Improving simulations of the upper-ocean by inclusion of surface waves in the Mellor-Yamada turbulence scheme

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Abstract

The Mellor-Yamada turbulence closure scheme, used in many ocean circulation models, is often blamed for overly high simulated surface temperature and overly low simulated subsurface temperature in summer due to insufficient vertical mixing. Surface waves can enhance turbulence kinetic energy and mixing of the upper ocean via wave breaking and non-breaking wave-turbulence interaction. The influences of wave breaking and wave-turbulence interaction on the Mellor-Yamada scheme and upper-ocean thermal structure are examined and compared with each other using one-dimensional and three-dimensional ocean circulation models. Model results show that the wave-turbulence interaction can effectively amend the problem of insufficient mixing in the classic Mellor-Yamada scheme. The behaviors of the Mellor-Yamada scheme, as well as the simulated upper-ocean thermal structure, are significantly improved by adding a turbulence kinetic energy production term associated with wave-turbulence interaction. In contrast, mixing associated with wave breaking alone seems insufficient to significantly improve the simulations since its effect is limited to the very near-surface layers. Therefore, it is concluded that the effects of wave-turbulence interaction on the upper ocean are much more important than the effects of wave breaking.
1. Introduction

Vertical mixing plays a significant role in the temperature structure and circulation of the world’s oceans. A proper parameterization of vertical mixing is essential for numerical ocean circulation models. Various vertical mixing schemes have been developed for different models; of those schemes, the turbulence closure scheme of Mellor and Yamada [1982; M-Y hereafter] has been used extensively in atmosphere and ocean numerical models.

Although models with the M-Y scheme can successfully reproduce many general features of the ocean, it does have some deficiencies, so various remedies have been proposed over the years to improve the scheme [Kantha and Clayson, 1994; Ezer, 2000; Mellor, 2001]. In particular, insufficient mixing in the upper ocean during summer is a common problem of ocean circulation models, no exception for models using M-Y scheme. The result of insufficient mixing is too high sea surface temperature [SST hereafter; Martin, 1985] and too low subsurface temperature [Ezer, 2000]. In fact, the M-Y scheme was developed for stratified boundary layers near rigid surfaces, in which the principal balance in the turbulence kinetic energy (TKE) equation is among the local shear production, buoyant production, and dissipation [Mellor and Yamada, 1982]. This assumption may be appropriate for atmosphere models, but not for ocean models if the presence of surface waves in the ocean is considered.

Surface waves at the air-sea interface play an important role in the TKE budget for the upper ocean. Wind energy input to surface waves is estimated as 60~70 TW [Wang and Huang, 2004; Rascle et al., 2008], which is much greater than the mechanical energy from all other sources in the ocean [Ferrari and Wunsch, 2009]. In order to simulate the upper ocean accurately, it is necessary to incorporate the effects of surface waves into the M-Y scheme, in the hope to improve the performance of ocean models.

Most of wave energy is locally dissipated through wave breaking [Donelan, 1998], which greatly enhances the TKE near the sea surface [Agrawal et al., 1992; Drennan
Wave breaking effects have been added to the original M-Y scheme by Mellor and Blumberg [2004] through modifying the surface boundary condition of the TKE equation. In Mellor and Blumberg [2004] and Sun et al. [2005], wave breaking improved somewhat the simulations of SST and mixed-layer depth (MLD) when compared with those by the original M-Y scheme, while D’Alessio et al. [1998], Burchard [2001] and Zhang et al. [2007] argued that wave breaking cannot affect the upper-ocean temperature because the strong turbulence induced by wave breaking is mainly confined within the near-surface zone with the depth scale of wave height [Rapp and Melville, 1990; Craig and Banner, 1994; Soloview and Lukas, 2003]. Thus, it remains a controversial issue over the years how much effect the wave breaking really have on ocean mixing.

Besides wave breaking, non-breaking surface waves can directly transfer energy from the surface wave to the turbulence because the surface waves are not truly irrotational or potential before breaking. This has been confirmed by theoretical analyses [Phillips, 1961; Kitaigorodskii et al., 1983; Arduin and Jenkins, 2006], laboratory experiments [Cheung and Street, 1988; Teixeira and Belcher, 2002; Babanin and Huas, 2009; Dai et al., 2010], and field observations [Anis and Moum, 1995; Gemmrich and Farmer, 2004; Veron et al., 2009]. The energy transfer associated with non-breaking surface waves is usually attributed to wave-turbulence interaction (or wave-current-turbulence interaction). Qiao et al. [2004, 2010] derived a parameterization scheme of non-breaking wave induced vertical mixing from the wave number spectrum, depicting the coupling of ocean models with wave models. Adopting this scheme to coastal and global circulation models [e.g., Qiao et al., 2004; Lin et al., 2006], as well as climate models [Song et al., 2007; Huang et al., 2008] resulted in significant improvements in the simulations of the upper ocean temperature structure. Other processes such as Langmuir cells, have also been tried to be incorporated into ocean models. The results show that these processes can affect the SST and MLD [D’Alessio et al., 1998; Kantha and Clayson, 2004].

Recently, regarding the wave-turbulence interaction as an additional source of TKE, Huang and Qiao [2010] introduced a parameterization scheme of
wave-turbulence interaction, and incorporated the scheme to a one-dimensional model with the M-Y scheme. The modified model shows significant improvements in simulating turbulence characteristics compared with the original model. In this paper, we further analyze the influences of wave-turbulence interaction on the M-Y scheme and upper-ocean simulation, along with comparison of the wave-breaking effect. Section 2 will discuss model modifications related to wave breaking and wave-turbulence interaction. Effects of wave breaking and wave-turbulence interaction on one-dimensional models are compared using idealized forcing experiments in Section 3 and ocean station data in Section 4. In Section 5, the wave-turbulence interaction effects are further discussed by using a three-dimensional model. Section 6 gives conclusions of this study.

2. Inclusion of wave-turbulence interaction and wave breaking

In the classic level-2.5 version of the M-Y turbulence closure model [Mellor and Yamada, 1982], the vertical viscosity $K_m$ and diffusivity $K_h$ are calculated from

$$K_m = q l S_m,$$  \hspace{1cm} (1a)  

$$K_h = q l S_h,$$  \hspace{1cm} (1b)  

where $q^2/2$ is the TKE, $l$ is a turbulence length scale, and $S_m$ and $S_h$ are stability functions associated with the Richardson number, which is defined as $G_h \equiv -l^2 N^2/q^2$. Two prognostic equations are solved for $q^2$ and $l$ as follows

$$\frac{D}{Dt} \left( \frac{q^2}{2} \right) - \frac{\partial}{\partial z} \left[ K_q \frac{\partial}{\partial z} \left( \frac{q^2}{2} \right) \right] = P_s + P_b - \varepsilon,$$  \hspace{1cm} (2a)  

$$\frac{Dq^2 l}{Dt} - \frac{\partial}{\partial z} \left[ K_q \frac{\partial q^2 l}{\partial z} \right] = E_1 l [P_s + E_3 P_b] - l \varepsilon \tilde{W},$$  \hspace{1cm} (2b)  

where $K_q$ is the vertical turbulence diffusivity, $P_b$ is the buoyant production, $\tilde{W}$ is a wall proximity function, $E_1$ and $E_3$ are non-dimensional constants,

$$P_s = K_m \left[ \left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2 \right]$$  \hspace{1cm} (3)  

is the shear production of TKE,
\[ \varepsilon = \frac{q^3}{B_1 l} \]  \hspace{1cm} (4)

is the TKE dissipation rate, \( u \) and \( v \) are horizontal velocity components, and \( B_1 (=16.6) \) is an empirical constant.

In the classic M-Y model, the surface boundary conditions for these two prognostic equations are

\[ q^2(0) = B_1^{2/3} u_*^2, \]  \hspace{1cm} (5a)
\[ q^2 l(0) = 0. \]  \hspace{1cm} (5b)

The wave breaking effect is usually incorporated into the M-Y scheme by modifying Eqs. (5a) and (5b). Mellor and Blumberg [2004] incorporated the wave breaking parameterization obtained by Craig and Banner [1994] into the M-Y scheme, in which the boundary conditions were changed to

\[ q^2(0) = (15.8 \alpha \beta_{CB})^{2/3} u_*^2, \]  \hspace{1cm} (6a)
\[ l(0) = \kappa z_w u_*^2 / g, \]  \hspace{1cm} (6b)

where \( \kappa = 0.41 \) is the von Karman constant, \( z_w \) is a surface roughness length, \( \alpha_{CB} (=100) \), and \( \beta_{CB} (=2\times10^5) \) are empirical constants; \( u_* = \sqrt{\tau_0 / \rho} \) is the friction velocity on the sea surface, here \( \tau_0 \) is the surface wind stress, and \( \rho \) is the density of sea water.

Recently, Huang and Qiao (2010) introduced a parameterization of the TKE dissipation rate induced by wave-turbulence interaction as

\[ \varepsilon_w = 148 \beta \sqrt{\delta} \frac{u_{d0} u_*^2}{L} e^{2k z}, \]  \hspace{1cm} (7)

where \( \beta \) is a dimensionless constant that is usually set to 1.0. \( \delta (=H_s/L) \) is the wave steepness, here \( H_s \) is the significant wave height, and \( L \) is the wavelength, \( u_{d0} = c(Ak)^2 \) is the magnitude of the Stokes drift at the surface (or at \( z=0 \)), here \( c \) is the wave phase velocity, \( A (=H_s/2) \) is the wave amplitude, and \( k (=2\pi/L) \) is the wave number, and \( \delta \) is the vertical coordinate with zero at the mean sea level and positive upward.

In this parameterization, the TKE dissipation rate induced by wave-turbulence interaction is a function of \( u_{d0} u_*^2 \) and wave parameters, and it decays with the depth.
away from the surface in the form of $e^{2kz}$. Since wave-turbulence interaction transfers energy from wave fields to the turbulence, it is reasonable for the TKE production in the M-Y scheme to include its effect when surface waves are present, i.e.,

$$P_s = K_m \left[ \left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2 + \mathcal{E}_w \right].$$

The influences of wave-turbulence interaction and wave breaking on the M-Y scheme and the upper-ocean simulation will be examined in the following sections by using the results of sensitivity numerical experiments. Each group of numerical experiments include three cases (Table 1): the first case is based on the classic/original M-Y scheme, in which effects of surface waves are not included; the second case includes wave-breaking effects via modifying the surface TKE flux and roughness length as in Eqs. (6a) and (6b); and the third case includes the effects of wave-turbulence interaction as in Eq. (8).

### 3. One-dimensional experiments with idealized forcing

In this section, we use a one-dimensional model to examine the behavior of the M-Y scheme by including wave breaking or wave-turbulence interaction, respectively. This model is similar to that used by Mellor and Blumberg [2004]. Following Martin [1985] and D’Alessio et al. [1998], several groups of experiments are carried out with idealized forcing, including wind-deepening (Exps. W1-W3), heating (Exps. H1-H3), and cooling (Exps. C1-C3). In those experiments, the water depth is set to 200 m with a model vertical resolution of 2 m. The time step is 120 s, and the latitude is taken to be 29.91°N, where the inertial period is 24 h.

The significant wave height $H_s$, the period $T_s$, and the surface wind stress $\tau_0$ are used for calculating the dissipation rate induced by wave-turbulence interaction via Eq. (7). Following Kantha and Clayson [2004], we set $H_s = 1.6$ m and $T_s = 6.2$ s, which is approximately corresponding to a fully developed sea state under the wind stress of 0.1 N m$^{-2}$.

### 3.1 Wind-deepening
For this group of numerical experiments, the initial conditions are as follows. The SST is equal to 24°C and thermal stratification is a uniform 0.05°C m⁻¹; the salinity is constant of 35 psu, and the velocity is zero. The surface wind stress is set to be 0.1 N m⁻², and the surface heat flux and freshwater flux are zero.

Figure 1 shows the evolution of the SST in the Exps. of W1 (control run), W2 (wave breaking), and W3 (wave-turbulence interaction). One can see that the SST in Exp. W2 is almost the same as that in Exp. W1, while it decreases significantly in Exp. W3. The changes in SST are attributed to changes in turbulence characteristics in the experiments. In Exp. W2, the enhanced turbulence velocity $q$ by wave breaking is limited to the first few meters near the surface (Fig. 2c). In the M-Y scheme, the dissipation rate $\varepsilon$ and the vertical diffusivity $K_h$ are closely associated with $q$, then the high values of the simulated $\varepsilon$ and $K_h$ are also confined to the near-surface zone (Figs. 2d, f). Thus, if the MLD is deeper than this wave breaking affected zone, wave breaking has little effect on either mixed-layer deepening or the upper thermal structure (Fig. 2a).

In Exp. W3, however, the enhanced $q$ by wave-turbulence interaction extends into much greater depth compared to that in Exp. W2 (Fig. 2c), although $l$ shows little change (Fig. 2b). This change in $q$ enhances the simulated $\varepsilon$ significantly (Fig. 2d). At the same time, they can cause an increase of the stability factor $S_h$ via changing the Richardson number $G_h$ (Fig. 2e). In the M-Y scheme, the vertical diffusivity $K_h$ is proportional to $q$, $l$, and $S_h$. Thus the simulated $K_h$ in Exp. W3 is much larger than that in Exp. W1 (Fig. 2f), which results in a decrease in the SST (Fig. 1) and an increase in the mixed layer depth (Fig. 2a).

### 3.2 Heating

For the heating experiments, the initial conditions and model parameters are set the same as those in the wind-deepening experiments, except that the initial temperature from the surface to 100 m is a uniform 19°C, and each experiment is forced with a constant wind stress of 0.1 N m⁻² plus a heat flux of 290.4 W m⁻² (i.e., 600 ly day⁻¹).
Figure 3 depicts the model turbulence characteristics on Day 2 in a similar fashion as shown in Fig. 2 for Exps. W1-W3. In Exp. H1 (control run) and Exp. H2 (wave breaking), there are large and unrealistic diffusivities below 25 m (Fig. 3f). The possible reason is that their stratifications are close enough to the threshold of instability, then generating too large stability factor $S_h$ [Ezer, 2000].

The temperature near the surface decreases slightly when the wave breaking effect is considered (Fig. 3a). The effects of wave breaking are quite sensitive to the magnitude of wind stress and the property of the mixed layer. Two additional groups of experiments are conducted in order to examine the wind stress effect. Under a weak wind condition (Exps. Ha1-Ha3), the mixed layer is very shallow, so that wave breaking plays a role in regulating the temperature near the surface (Fig. 4a). For example, the simulated SST in Exp. Ha2 (wave breaking) is about 0.3 °C lower than that in Exp. Ha1 (control run) on Day 2. Under strong winds, the wind-driven mixed layer is much deeper than the wave height, so that the wave-breaking effects are negligible (Fig. 4b). On the other hand, if the mixed layer is weakly stratified under a strong surface heating condition, wave breaking can still affect the SST (Fig. 3a).

The effects of wave-turbulence interaction are also dependent on the wind stress magnitude. The TKE production induced by wave-turbulence interaction is a function of $u_s u_z^c$, and decays with the depth away from the surface in the form of $e^{2kz}$ as in Eq. (7). Under the weak wind condition, wave-turbulence interaction plays an important role in regulating the TKE budget and vertical mixing of the upper ocean (Fig. 5a). The simulated SST in Exp. Ha3 (wave-turbulence interaction) is about 1.5 °C lower than that in Exp. Ha1 on Day 2. Under high wind conditions, however, the shear production of TKE is much larger than that by wave-turbulence interaction (Fig. 5b), so that wave-turbulence interaction has relatively little effect on the temperature in the upper ocean (Fig. 4b).

3.3 Cooling

For the cooling experiments, the initial conditions and model parameters are the
same as that for the wind-deepening experiments. The models were forced by a constant wind stress of 0.1 N m$^{-2}$ plus a heat flux of -96.8 W m$^{-2}$ (i.e., -200 ly day$^{-1}$).

Figure 6 shows the model results of Day 5. It is obvious that the inclusion of wave-turbulence interaction increases the simulated $q$ and $\varepsilon$ throughout large depth (Figs. 6c,d). The Richardson number $G_h$ is positive in the unstable stratification, so that the increase of $q$ can decrease $G_h$, and the stability factor $S_h$ (Fig. 6e). Thus the effects of wave-turbulence interaction on the vertical diffusivity $K_h$ and on the upper-ocean thermal structure are insignificant under unstable condition, since convective mixing dominates over all other processes (Figs. 6a, f).

4. One-dimensional experiments for ocean weather station Papa

In this section, the model is tested against observational data from June 9, 2007 to January 16, 2008 taken from the ocean weather station Papa, which is located in the eastern North Pacific at 145°W 50°N. Data from this station had served over the years as a measuring stick to evaluate various one-dimensional mixing models [Martin, 1985; Mellor, 2001; Mellor and Blumberg, 2004].

4.1 Model description

The model is the same as that used in Section 3. The initial temperature and salinity are taken from that on June 9, 2007, and linearly interpolated to the model grid; the initial velocity is set to zero. Hourly surface data are used to force the numerical model. The downward solar radiation and long wave radiation are taken from the observed 2-minute high resolution data. The rain data are taken from observed 10-minute data. The sea surface albedo is taken from Payne [1972], and the seawater optical type is assumed to be Type II [Paulson and Simpson, 1977]. Upward long wave radiation is obtained from the Stefan-Boltzmann law. The surface wind stress, sensible heat flux, and latent heat flux are obtained from the observed wind speed (Fig.7a), SST, air temperature, and relative humidity using the COARE algorithm [Fairall et al., 2003]. Sea-surface pressure data are not available, so pressure
is simply set to 101.325 kPa. A net surface heat imbalance was found at this site, so 15 W m\(^{-2}\) is added to the outgoing heat flux to decrease the simulated deviation [Large et al., 1994]. In addition, the SST and sea surface salinity (SSS) are relaxed to the observations, with relaxations of 50 W m\(^{-2}\) K\(^{-1}\) for SST and 30 days for SSS (assuming a mixed layer depth of 25 m). The Coriolis parameter \(f\) is for latitude of 50°N, and the background mixing is set to 2×10\(^{-6}\) m\(^2\) s\(^{-1}\) [Martin, 1985].

Three cases are carried out: with and without wave breaking, and with wave-turbulence interaction (see Table 1). Surface wave characteristics are not available in the observations. Instead we use the outputs from the WAVEWATCH III model [Tolman, 2002], as shown in Figs 7b, and 7c. The significant wave period is calculated by a linear formula \(T_s=0.91T_p\) [Wen et al., 1989], here \(T_p\) is the peak wave period.

### 4.2 Results

Figure 8 shows the simulated SST and MLD. The MLD is defined as the depth at which the temperature drops by 0.5°C from the surface. In Exp. P1 (control run), the simulated SST from June to October is warmer while the MLD is shallower than the observations; this is a common problem of the classic M-Y scheme caused by insufficient upper-ocean mixing as mentioned before [Martin, 1985; Kantha and Clayson, 1994; Ezer, 2000]. In Exp. P2 (wave breaking), the changes in SST and MLD are small compared to Exp. P1. The results in Exp. P3 (with wave-turbulence interaction), on the other hand, are in very good agreement with the observations in terms of the simulated SST and MLD. From November to January there is a clear discrepancy between the model and observations; possible reasons for that could be the absence of advective processes in one-dimensional models, and errors in the surface heat fluxes [Large et al., 1994].

Too cold model subsurface temperature is another problem of the classic M-Y scheme, which is usually associated with a too shallow summertime thermocline [Ezer, 2000]. Figure 9 shows simulated and observed daily-mean temperature profiles in the upper ocean. The simulated seasonal thermocline in Exp. P1 is somewhat shallower
and sharper than observed (Figs. 9a, 9d), so that the subsurface temperature is
significantly underestimated with a maximum bias of up to -4°C (Fig. 10a).

At Station Papa, the shallowest MLD is about 20 m, which is much deeper than
that achieved by wave breaking. On the other hand, the temperature in the mixed layer
is nearly uniform (Fig. 9). As discussed in Section 3, wave breaking alone has very
little effect on the upper-ocean thermal structure under these conditions. Thus the
simulated temperature profile and its deviation from observations in Exp. P2 are quite
similar to those in Exp. P1 (Figs. 9b and 10b). As seen before, wave-turbulence
interaction can affect greater depth than wave breaking, and thus the simulated
seasonal thermocline is more realistic in Exp. P3 (Fig. 9c), with maximum
temperature error of only 1°C (Fig. 10c).

These improvements of the upper-ocean thermal structure can also be seen from
Fig. 11, which shows comparison of simulated and observed daily-mean temperature
profiles on September 1, 2007 and October 1, 2007, respectively. In Exp. P1, the
simulated surface temperature is overestimated, while the subsurface temperature is
underestimated. Meanwhile the MLD is about 4-10 m shallower than the observations.
These biases are attributed to insufficient mixing (Fig. 12), which prevents the heat
transport from the surface layer to the subsurface. A large amount of heat accumulates
in the surface, which results in the simulated temperature being too warm in the
surface, while too cold in the subsurface.

Wave breaking can greatly enhance the turbulence, therefore the vertical mixing.
However, the wave-breaking effect is limited to the near-surface layer, so that it is
insignificant in regions with a deep MLD. The TKE dissipation rate induced by
wave-turbulence interaction, however, decays in the form of $e^{-2kz}$ away from the
surface, so that it can affect the depth much greater than that of wave breaking,
especially for the long swells with small wave numbers. The observations also show
that strong turbulence can extend to tens of meters in the ocean when swells are
present [Anis and Moum, 1995]. Moreover, the swells can radiate away from their
generation areas, and propagate over a large distance across an ocean basin.
Wave-turbulence interaction should be a non-negligible TKE source in regions where
the wind is weak but the swell is strong.

In our experiments, when wave-turbulence interaction is incorporated into the model, the simulated vertical mixing is greatly enhanced (Fig. 12). Strong mixing can transport a large amount of heat from the surface layer to the subsurface, resulting in a decrease in the surface temperature and an increase in subsurface temperature (Fig. 11). Thus, the temperature distribution of the upper ocean is improved greatly in Exp. P3 due to the effect of wave-turbulence interaction and the simulated temperature is close to the observed temperature.

5. Global ocean general circulation model

The behavior of the M-Y scheme in three-dimensional models is sometimes different from that in one-dimensional models due to horizontal advection and other three-dimensional processes [e.g., Mellor, 2001, discussed some of these limitations]. In this section, we examine effects of the wave-turbulence interaction and wave breaking on upper-ocean simulations by the Princeton Ocean Model [POM; Blumberg and Mellor, 1987].

5.1 Model description

The model domain covers the global ocean between 72°S and 65°N with a zonal resolution of 1°. The meridional resolution is 1/3° within 10°S-10°N, and increases linearly to 1° by 20°N (20°S). The model has 32 sigma layers in the vertical direction, with at least 6 layers in the top 60 m and at least 10 additional levels between 60 and 250 m. The bottom topography is obtained from ETOPO5, and the maximum water depth is set as 5000 m.

The model is initiated with the temperature and salinity conditions in January taken from the WOA01 climatology [Conkright and Boyer, 2002]. The climatological monthly-mean wind stress is taken from the QuickSCAT/NCEP blended ocean winds from 2000 to 2005 [Milliff et al., 2004]. The surface heat flux and freshwater flux are taken from the monthly-mean COADS climatology [da Silva et al., 1994]. The penetration of the solar radiation is taken into account, and the seawater optical type is
taken as Type I [Paulson and Simpson, 1977]. Moreover, to keep surface density values from drifting too far from climatology, the SST and SSS are relaxed to the monthly-mean climatology of WOA01, with a relaxations of 50 W m⁻² K⁻¹ for SST and 60 days for SSS (for a mixed layer depth of 25 m). Sponge layers of 5° are placed along the northern and southern boundaries, in which both temperature and salinity are relaxed toward observed monthly climatological fields.

The horizontal viscosity and diffusion are calculated by a Smagorinsky type formula [Smagorinsky et al., 1963], but the minimum horizontal viscosity is set to 2×10³ m² s⁻¹ [Pezzi and Richards, 2003]. A time-splitting scheme is used, with a barotropic time step (for the vertically averaged equations) of 60 s and a baroclinic time step (for the 3D equations) of 1200 s. The vertical viscosity and diffusivity are calculated from the level-2.5 turbulence closure scheme [Mellor and Yamada, 1982], with background coefficients of 1×10⁻⁴ and 1×10⁻⁵ m² s⁻¹, respectively. Three experiments (Exps. G1-G3) are carried out (Table 1). The three numerical experiments are run for 10 years from rest (cold start), in which the outputs of the last year are used for the analysis. Similar as Exp. P3, surface wave characteristics in Exp. G3 are also obtained from the outputs from the WAVEWATCH III model [Tolman, 2002].

5.2 Results

The most outstanding features induced by the wave breaking and wave-turbulence interaction are improvements of the subsurface thermal structure during summer. Figure 13 shows monthly-mean temperature differences along the dateline in August of the three experiments relative to the WOA01 data. In Exp. G1 (control run), the simulated subsurface temperature between 20°N and 60°N is seriously underestimated by insufficient mixing as compared to the observations, with a maximum bias up to 3°C (Fig. 13a). The temperature biases in Exp. G2 (wave breaking) are quite similar to that in Exp. G1, but their amplitudes seem to be reduced slightly due to the enhanced mixing of wave breaking (Fig. 13b). Although it is believed that the mixing induced by wave breaking is mainly limited to the upper few
meters, its effect can extend to a greater depth in the three-dimensional model, compared with the more limited impact seen in the one-dimensional model discussed before.

When the wave-turbulence interaction is incorporated into the 3D model, the simulation of the upper-ocean temperature is significantly improved. The biases are less than 1°C in most areas. The improvements can also be seen from Fig. 14, which shows monthly-mean model temperature errors along 30°N in August for the three experiments.

The improvements in Exps. G2 and G3 are mainly located in middle and high latitudes. In the low latitudes, the influences of wave breaking and wave-turbulence interaction are less important. This is attributed to the fact that wave energy is smaller in tropical areas [Wang and Huang, 2004]. The other reason may be due to the fact that the turbulence production in the ocean mixed layer is dominated by TKE flux from the sea surface in the extra-tropical ocean, while it is dominated by shear production in the equatorial ocean [Noh, 2004].

Figure 15 shows the temperature profiles at the central point of the North Pacific (30°N, 180°E). In summer, the seasonal thermocline usually appears under the mixed layer at middle and high latitudes due to strong solar radiation and weak wind, while in winter, the mixed layer deepens and the seasonal thermocline disappears due to strong cooling-induced convection. The too shallow summer thermocline associated with insufficient mixing is a common problem of ocean general circulation models with the M-Y scheme. Ezer [2000] suggested that shortwave radiation penetration can improve the simulation, which can then reproduce a more realistic summer thermocline. However, the thermocline depth is still underestimated to some extent in his model. Similar to his study, the seasonal thermocline is somewhat shallow in Exp. G1, although the model has included shortwave radiation penetration effect. This shallow thermocline results in the simulated temperature in summer being seriously underestimated in the subsurface with a maximum bias of 3.5°C, while overestimated in the surface (Fig. 16a).

In Exp. G2, the simulated thermocline seems to be slightly improved due to
wave-breaking effect (Fig. 15b), and the temperature biases decrease slightly compared to that in Exp. G1 (Fig. 16b). In Exp. G3, the seasonal thermocline is greatly improved by including the wave-turbulence interaction (Fig. 15c), and the maximum deviation from the climatology decreases to 2.0°C (Fig. 16c). This indicates that other improvements, such as more realistic surface forcing [Ezer, 2000], a Richardson number dependent dissipation correction [Mellor, 2001], and internal wave breaking [Kantha and Clayson, 1994], may also be needed in order to simulate the upper ocean accurately.

6. Conclusions

Wind energy input into ocean surface waves is a dominate source of external mechanical energy for the ocean [Ferrari and Wunsch, 2009]. Surface waves can transfer TKE from wave fields to the turbulence via wave breaking and wave-turbulence interaction. This energy transfer exerts important influence on the TKE budget of the upper ocean. In this study the effects of wave breaking and wave-turbulence interaction on the M-Y scheme and upper-ocean simulations are examined and compared with each other using one-dimensional and three-dimensional ocean circulation models.

The classic M-Y scheme is obtained for stratified boundary layers near rigid surfaces, in which the effects of surface waves are missing [Mellor and Yamada, 1982]. Although the scheme can successfully reproduce the general features of the ocean, there are usually systematic deficiencies due to insufficient mixing in the upper ocean: the simulated temperature in summer is often too high near the surface, while too low in the subsurface associated with too shallow MLD and seasonal thermocline [Martin, 1985; Ezer, 2000].

Although a large amount of wave energy is lost via breaking, its induced turbulence and mixing are mainly limited to the upper few meters. Furthermore, to account for surface wave breaking in ocean models will require an extremely fine resolution near the surface [Mellor and Blumberg, 2004]; such high resolution is not possible in most general ocean circulation models. In regions with weak wind and
strong surface heating, the mixed layer is usually shallow with weak stratification. Wave breaking may play a role in the behavior of the temperature near the surface, but its effects are very confined. Under moderate and high wind conditions, the mixed layer is usually much deeper than the depth affected by wave breaking, so that wave breaking effects are negligible.

Wave-turbulence interaction can affect much greater depths than wave breaking. Wave-turbulence interaction can be incorporated into the M-Y scheme as an additional TKE production, and other empirical constants need not to be adjusted. Our results show that wave-turbulence interaction can effectively amend the problem of insufficient mixing in the classic M-Y scheme. In the models with improved M-Y scheme, where wave-turbulence interaction is added, the behavior of the M-Y scheme, as well as the simulated upper-ocean thermal structure, are significantly improved as compared to that by the classic M-Y scheme. In summary, our results clearly demonstrate that the effects of wave-turbulence interaction on the upper ocean is in most cases more important than that of wave breaking; these results are due to two main reasons: 1. the large depth affected by wave-turbulence interaction and 2. the impact of non-locally generated swell in the wave-turbulence interaction.
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Figure captions:

Figure 1. Simulated SST for Exps. W1-W3.

Figure 2. Comparisons of simulated (a) temperature profile $T$, (b) turbulence length scale $l$, (c) turbulence velocity magnitude $q$, (d) dissipation rate $e$, (e) stability factor $S_h$, and (f) vertical diffusivity $K_h$ for Exps. W1-W3 on Day 5.

Figure 3. Same as Fig. 2, except for Exps. H1-H3 on Day 2.

Figure 4. Same as Fig. 3a, except using surface wind stress of (a) 0.037 N m$^{-2}$, and (b) 0.2 N m$^{-2}$.

Figure 5. TKE productions caused by the velocity shear (blue lines) and wave-turbulence interaction (pink lines) on Day 2 for (a) Exp. Ha3, and (b) Exp. Hb3.

Figure 6. Same as Fig. 2, except for Exps. C1-C3 on Day 5.

Figure 7. (a) Wind speed, (b) significant wave height, and (c) period at 50°N, 145°W from June 9, 2007 to January 16, 2008. The surface wave data were obtained from the WAVEWATCH III model.

Figure 8. Simulated and observed daily-mean (a) SST, and (b) MLD at Station Papa from June 9, 2007 to January 16, 2008. The MLD is defined as the depth at which the temperature drops by 0.5°C from the surface.

Figure 9. Simulated and observed daily-mean temperature profiles (units: °C) from (a) Exp. P1, (b) Exp. P2, (c) Exp. P3, and (d) observations.

Figure 10. Simulated daily-mean temperature deviations from the observation (units: °C): (a) Exp. P1, (b) Exp. P2, (c) Exp. P3.

Figure 11. Simulated and observed daily-mean temperature profiles at Station Papa (units: °C) on (a) September 1, 2007 and (b) October 1, 2007.

Figure 12. Simulated daily-mean vertical diffusivity profiles (units: m$^2$ s$^{-1}$) on (a)
September 1, 2007 and (b) October 1, 2007.

Figure 13. Monthly-mean temperature differences from the climatology along the dateline in August: (a) Exp. G1, (b) Exp. G2, and (c) Exp. G3. Contour interval is 1°C.

Figure 14. Monthly-mean temperature differences from the climatology along 30°N in August: (a) Exp. G1, (b) Exp. G2, and (c) Exp. G3. Contour interval is 1°C.

Figure 15. Monthly-mean temperature as a function of depth and time at 30°N, 180°E: (a) Exp. G1, (b) Exp. G2, (c) Exp. G3, and (d) the climatology. Contour interval is 2°C.

Figure 16. Same as Fig. 15, except for monthly-mean temperature anomalies from the WOA01 climatology. Contour interval is 0.5°C.
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Figure 2 Comparisons of simulated (a) temperature profile $T$, (b) turbulence length scale $l$, (c) turbulence velocity magnitude $q$, (d) dissipation rate $\varepsilon$, (e) stability factor $S_h$, and (f) vertical diffusivity $K_h$ for Exps. W1-W3 on Day 5.
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Table 1 Model experiments conducted. Every group of experiments included three cases: the first case is a control experiment, in which effects of surface waves are not included; the second case includes wave-breaking effects as in Eq. (6a) and (6b); and the third case includes the effect of wave-turbulence interaction as in Eq. (8).

<table>
<thead>
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<th>Effect</th>
<th>Wind-deepening</th>
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<th>Cooling</th>
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<td>Hb1</td>
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