ABSTRACT

Simulation and prediction of the South Asian summer monsoon in a climate model remain a challenge despite intense efforts by the atmosphere and ocean research community. Because the phenomenon arises from the interaction of the atmosphere with the upper ocean, a deficiency in the simulation of the latter can lead to a poor simulation of the atmospheric meridional circulation. This study demonstrates that a significant improvement can be obtained in the simulation of the summer monsoon by correcting a prevailing deficiency in the mixed layer simulation of the Indian Ocean. A particular physical process of the nonbreaking wave–ocean mixing parameterized as $B_y$, which has not been considered in any climate model, is included in this study to enhance the vertical mixing in the upper ocean. Results show that the inclusion of this mixing process in a climate model leads to a better simulation of the ocean mixed layer, especially in the regions where the mixing was previously underestimated. The improved mixed layer simulation further results in stronger meridional differential heating, which drives stronger low-level monsoonal winds and results in stronger moisture transport and convergence, especially in the northern Indian Ocean. Moisture convergence into the Bay of Bengal is significantly enhanced and in general the spatial distribution of moisture is more consistent with observations. The directly driven monsoonal winds by the differential heating are further amplified by the resultant latent heating, which generates not only a wind amplitude comparable to the observations but also a correct vertical structure.

1. Introduction

The English word “monsoon” has its origin in the Arabic word *mausam*, meaning season (Webster 1987; Slingo 2002), and was used to describe the seasonal reversals of winds and wetness over tropical regions (Ramage 1971). It was later extended to reversal-like changes of winds and wetness in subtropical regions such as East Asia (Tao and Chen 1987). Monsoons are one of the major components of the global climate system; they have tremendous societal and economic impacts not only on the monsoonal regions but also over the globe (Xu 1958; Webster 1994; Simmonds et al. 1999; Qi 2009).

The South Asian summer monsoon (SASM) is believed to be driven by differential heating caused by land–sea contrast in association with the solar seasonal cycle (Webster 1987; Slingo 2002). This near-surface differential heating drives massive sea–land breezes in the lower troposphere of the atmosphere, but the response to the heating is confined to the heating layers (Wang et al. 1996; Wu et al. 1999, 2000; Wu 2003). As a consequence, the differential heating itself cannot directly drive monsoons of observed amplitude and the oppositely directed circulations in the lower and upper troposphere.

To drive monsoons of the observed amplitude and structure, deep latent heating is needed. The low-level circulation directly driven by differential heating brings a large amount of moisture from the ocean to the land, which induces deep thermal heating over the land associated with the condensation of the converged moisture, in particular under the influence of elevating topography (Yanai et al. 1992; Wu and Zhang 1998). The thermally driven circulation further strengthens the monsoonal circulation and provides more moisture for condensation. The diagnosed thermal heating from observation can drive monsoonal circulations approximately the same as observation (Zhang and Krishnamurti 1996). However, the moisture that is brought to the land from the ocean is constrained by the ocean surface conditions...
such as the ocean surface temperature and winds that are closely tied to the mixed layer (ML) of the upper ocean.

In this study, we focus on how the change of vertical mixing in the upper ocean can change the monsoon circulation, by comparing coupled model simulations in which the mixed layer parameterization scheme is changed. A component that we add to the mixed layer parameterization scheme is the nonbreaking wave-induced mixing (Qiao et al. 2004, 2010). Previously, we have demonstrated that incorporating this nonbreaking wave-induced mixing can lead to a more realistic simulation of the mixed layer depth over nearly all regions of the global oceans in various forced ocean models and in coupled climate system models such as the Princeton Ocean Model (POM; Qiao et al. 2010), Regional Ocean Modeling System (ROMS; Wang et al. 2010), and Modular Ocean Model, version 4 (MOM4; Shu et al. 2011). Here we show that such improved ocean mixed layer and the accompanying fluxes at the air–sea interface lead naturally to an improved simulation of SASM in a coupled atmosphere–ocean general circulation model.

The paper is organized as follows: Section 2 introduces the theory of the nonbreaking wave-induced mixing. The model, experimental design, and observation data are described in section 3. Section 4 compares the simulated monsoonal circulation and analyzes the physical processes in different model settings. Conclusions are given in section 5.

2. Wave-induced mixing

The surface wave is the most energetic component in the upper oceans and plays an important role in the vertical mixing process. There are two well-recognized ways for surface waves to influence the ocean: breaking (Drennan et al. 1996; Mellor et al. 2008) and nonbreaking (Qiao et al. 2004; Babanin and Haus 2009; Dai et al. 2010) waves. Both processes can enhance the viscosity and diffusivity in the ocean. It is noticed that the effects of turbulence generated by wave breaking are mostly limited within a depth on the order of wave amplitude, while the effects of turbulence generated by nonbreaking wave can penetrate into a depth on the order of the wavelength. Furthermore, the Reynolds stress induced by wave movement can transfer kinematic energy from surface waves to the ocean circulation. Vertical mixing in ocean circulation models can be calculated using two types of turbulence closure schemes: the intensive mixing scheme and the continuous mixing layer scheme. The difference between them is this: by using the continuous mixing layer scheme based on $K$-theory and turbulence closure methods, we can simulate the vertical structure of the upper mixed layer. Some parameterization schemes are widely used in current ocean circulation models, such as the $K$-profile parameterization (KPP scheme; Large et al. 1994), the Pacanowski and Philander scheme (PP scheme; Pacanowski and Philander 1981) and the Mellor–Yamada scheme (M-Y scheme; Mellor and Yamada 1982).

Generally, the mixing in ocean models resulting from these mixing schemes is relatively weak compared with observations, especially in summer (Martin 1985; Kantha and Clayson 1994; Ezer 2000; Mellor 2001). Previously, it was assumed that in these schemes some processes, such as internal waves at the bottom of the mixed layer, were not well represented. To include the effects of these vertical mixing processes, a Richardson number–dependent mixing below the ML (Kantha and Clayson 1994) or a Richardson number–dependent dissipation (Ezer 2000; Mellor 2001) were attempted. While this approach shows some positive signs of correction, in particular when high-frequency winds are used and penetrating shortwave radiation under the ocean surface is employed (Ezer 2000), nonetheless vertical mixing still remains underestimated in models.

Another attempt is to consider the contribution of surface waves to ocean mixing and wave–current–turbulence interaction processes (Craig and Banner 1994; Terray et al. 1996; Stacey 1999; Burchard 2001; Malcherek 2003; Mellor 2003, 2008; Mellor and Blumberg 2004; Kantha and Clayson 2004; Ardhuin and Jenkins 2006). The mixing related to the breaking waves is mostly confined to the top few meters, on the order of wave amplitude near the surface. It may also require very fine vertical resolution in the surface layer of the model in order to even resolve it.

To further alleviate the problem of the vertical mixing underestimation, the nonbreaking wave-induced vertical mixing is introduced into the mixing schemes (Qiao et al. 2004, 2010). While the exact physical mechanism of the nonbreaking wave-induced vertical mixing has not been fully understood, the underlying assumption is that the wave-related motions have scales comparable to that of shear-induced turbulence; thus, the interaction between wave-induced motion and circulation-related turbulence is employed to derive a so-called nonbreaking wave-induced mixing, represented by a new mixing coefficient $B_w$. In this process, the waves that have a significant role in inducing mixing are the surface gravity waves with horizontal scale on the order of about 100 m. Although the horizontal scale of these waves is much smaller than the scale of horizontal ocean circulation, the scale of the wave-induced vertical velocity in the upper ocean can be comparable to or even greater than the vertical velocity variation of the circulation. Similar ideas
concerning wave-induced turbulence can also be found in Babanin (2006) and Babanin and Haus (2009).

Qiao et al. (2004, 2010) express the wave-induced viscosity (or diffusivity) analytically as a function of the wavenumber spectrum, which can be obtained from a third-generation wavenumber spectrum numerical model (Yuan et al. 1991; Yang et al. 2005). This approach allows the coupling of a wave model with ocean circulation models. Numerous model simulations using this approach show that the additional nonbreaking wave-induced mixing significantly improves the model ML and SST when compared with model runs without the wave effects (Qiao et al. 2004, 2010; Lü et al. 2006; Qiao et al. 2006; Xia et al. 2006; Song et al. 2007; Wang et al. 2010; Shu et al. 2011).

The details of the mathematical derivation of the nonbreaking wave-induced mixing can be found in Qiao et al. (2010). Here we briefly introduce the equations to express such mixing in ocean circulation models. By splitting the velocity fluctuation $u$ into its current-related part $c$ and wave-related part $w$ (Yuan et al. 1999)—that is,

$$ u_i = u_{iw} + u_{ic}, \quad (1) $$

where $i = 1, 2, 3$ represents zonal, meridional, and vertical directions, respectively—we invoke Prandtl mixing length theory to parameterize the momentum mixing induced by the nonbreaking surface wave, and we assume that the mixing length $l_{iw}$ be proportional to the range of the particle displacement in the $i$th direction. Conceptually $u'_{iw}$ should be understood as the increment of the wave motion velocity at the spatial interval of $l_{iw}$ in the $i$th direction, and so can be expressed as

$$ u'_{iw} = l_{iw} \frac{\partial}{\partial x_i} (u_{iw} u_{iw})^{1/2}, \quad (2) $$

where

$$ l_{iw} \sim \int \frac{A(k) \exp(kz) \exp[i(k \cdot x - \omega t)]}{k} dk. \quad (3) $$

Since ocean waves are locally uniform, the horizontal change of statistic parameters for ocean waves within the length of $l_{iw}$ is nearly zero. Therefore,

$$ u'_{1w} = 0, \quad u'_{2w} = 0. \quad (4) $$

For the vertical direction, however,

$$ u'_{3w} = l_{iw} \frac{\partial}{\partial z} \left[ \int \frac{\omega^2 E(k) \exp(2kz)}{k} dk \right]^{1/2}. \quad (5) $$

The analytical expression of the nonbreaking surface wave-induced mixing $B_v$ is expressed as

$$ B_v = \alpha \left[ \int \frac{E(k) sh^2[k(H + z)]}{sh^2(kH)} dk \right]^{1/2} \times \frac{\partial}{\partial z} \left\{ \int \frac{\omega^2 sh^2[k(H + z)]}{sh^2(kH)} E(k) dk \right\}. \quad (6) $$

### 3. Model and experiment design

As mentioned in the introduction, the purpose of this study is to demonstrate that the improvement of ML simulation can significantly improve the simulation of SASM in a coupled climate model. The improvement of ML simulation is realized by adding the nonbreaking wave-induced vertical mixing, the so-called $B_v$, scheme developed by Qiao et al. (2004, 2010), into the coupled climate model.

The coupled GCM used in this study is the Institute of Atmospheric Physics (IAP), Chinese Academy of Sciences–State Key Laboratory of Numerical Modeling of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG) model, called the Flexible Coupled Ocean–Atmosphere General Circulation Model (FGCM), which is built on the first version of the National Center for Atmospheric Research (NCAR) Climate System Model (CSM-1). The most significant difference between CSM-1 and FGCM-0 is the ocean component of CSM-1 being replaced by the IAP–LASG global ocean model, which has 30 vertical layers and a horizontal resolution of T63 ($1.875^\circ \times 1.875^\circ$) and is therefore often called L30T63. To make this replacement work, some necessary modifications to other component models of CSM-1 are also carried out. More details about FGCM-0 can be found in Yu et al. (2002).

In L30T63, the Pacanowski and Philander scheme (Pacanowski and Philander 1981) is employed to calculate the vertical turbulence mixing. In this scheme, the mixing coefficient is determined by the vertical shear of mean current and the density gradient of seawater. The sharp and abrupt change of the ocean vertical mixing rate renders the FGCM-0 unable to reproduce the structure of the upper ocean. To correct this, we incorporate the $B_v$ into FGCM-0. Since $B_v$ involves surface wave spectra, the Marine Science and Numerical Modeling (MASNUM) wave model (Yuan et al. 1991; Yang et al. 2005) is used to calculate the nonbreaking wave-induced mixing $B_v$, we then include $B_v$ in the ocean circulation model with data transfer through a coupler.
Two numerical experiments are designed to evaluate the effect of $B_v$ on the SASM. Experiment 1 is a direct run of the FGCM-0 without $B_v$ (denoted Exp.no), while Experiment 2 adds $B_v$ to FGCM-0 (denoted Exp.with). After spinning up the L30T63 ocean circulation model for 1000 yr, we run both numerical experiments of Exp.no and Exp.with for an extra 70 yr to reach their respective equilibrium, and the last 50 yr (years 21–70) are analyzed.

The observation data used for comparison in this study are as follows: 1) National Centers for Environmental Prediction (NCEP) reanalysis data provided by the National Oceanic and Atmospheric Administration Office of Oceanic and Atmospheric Research (NOAA/OAR) Earth System Research Laboratory Physical Sciences Division (ESRL PSD) (see http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.derived.html) from 1948 to 2011 (Kalnay et al. 1996). The variables that are compared include 1) wind speed, specific humidity, and sea level pressure; 2) the Climate Prediction Center (CPC) Merged Analyzed of Precipitation (CMAP) data from 1979 to 2008 (Xie and Arkin 1997); and 3) seawater temperature data provided by NOAA National Oceanographic Data Center (NODC) World Ocean Atlas (Levitus) data (Locarnini et al. 2010).

4. Analysis of model results

Figure 1 shows the difference between the simulated precipitation and CMAP data over a 50-yr period for summer months from June to August (JJA). Spatial distribution of precipitation in the SASM region is simulated quite poor in Exp.no, with negative bias (insufficient rainfall) extending from the Northwest Pacific through the South China Sea to the Bay of Bengal, and positive bias (excessive rainfall) from the Arabian Sea to the southern Indian Ocean (Fig. 1a). In Exp.with, the model has an improved performance in simulating the patterns of precipitation. The simulation errors are significantly reduced by inducing $B_v$ in the model. As shown in Fig. 1b, the bias over South China, the Bay of Bengal, the northwestern Pacific, and the southern Indian Ocean has been reduced by 2 mm day$^{-1}$. However, the inclusion of $B_v$ has not completely eliminated the errors, especially in the western Indian Ocean. The domain-averaged root-mean-square error (RMSE) of precipitation of two simulations against CMAP data is decreased from 3.23 in Exp.no to 2.36 in Exp.with, suggesting that $B_v$ is a major factor responsible for the improvement of the SASM simulation in FGCM-0.

Water vapor transport and convergence provide good indicators of latent heat release associated with SASM. Since the upper atmosphere contains a relatively small amount of water vapor (Zhou and Yu 2005), our water vapor flux is based on the sum of water vapor contained between surface to 300 hPa, as expressed below:

$$Q = \frac{1}{g} \int_{P_s}^{P_t} qV dp, \quad (7)$$

where $Q$ is water vapor flux, $q$ and $V$ represent the specific humidity and the velocity vector, $g$ is the acceleration of gravity, and $P_t$ and $P_s$ are 300 hPa and surface pressure, respectively.

Figure 2 displays the distributions of averaged JJA water vapor transport in the SASM region based on NCEP reanalysis data and model outputs. A characteristic
pattern of JJA water vapor transport in the SASM region is a conveyor belt (Fig. 2a): water vapor is transported from east to west in the southern Indian Ocean, then northward across the equator following the Somali jet in the western Indian Ocean, and gradually turns eastward in the northern Indian Ocean, across the India subcontinent to the Bay of Bengal and the South China Sea. Eventually, this conveyor belt meets the Pacific subtropical high and turns northward. The water vapor flux in the Arabian Sea and the Bay of Bengal is strongest, with the intensity reduced over the South China Sea. Figure 2b shows the water vapor transport simulated by FGCM-0. The intensity of the water vapor conveyor belt in general is much weaker in original FGCM-0 compared with NCEP reanalysis data; it reaches a maximum near the coast of Somalia and weakens rapidly from the South Asian monsoon region to the East Asian monsoon region. In the eastern Arabian Sea, water vapor flux reduces to 200 kg m$^{-1}$ s$^{-1}$ and the belt structure of water vapor transport become difficult to identify. Figure 2c displays the difference of water vapor flux between Exp.no and Exp.with, indicating the effect of $B_v$ on the modeling of water vapor transport. It shows that water vapor transport is significantly enhanced over the eastern Arabian Sea, the Bay of Bengal, and the South China Sea. The most noticeable difference in water vapor transport between Exp.no and Exp.with is found in the Bay of Bengal, where water vapor flux features the largest increase. Spatial distribution of water vapor transport simulated in Exp.with (Fig. 2d) appears to be very consistent with observations (Fig. 2a).

So far we have shown the effect of incorporating $B_v$ into FGCM-0 in improving the simulation of SASM. The question that naturally follows is how and why this improvement is accomplished. To answer this question we examine the ocean responses to $B_v$.

We use the depth of 20°C thermocline of observation to reflect the JJA climatology of thermocline structure in the Indian Ocean, and it also can be regarded as the bottom of the upper mixed layer. As shown in Fig. 3a, the thermocline is deeper in the Arabian Sea, the northwestern Pacific Ocean, and at 15°-20°S in the southern

FIG. 2. The JJA mean water vapor transport (kg m$^{-1}$ s$^{-1}$): (a) NCEP reanalysis data results, (b) Exp.no simulation, (c) Exp.with minus Exp.no, and (d) Exp.with simulation.
Indian Ocean. The minimum depth appears in the tropical southern Indian Ocean from the equator to about 10°S. The depth of 20°C thermocline simulated by Exp.no is smaller in the southern Indian Ocean compared with the Levitus data, indicating that the ML is too shallow and the strength of summer thermocline is underestimated by FGCM-0 (Fig. 3b).

Parameter $B_y$, expressed in terms of wavenumber spectrum, has spatial nonuniform distribution (Qiao et al. 2004), which is shown in Fig. 4. Its value is larger in the southern Indian Ocean and the Arabian Sea with a maximum larger than $6 \times 10^{-4}$ m$^2$ s$^{-1}$. As indicated by Eq. (6), a large value of $B_y$ corresponds to stronger vertical mixing. In the tropical southern Indian Ocean, the region of large $B_y$ coincides with the region of relatively shallow thermocline, leading to large relative deepening of the ML and thereby reducing the sea surface temperature in that large domain, increasing the differential heating in the north–south direction (Fig. 5). The stronger low-level wind directly driven by the stronger differential heating is further amplified by the deep convective heating associated with the moisture convergence in the northern precipitation areas. This process is reflected in Fig. 2. It can be verified that the atmospheric moisture changes little in two different runs while the wind vector changes significantly (Fig. 2).

To further solidify the scenario discussed above, we diagnose the latent heat flux and surface pressure fields in two runs. Figure 6 shows the variation of latent heat flux induced by $B_y$, in which positive latent heat flux appears in the southern Indian Ocean, Arabian Sea, and the Bay of Bengal, reflecting stronger convective heating in these regions. Figure 7 shows the SLP difference between Exp.with and Exp.no. Up to a 4-hPa surface pressure difference is directly or indirectly caused by $B_y$ in the southern Indian Ocean from 80° to 100°E, and a band of negative surface pressure difference extends from the Arabian Sea to the Bay of Bengal and to the South China Sea, with the strongest anomaly depression (<−3 hPa).
appearing in the Bay of Bengal. These diagnoses clearly support the scenario described above.

The next question is how good the simulated monsoonal circulation is with the inclusion of $B_v$. Figure 8 compares the 500-hPa circulations among observations and the Exp.no, and Exp.with simulations. In the observation, atmospheric flows from the Indian Ocean, the Indonesian Sea, and the northwestern Pacific subtropical high converge in the South China Sea and then the converged flow turns northward. Exp.no simulates the location and strength of the subtropical high relatively well, but the subtropical high itself extends too far west, even reaching the Bay of Bengal, and so the position of the simulated Indian low stays far west to the north of Somalia. The corresponding simulation in Exp.with gives a much better result: the strength and location of the subtropical high and Indian low are quite consistent with the observation, if not perfect.

The consistency between the observation and the simulation in Exp.with can also be recognized in the meridional winds along the equator (Fig. 9). As displayed in Fig. 9a, the observed cross-equatorial flow in lower atmosphere is northward and strong in the western Indian Ocean, and the Somalia jet reaches its maximum strength at 40°E. This feature is reproduced in both Exp.no and Exp.with. However, the observed extended more northward in the upper troposphere from 40° to 160°E. This is reproduced quite well in Exp.with.

5. Conclusions

The present work has demonstrated that the inclusion of the nonbreaking wave-induced ocean mixing parameterized by parameter $B_v$ in the climate model of FGCM-0 can much improve the SASM simulation. The results from two experiments indicate that the improvement is mainly attributed to a better simulation of the horizontal structure and the upper ocean mixed layer, which leads to improved air–sea heat flux in the Indian Ocean and resulting better transport of moisture in atmosphere circulation model. The major results are summarized as follows.

Compared with the CMAP data, original FGCM-0 simulation overestimates the precipitation in the Arabian Sea, the southern Indian Ocean, and western China, while it underestimates monsoon rainfall in the northwestern Pacific, the South China Sea, and the Bay of Bengal. Adopting $B_v$ as an additional ocean vertical mixing parameter, the simulated domain-averaged RMSE of precipitation is decreased from 3.23 to 2.56. The differences of water vapor transport between two numerical experiments are evident over the east of the Arabian Sea and the Bay of Bengal characterized previously by insufficient water vapor flux. The model results with $B_v$ significantly enhance water vapor transport in this region, leading to better agreement with observations.

A quite realistic simulated result of ocean mixed layer depth, by lowering the southeastern Indian Ocean SST and raising SST in the northern Indian Ocean and the southwestern Indian Ocean, results in the improved meridional differential heating. In addition, the latent heat flux forced by SST increases in the southern Indian Ocean, the Arabian Sea, and the Bay of Bengal. Stronger convective heating appears with the adoption of $B_v$ in the climate model, which has an important influence on monsoon simulation. Our results indicate that the stronger low-level wind directly driven by the stronger differential heating is further amplified by the deep convective heating associated with the moisture convergence and precipitation, which leads to the improved vertical structure of monsoonal circulation. This implies that the latent heating plays a fundamental role in driving the upper–lower-level reversal of monsoon circulation of the right amplitude.
The consistency between the observations and the simulation with $B_v$ also appears in the simulation of the monsoonal circulation systems, such as the meridional winds along the equator, the subtropical high, and the Indian low system in the atmosphere. The observed northerly cross-equatorial flow in the upper troposphere is reproduced favorably in the climate model with $B_v$. As shown in 500-hPa isobaric chart, the strength and location of the subtropical high and the Indian low have been much improved when $B_v$ is incorporated into the climate model.

Although adoption of $B_v$ is unable to eliminate all the simulation bias by the climate model of FGCM-0, the thermal structure in the upper ocean is much improved, resulting in better simulation of SASM. It is of interest to investigate whether the inclusion of $B_v$ would improve monsoon simulation in other climate models.

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**FIG. 8.** The general circulation in 500 hPa: (a) NCEP reanalysis data, (b) Exp.no simulation, and (c) Exp.with simulation.

**FIG. 9.** JJA mean cross-equator flow depicted by the vertical profile of meridional wind (m s$^{-1}$) along the equator: (a) NCEP reanalysis data, (b) Exp.no simulation, and (c) Exp.with simulation.
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