOB, SCS, and the Philippine Sea (PHS), with daily precipitation ranging from 6 to 12 mm per day. Particularly, the north BOB is known as the strongest large-scale rainfall center in the world (Figure 5a). Although Exp1 can reproduce the basic pattern, the precipitation magnitude is about twice larger than that in nature in IDO and BOB but too weak in SCS and PHS (Figure 5b). Indeed, during the history of the SAMIL development, continuous efforts, such as increasing the resolution of the model in both horizontal and vertical, replacing the old convection parameterization scheme, and removing the impacts of negative orography, have been made to reduce these precipitation errors. However, the effect is not satisfactory. After coupling the OMLM to the AGCM, the improvement of ASM precipitation simulation in Exp2 (Figure 5c) is substantial not only in distribution but also in magnitude. The precipitation in Exp2 reduces to ~12 mm/day in IDO and ~18 mm/day in BOB. On the other hand, it enhances to ~9 mm/day and ~12 mm/day in SCS and PHS. The precipitation magnitude in Exp2 is still somewhat larger than that in CMAP for most ASM areas, which might be related to the parameterization scheme for convection and condensation processes in
SAMIL2.08. Mean state of monsoon precipitation is very similar in BOB and IDO between Exp2 and Exp3 (Figure 5d), denoting the key role of SST as the lower boundary condition. However, excessive rainfall in SCS and PHS in Exp3 compared to that in Exp2 suggests other factors might be also important in these basins, and it will be further discussed in sections 4 and 5. In East China, Korea, and Japan, the difference is not obvious among three experiments and the observation. It may ascribe to much less rainfall magnitude and more controlling factors (e.g., synoptic systems from the higher latitudes, thermal forcing over the Tibetan Plateau, and variation of the Subtropical High) in determining the East ASM precipitation.

Figure 5. Spatial distribution of the monsoon precipitation in (a) CMAP, (b) Exp1, (c) Exp2, and (d) Exp3 during May to July (MJJ). Units: mm × day⁻¹.

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The ASM onset timing and propagation have been documented by countless studies. Wu and Zhang [1998] revealed that the whole procedure of the outbreak of the ASM onset consists of three consequential stages. The first is the monsoon onset over the eastern coast of the BOB in early May. It is followed by the onset of the East ASM over the SCS by 20 May, then the onset of the South Asian monsoon over India by 10 June. This result is further confirmed by Mao et al. [2004]. Other studies [Wang, 1994; Murakami and Matsumoto, 1994] argue that the ASM marches northeastward over the SCS and western North Pacific with three distinct stages observed in mid-May, mid-June, and late-June [Wu and Wang, 2001]. To diagnose the ASM onset evolution, here we use the Webster and Yang monsoon index (1992, WYI hereafter), which is defined as

\[
WYI = \frac{(U_{850} - U_{200})}{6.5}
\]

Where U means zonal wind and the subscript represent isobar.

Boreal spring represents a period of remarkable transition for the Indian Ocean because it is accompanied by the onset of the south ASM. The sustained monsoon winds drive vigorous ocean transports of heat and momentum that are responsible for the large annual cycle in SST in the Indian Ocean [Loschnigg and Webster, 2000]. Figure 6 shows the pentad mean WYI evolution for those four monsoon subregions mentioned before. In NCEP/DOE (Figure 6a), the sequence of monsoon onset is BOB in early May, SCS in mid-May, IDO in late-May, and PHS in mid-June (Table 2). This result is similar to previous studies, small differences arise from the various selected monsoon index, averaged domains, and data sources. The orders of the ASM onset can be well simulated in all experiments. Thus the SAMIL2.08 can simulate the basic monsoon evolution, including the onset and peak in monsoon precipitation over India and the western North Pacific. However, the monsoon onset timing in our modeling results is somewhat earlier than the observation. For example, the onset time in BOB is about 4 pentads earlier in the Exp1 (Figure 6b), and 2 pentads earlier in Exp2 (Figure 6c) and Exp3 (Figure 6d). Like other current AGCMs, this deficiency arises mainly from the exaggerated land surface solar radiation flux, which induces the higher surface air temperature and the subsequently intensified land-sea thermal contrast. The coupled model can improve the simulation of the ASM onset timing when local air-sea feedbacks are included. It is well known that two fundamental factors for origin of monsoon are seasonally varying solar radiation and land-sea thermal contrast. In the GWP, however, local air-sea
interaction acts as an important modulator in the ASM evolution.

5. Role of Local Air-Sea Interaction in the GWP

5.1. Effects of SST on AMS Onset

Regional oceans have been considered as a requirement in setting up the background forcing of the monsoon by heating the atmosphere and supplying moisture via evaporation. A statistically significant positive correlation exists between equatorial Indian Ocean SSTs and Indian precipitation in the winter prior to the monsoon wet season, after the removal of ENSO’s influence [Clark et al., 2000].

Near the time of the initial monsoon onset, warm episodes are often observed, yet as the monsoon becomes established, SST near the equator cools. To clarify the feedbacks

Table 2. Occurrence Time (in Pentad) of Monsoon Onset and Peak SST (in Brackets) in NCEP/DOE and Experiments in Four AMS Subregions

<table>
<thead>
<tr>
<th></th>
<th>BOB</th>
<th>SCS</th>
<th>IDO</th>
<th>PHS</th>
</tr>
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<tbody>
<tr>
<td>NCEP/DOE</td>
<td>26(23)</td>
<td>27(28)</td>
<td>29(28)</td>
<td>36(35)</td>
</tr>
<tr>
<td>Exp1</td>
<td>22(23)</td>
<td>26(28)</td>
<td>27(28)</td>
<td>36(35)</td>
</tr>
<tr>
<td>Exp2</td>
<td>24(23)</td>
<td>25(25)</td>
<td>27(28)</td>
<td>29(31)</td>
</tr>
<tr>
<td>Exp3</td>
<td>24(23)</td>
<td>26(25)</td>
<td>27(28)</td>
<td>28(31)</td>
</tr>
</tbody>
</table>

Figure 6. Evolution of Webster and Yang monsoon onset index (WYI) in (a) NCEP/DOE, (b) Exp1, (c) Exp2, and (d) Exp3 in units of m × s⁻¹. Black, red, green, and blue curves are for BOB, SCS, IDO, and PHS, respectively. Vertical dashed lines denote the monsoon onset date.
of the air-sea interaction in the GWP, SST and precipitation evolution for the four subregions during the ASM period are depicted in Figure 7. The orders of seasonal maximum SST are consistent with those of monsoon onset, with a slightly leading or overlapping with them (Table 2). In Exp2 and Exp3, for example, SST attains the peak at the mid-April (23rd pentad), end of April (25th pentad), mid-May (28th pentad), and end of June (31st pentad) in BOB, SCS, IDO, and PHS. The SST maximum leads obviously that in Exp1 in SCS (3 pentad) and PHS (6 pentad), where the simulated precipitation in Exp1 is systematically less than the observation and the Exp2. The responsible mechanism for SST seasonal evolution might be different in each subregion. For example, from winter to spring in SCS, as the mixed layer depth shoals against the diminishing of the northeast monsoon, heat gained from the atmosphere is trapped in a shallower surface mixed layer and eventually warms up the SST [Qu, 2001]. However, a common feature in Figure 7 is that SST change usually precedes rainfall change, i.e., SST increases persistently and approaches to the annual-cycle peak, and then precipitation starts to increase sharply and the monsoon onset occurs.

The close linkage between SST change and monsoon onset can be further proved by the lead/lag correlation test shown in Figure 8. A significant leading and simultaneously correlation between SST and precipitation can be detected in observation and the simulations in most cases. In NCEP/DOE, the largest positive correlation happens when SST precedes rainfall 9 pentads in BOB, 2 pentads in SCS, 6 pentads in IDO, and 1 pentad in PHS. The situation is similar in Exp2 and Exp3, with a slightly weaker correlation value especially in BOB. In Exp1, although the SST-precipitation correlation in BOB, SCS, and IDO is still consistent with the observed one, the result in PHS is almost opposite, i.e., maximum precipitation precedes SST by 6 pentads in Exp1. Typically, the SST-rainfall correlation in PHS is predominantly negative in Exp1 while the observed correlation is positive, indicating that the Exp1 cannot represent properly the air-sea feedback in the PHS and subsequently a poor monsoon evolution there.

From winter to summer, SST increasing provides a favorable background for large-scale convection in the tropical oceans. Warmer SST can increase surface air temperature and humidity via surface heat fluxes and turbulent mixing in the atmospheric boundary layer. Moreover, SST gradient can enhance low-level moisture convergence through differential surface heat fluxes [Lindzen and Nigam, 1987]. In the Javanese coast, for instance, the strong

Figure 7. Evolution of the pentad-mean SST (dotted curves, in units of °C) and precipitation (solid curves, in units of mm × day⁻¹) in (a) BOB, (b) SCS, (c) IDO, and (d) PHS. Black, blue, red, and green curves are for NCEP/DOE, Exp1, Exp2, and Exp3, respectively. SST in NCEP/DOE and Exp1 is represented by square curve, and SST in Exp2 and Exp3 by open circle curve in each panel. Vertical dashed lines denote SST peak.
SST gradient around the 28°C isotherm contributes toward forced horizontal convergence of moist air and subsequent deep convection during summer monsoon [Rajendran et al., 2004]. When SST attains a threshold value (above 30°C in Indian parts and 29°C in the SCS and PHS), the atmospheric instability near the surface enhances gradually. Also, the moisture convergence increases over there. As the atmospheric low-frequency oscillations propagate there, monsoon onset happens accompanying by abundant precipitation.

The scatter diagram shown in Figures 9a, 9c, 9e, and 9g demonstrates consistent changes in SST and instability index (defined as the vertical gradient in saturation equivalent potential temperature between 1000 hPa and 700 hPa). In addition, surface moisture convergence tends to increase with SST (Figures 9b, 9d, 9f, and 9h), especially when SST above 28°C. The atmospheric instability and moisture convergence indexes used in this study equal to those of Wu and Wang [2001] and Wu [2002]. Surface moisture divergence was calculated by using daily mean surface wind.

Figure 8. Lead-lag correlation between pentad mean SST and precipitation in (a) NCEP/DOE, (b) Exp1, (c) Exp2, and (d) Exp3. Black, red, green, and blue curves are for BOB, SCS, IDO, and PHS, respectively. Hatchings in each panel denote significance at 95% level. Negative (positive) value in abscissa means SST lead (lag) precipitation.
Both of them are 10-year-mean average over the GWP. Clearly, these relationships are significant in the observation and all three experiments, suggesting the overwhelming role of SST in triggering the convective activity in the GWP. Differences in some details among the observation and simulations also exist. In the NCEP/DOE, the lower atmosphere seems to be more unstable and the atmospheric instability index exceeds 70 K/1000 hPa when SST above the 28°C, whereas in all three experiments it always below 70 K/1000 hPa.

5.2. Monsoon Onset Induced SST Change

Now we need discuss the feedbacks of monsoon on SST. The processes for SST change include surface heat flux, oceanic mixing, upwelling, and oceanic advection. As indicated by Webster et al. [1998], surface heat flux is a chief factor in the SST seasonal change in the western Pacific warm pool.

To further demonstrate the role of cloud-radiation and wind-evaporation processes in alternating SST change, in Figure 10 we present the tendencies of SST, surface net short-wave radiation flux, as well as latent heat flux in BOB before and after monsoon onset for three experiments. Here the pentad mean data is used and the monsoon onset date is defined as the Webster-Yang index aforementioned for each case. Typically, monsoon onset induces changes in heat fluxes by altering cloudiness and surface wind. The increased cloudiness reduces locally the incoming short-wave radiation. On the other hand, the increased wind speed and intrusion of monsoon disturbances enhances evaporation and oceanic mixing. In Exp1 SST is a forcing field without any influence by the atmosphere and increases by 0.2–
0.3°C per pentad even with remarkable decreasing in solar radiation and increasing in latent heating after the monsoon onset (top of Figure 10). Because of the absence of air-sea feedback, the persistent high SST maintains a favorable boundary condition for strong convection and the subsequent overestimated rainfall magnitude in BOB shown in Figure 5b. However, after the BOB monsoon onset in Exp2, the incoming short-wave radiation reduces continuously [nearly \(-20\) \(\text{W m}^{-2}\) \(\text{Pentad}^{-1}\), Figure 10e] and evaporation further results in substantial increasing of upward latent heat flux [nearly \(60\) \(\text{W m}^{-2}\) \(\text{Pentad}^{-1}\), Figure 10f]. Two feedbacks, i.e., cloud-radiation and wind-evaporation results in the remarkable cooling in SST in southern BOB [\(-0.4^\circ\text{C Pentad}^{-1}\) in average, Figure 10d]. Further study reveals that the increase in latent heat flux is mainly due to enhanced surface winds (figures omitted here). In Exp3 (bottom of Figure 10), SST as the same as that in Exp2 and the atmosphere acts as a slave to the SST forcing. The response of surface solar radiation and latent heating fluxes after the monsoon onset is similar to that in Exp2, only with a smaller change.

Moreover, the remarkable decreasing (increasing) of solar radiation (latent heating) flux after the monsoon onset also happens in the rest three AMS subregions (figures not shown here). As shown in Table 2, SST peak in IDO or PHS lags the in situ monsoon onset one or two pentads in the Exp2. Thus the oceanic signal after the monsoon onset is not a negative SST tendency, instead it is a significantly reduced SST-increasing tendency and turns to a decreasing tendency one or two pentads later.

6. Summary and Discussions

[32] An AGCM-OMLM coupled model is constructed in this work. By comparing the simulation results with the observation and two other AGCM experiments, the local air-sea interaction processes and its impacts on the ASM are investigated. Some main conclusions are summarized as follows:

[33] 1. The simulation results given by an OMLM indicate that the north Indian Ocean and the tropical western Pacific is characterized by particularly active local air-sea interactions.

[34] 2. When the OMLM is coupled to an AGCM in the GWP, a much more reasonable representation of the seasonal mean monsoon precipitation in the magnitude and pattern is produced compared to the AGCM control run. The overestimated (underestimated) precipitation amount in BOB and IDO (SCS and PHS) are improved substantially. Moreover, the local air-sea feedback generates a more reasonable ASM monsoon onset timing.

[35] 3. Seasonal evolution of SST in the GWP provides a favorable background for the tropical monsoon onset in late spring or early summer. As SST reaches its annual peak, the atmosphere instability increases sharply accompanied by increased moisture convergence.
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Appendix A: Abbreviations List in the Text

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Full name</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGCM</td>
<td>Atmospheric General Circulation Model</td>
</tr>
<tr>
<td>AMIP</td>
<td>Atmospheric Model Intercomparison Project</td>
</tr>
<tr>
<td>ASM</td>
<td>Asian Summer Monsoon</td>
</tr>
<tr>
<td>BOB</td>
<td>Bay of Bengal</td>
</tr>
<tr>
<td>CGCM</td>
<td>Coupled General Circulation Model</td>
</tr>
<tr>
<td>CMAP</td>
<td>The CPC Merged Analysis of Precipitation</td>
</tr>
<tr>
<td>GWP</td>
<td>Great Warm Pool</td>
</tr>
<tr>
<td>IAP</td>
<td>Institute of Atmospheric Physics</td>
</tr>
<tr>
<td>LASG</td>
<td>the State Key Laboratory of Numerical Modelling for Atmospheric Sciences and Geophysical Fluid Dynamics</td>
</tr>
<tr>
<td>OMLM</td>
<td>Ocean Mixed Layer Model</td>
</tr>
<tr>
<td>SCS</td>
<td>South China Sea</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperature</td>
</tr>
<tr>
<td>IDO</td>
<td>Indian Ocean</td>
</tr>
<tr>
<td>PHS</td>
<td>Philippine Sea</td>
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</table>

[36] 4. After the monsoon onset, the atmospheric feedbacks cool SST through enhanced evaporation and reduced solar heating caused by deep convection during the active phase of monsoon. Consistently, the relationship between SST anomalies and surface fluxes implies the dominance of the cloud-radiation and wind-evaporation feedbacks over convective regions of the Indian and western Pacific Oceans.

[38] 5. Absence of such kind of feedbacks in the AGCM maintains the high SST even after the monsoon onset, which in turn generates more precipitation and the exaggerated precipitation centers in IDO and BOB.

[39] Within the GWP, monsoon precipitation among individual subregions seems to be closely related to each other. More precipitation in BOB and IDO often accompanied by less precipitation in SCS and PHS, and vice versa, suggesting the intrinsic linkage in the whole ASM system. Further investigation is then required to clarify this point.

[40] In this work the parameters in OMLM are chosen as same values across the GWP. However, wind stresses, heat fluxes, and ocean dynamics vary in ocean basins and seasonal dependent. To provide more reasonable simulation results, those parameters should be examined systematically. Also, the contribution of change in salinity to SST in the AGCM-OMLM coupled model need an evaluation.

[41] Moreover, intraseasonal oscillation is closely related to the summer monsoon variation. The simulation of intraseasonal oscillation, together with the impacts on the ASM by using the coupled model, will be conducted in our future work.

References


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