Dynamical roles of mixed layer in regulating the meridional mass/heat fluxes

Rui Xin Huang,[1] Chuan Jiang Huang,[2] and Wei Wang[3]

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[1] The mixed layer is an important component of the oceanic circulation system. Recent progress in energetics of the oceanic circulation suggests that the amount of external mechanical energy available for mixing is directly linked to the strength of the meridional overturning circulation. Using an analytical two-dimensional model and a three-dimensional numerical model, it is shown that the meridional distribution of mixed layer depth plays an important role in regulating the meridional overturning circulation and poleward heat flux. In fact, if the mixed layer at low and middle latitudes is deeper because of increase in mechanical energy input to the turbulence in the upper ocean, the meridional overturning circulation and poleward heat flux are enhanced in a steady circulation system, and at the same time, it may take less mechanical energy to support the subsurface diapycnal mixing.

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1. Introduction

[2] Theory of ocean general circulation has gradually evolved over the past decades. In particular, a new energetic theory of ocean general circulation has been postulated. According to the energetic theory, the maintenance of both the wind-driven and thermohaline circulation requires external sources of mechanical energy, including the winds and tides [*Munk and Wunsch*, 1998; *Huang*, 1999]. Although the ocean receives a huge amount of thermal energy, it cannot convert this thermal energy into mechanical energy very efficiently because the ocean is heated and cooled from roughly the same geopotential level, i.e., the upper surface. The inability of thermal forcing alone to drive meridional overturning circulation (MOC) observed in the world oceans can be traced back to the classical postulation of *Sandström* [1908, 1916]. A closed steady circulation can be maintained in the ocean only if the heating source is situated at a lower level than the cooling source. Sandström's postulation has also been discussed in many recent studies, e.g., *Huang* [1999], *Paparella and Young* [2002], and *Wang and Huang* [2005].

[3] Sandström's postulation was not based on accurate analysis in terms of the modern fluid dynamics language due to the lack of diapycnal mixing sustained by external mechanical energy input [*Huang*, 1999], so it is not a very accurate statement. Recent analyses reveal that the oceanic general circulation is energetically driven by mechanical energy input from the winds and tides [e.g., *Wunsch and Ferrari*, 2004; *Huang*, 2004]. As a result, for a quasi-steady ocean circulation, the strength and distribution of mechanical energy sources/sinks dedicate the strength of circulation and regulate its variability.

[4] The primary sources of mechanical energy for the ocean include tidal dissipation and wind energy input [*Munk and Wunsch*, 1998]. It is well known that tidal dissipation does not vary much for timescales shorter than centennial. The recent estimate of tidal dissipation in the open ocean is 0.7–0.9 TW [*Munk and Wunsch*, 1998]. However, tidal dissipation in the open ocean takes place primarily over rough topography, where barotropic tidal energy is converted into energy for internal tides and internal waves that sustain bottom-intensified diapycnal mixing [*Polzin et al.*, 1997; *Ledwell et al.*, 2000].

[5] Wind energy input plays a vitally important role in setting up the density structure and circulation in the upper kilometer of the open ocean. According to recent studies, wind energy input to the geostrophic current is estimated as 0.9 TW [*Wunsch*, 1998], wind energy input to surface waves is estimated as 60 TW [*Wang and Huang*, 2004a], those to the Ekman layer is estimated as 0.5–0.7 TW over near-inertial frequencies [*Alford*, 2003; *Watanabe and Hibiya*, 2002] and about 2.4 TW over subinertial ranges [*Wang and Huang*, 2004b]. Over the past decades, wind energy input has greatly changed on interannual and decadal timescales. For example, wind energy input to the geostrophic current, surface waves, and Ekman layer increases nearly 25% over the past 50 years. Especially over the past two decades, when more reliable data have been collected through satellites, wind energy input to the world oceans has increased approximately 15% [*Huang et al.*, 2006].

[1] Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, USA.
[2] Key Laboratory of Marine Science and Numerical Modeling, The First Institute of Oceanography, State Oceanic Administration, Qingdao, China.
[3] Physical Oceanography Lab, Ocean University of China, Qingdao, China.

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[6] As discussed above, the oceans receive a huge amount of mechanical energy through both the surface waves and ageostrophic currents in the Ekman layer. It is commonly believed that most of such energy is dissipated in the upper ocean, primarily through the turbulence in the oceanic mixed layer [*Craig and Banner*, 1994; *Noh et al.*, 2004]. However, most of recent studies related to energetics of the oceanic circulation have been primarily focused on the importance of subsurface diapycnal mixing in regulating MOC and poleward heat flux, e.g., *Munk and Wunsch* [1998] and *Huang* [1999]. Thus there is an open question: Does the huge amount of mechanical energy input to the surface waves and ageostrophic currents somehow contribute to the maintenance and regulation of MOC and poleward heat flux in the steady state of the oceanic circulation?

[7] Oceanic mixed layer is one of the most important components of the oceanic general circulation because it is the connection between the ocean interior and the atmosphere. In general, mixed layer depth (MLD) is controlled by external mechanical energy input and surface buoyant forcing [*Kraus and Turner*, 1967; *Hallberg*, 2003]. Since wind energy input has changed greatly over the past decades, MLD should change accordingly. On the basis of observed data, recent studies indicated that MLD varies greatly on interannual and interdecadal timescales [*Bissett et al.*, 1994; *Polovina et al.*, 1995; *Deser et al.*, 1996; *Li et al.*, 2005]. For example, *Polovina et al.* [1995] pointed out that MLD in winter and spring was 30–80% deeper during 1977–1988 than during 1960–1976 in the subtropical and transition zones, and 20–30% shallower in the subarctic zone in the North Pacific because of changes in the strength and distribution of wind stress induced by variations in the strength of the Aleutian Low. Moreover, tropical cyclones can deepen MLD at low latitudes greatly [*Price*, 1981]. Power dissipation of tropical cyclones has more than doubled in the North Atlantic and increased by about 75% in the western North Pacific in the past 30 years [*Emanuel*, 2005], so that MLD in these parts of ocean should change greatly during this period.

[8] The main goal of this study is to explore the role of MLD distribution in regulating MOC and poleward heat flux. In particular, we will explore the connection between MLD, energy required to sustain the thermohaline circulation, MOC, and poleward heat flux associated with the resulted circulation.

[9] This study is organized as follows. First, in section 2, we discuss the importance of MLD distribution in setting up the meridional pressure gradient controlling MOC. Using a simple two-dimensional analytical model ocean, we argue that deep penetration of the mixed layer due to stronger turbulence activity can lead to stronger MOC and poleward heat flux in a steady circulation system; at the same time, it may require less mechanical energy to sustain subsurface diapycnal mixing.

[10] To further verify this theoretical point, we carried out numerical experiments based on an isopycnal-coordinate ocean general circulation model in section 3. Different turbulence kinetic energy parameters of wind energy input were applied to the model, and the numerical results obtained will be compared with that inferred from the simple analytical model discussed in section 2. Finally, we concluded in section 4.

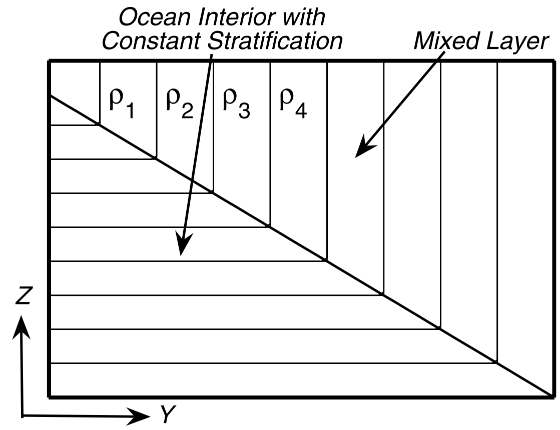

Figure 1. Sketch of the model ocean, with a mixed layer with vertically homogenized density and an ocean interior with constant stratification.

2. Analysis Based on a Two-Dimensional Model

[11] We begin with a starting point that the meridional pressure gradient is directly related to the meridional distribution of MLD. For simplicity, we assume that the density in the mixed layer is vertically homogenized. In addition, in a steady state near the northern boundary, the mixed layer penetrates to the bottom of the model ocean. Thus a deep mixed layer at lower latitudes gives rise to a large meridional baroclinic pressure difference in the upper ocean. As a result, both the meridional overturning cell and poleward heat flux are intensified. At the same time, external mechanical energy required for sustaining subsurface diapycnal mixing may be less.

[12] Our first step is to demonstrate this idea analytically, using a highly-simplified two-dimensional model with depth H and width L. Stratification below the mixed layer is assumed to be constant, and the density in the mixed layer is vertically homogenized (Figure 1). For a steady state solution obtained under the assumption of a linear equation of state (so that there is no cabbeling effect) and without the seasonal cycle, MLD at the northern boundary of the model basin should be equal to the depth H of the ocean. For simplicity, we further assume that MLD is a linear function of the meridional coordinate

$$h(y) = H(ay + 1 - a) \tag{1}$$

where $y = Y/L$, Y is the meridional distance from the southern boundary and $a \leq 1$ is a constant in nondimensional units. This linear profile of MLD is a high idealization of the complicated MLD distribution in the world oceans; however, it includes the most essential ingredient of MLD: Annual mean MLD is shallow at low latitudes and deep at high latitudes. Of course, some of the meridional variation of MLD in the world oceans is not included in such a simple profile. For example, in summer, MLD at middle and high latitudes is shallower than that near the equator [*Monterey and Levitus*, 1997; *Kara et al.*,

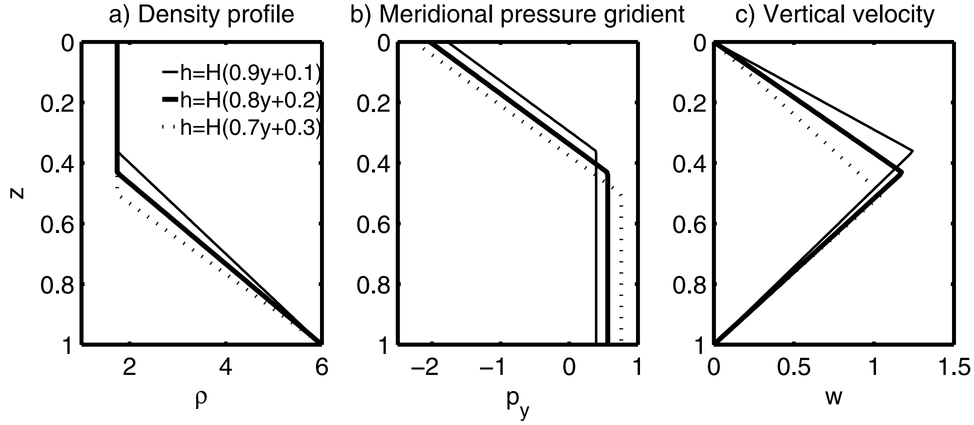

Figure 2. Structure of the solution at a station y = 0.3 for three different mixed layer depth profiles. (a) Density profile; (b) meridional pressure gradient; and (c) vertical velocity (all in nondimensional units).

2003]. Nevertheless, we hope a close examination of such a simple linear profile may help us to understand the interplay between mixed layer depth distribution, mechanical energy required for overcome friction and dissipation, and the thermohaline overturning circulation.

2.1. Stratification

[13] We assume that the density in the mixed layer is vertically homogenized and it is also a linear function of latitudinal position

$$\rho_{\rm m} = \rho_0 + \Delta\rho y, \quad -h \le z \le 0,\ 0 \le y \le 1 \tag{2}$$

[14] Since we have assumed that the stratification below the mixed layer is constant, the density in the ocean interior can be written as

$$\rho_{\rm i} = \rho_0 + \Delta\rho - \frac{\Delta\rho}{aH}(H+z), \quad -H \le z < -h \tag{3}$$

where subscripts m and i indicate the mixed layer and interior, ρ_0 is the mixed layer density at the southern boundary of the model basin, and $\Delta\rho$ is the mixed layer density difference between the southern and northern boundaries. Typical density profiles under three different MLD distributions are shown in Figure 2a. If MLD increases at a certain position (Figure 3a), the stratification below the mixed layer (i.e., $\Delta\rho/aH$ in the above formula) is enhanced actually in a steady state.

2.2. Pressure Gradient

[15] Pressure can be calculated by integrating the hydrostatic equation over depth. For a model based on the rigid-lid approximation, there is an unknown "atmospheric pressure" $p_{\rm a}(y)$, at the "rigid lid" z = 0. Integrating from the rigid lid z = 0 downward, pressure in the mixed layer is

$$p_{\rm m} = p_{\rm a} - g\rho_{\rm m} z = p_{\rm a} - g(\rho_0 + \Delta\rho y)z \tag{4}$$

Pressure in the ocean interior below the mixed layer is

$$p_{\rm i} = p_{\rm a} + g\rho_{\rm m} h + g\int_z^{-h} \rho_{\rm i} {\rm d}z \tag{5}$$

From these two relations, the meridional pressure gradients in these two regions are

$$p_{{\rm m},y} = p_{{\rm a},y} - g\Delta\rho z \tag{6a}$$

$$p_{{\rm i},y} = p_{{\rm a},y} + g\Delta\rho h \tag{6b}$$

where the second subscript y indicates the partial derivative with respect to y. The unknown atmospheric pressure gradient $p_{{\rm a},y}$ can be eliminated as follows:

[16] First, we assume that the meridional velocity is proportional to the meridional pressure gradient, i.e.,

$$v = -cp_y \tag{7}$$

where c is a constant.

[17] Second, under the Boussinesq approximation, the continuity equation requires that the vertically integrated meridional volume flux crossing each meridional position y must be zero; thus

$$\int_{-H}^{0} \nabla p {\rm d}z = 0 \tag{8}$$

Substituting equations (6a) and (6b) into equation (8) leads to the relation determining the unknown atmospheric pressure gradient

$$p_{{\rm a},y} = -g\Delta\rho h\left(1 - \frac{h}{2H}\right) \tag{9}$$

Thus the corresponding pressure gradients in the mixed layer and below are

$$p_{{\rm m},y} = -g\Delta\rho\left[h\left(1 - \frac{h}{2H}\right) + z\right] \tag{10a}$$

$$p_{{\rm i},y} = \frac{g\Delta\rho}{2H}h^2 \tag{10b}$$

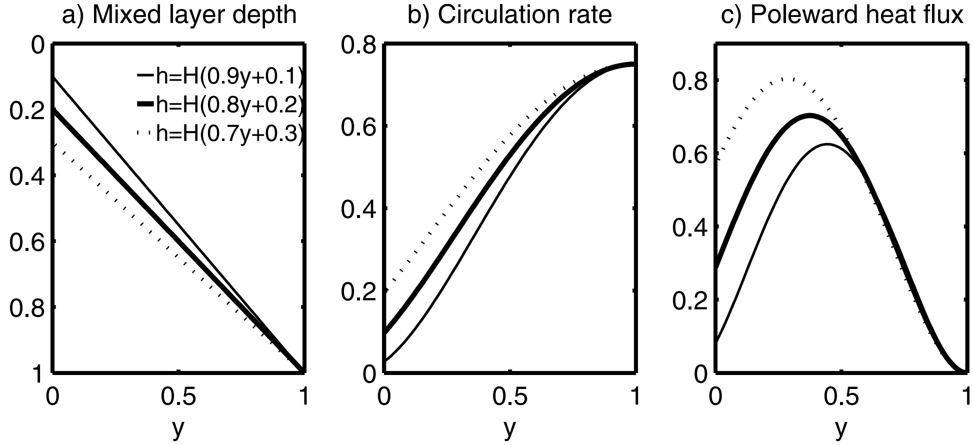

Figure 3. Meridional structure of the solution for three different mixed layer depth profile (in non-dimensional units). (a) Mixed layer depth; (b) meridional overturning rate; and (c) poleward heat flux.

[18] Typical meridional pressure gradient profiles are shown in Figure 2b. From this figure, it is readily seen that if MLD increases at a given station, the meridional pressure gradients in the mixed layer and in the ocean interior are enhanced. Therefore MOC should strengthen because the meridional velocity is proportional to the meridional pressure gradient in the model.

2.3. Velocity

[19] Substituting equations (10a) and (10b) into equation (7), the corresponding meridional velocity profiles in the mixed layer and in the interior below are

$$\nu_{\mathrm{m}} = cg\Delta\rho\left[h\left(1-\frac{h}{2H}\right)+z\right] \tag{11a}$$

$$\nu_{\mathrm{i}} = -\frac{cg\Delta\rho}{2H}h^2 \tag{11b}$$

The continuity equation is

$$\nu_y + w_z = 0 \tag{12}$$

Since vertical velocity at the surface or the bottom vanishes, the vertical velocity can be calculated from the vertical integration at each meridional location

$$w = -\int_z^0 \nu_y \mathrm{d}z \tag{13}$$

Typical vertical velocity profiles are shown in Figure 2c.

2.4. Meridional Overturning Rate

[20] In order to find the boundary between the northward and southward flows at each station, we use equation (11a) and set $\nu_{\mathrm{m}} = 0$, and this leads to the vertical position of the zero-meridional velocity layer

$$h_0 = -h(1-\frac{h}{2H}) \tag{14}$$

By definition, the meridional velocity changes its sign at h_0. Thus the meridional overturning rate is

$$\Psi = -\int_0^{h_0} \nu_{\mathrm{m}}\mathrm{d}z = \frac{cg\Delta\rho}{2}h^2(1-\frac{h}{2H})^2 \tag{15}$$

2.5. Poleward Heat Flux

[21] The poleward heat flux can be calculated as follows. Assuming a linear equation of state $\rho = \bar{\rho}(1-\gamma T)$, where γ is the thermal expansion coefficient, the poleward heat flux is

$$H_{\mathrm{f}} = c_{\mathrm{p}}\int_{-H}^0 \bar{\rho}\nu T\mathrm{d}z = -\frac{c_{\mathrm{p}}}{\gamma}\left(\int_{-h}^0 \nu_{\mathrm{m}}\rho_{\mathrm{m}}\mathrm{d}z + \int_{-H}^{-h}\nu_{\mathrm{i}}\rho_{\mathrm{i}}\mathrm{d}z\right)$$
$$= \frac{c_{\mathrm{p}}}{\gamma}\frac{cg\Delta\rho^2}{4aH^2}h^2(H-h)^2 \tag{16}$$

2.6. Energy for Sustaining Diapycnal Mixing in the Interior

[22] First, we assume that the density balance in the oceanic interior can be approximately treated in terms of a one-dimensional balance between the vertical advection and vertical diffusion

$$w\frac{\partial\rho}{\partial z} = \kappa\frac{\partial^2\rho}{\partial z^2} \tag{17}$$

Using this equation, the scale for the mixing coefficient is $\kappa = WD$, where W and D are the scales of vertical velocity and thickness of the stratified water. Therefore the mechanical energy required for sustaining mixing in the ocean interior is estimated as

$$e_i = \kappa(\rho_0 + \Delta\rho - \rho_{\mathrm{m}}) = W(H-h)(\Delta\rho - \Delta\rho y)$$
$$= \frac{cg\Delta\rho^2}{H}h(H-h)^3 \tag{18}$$

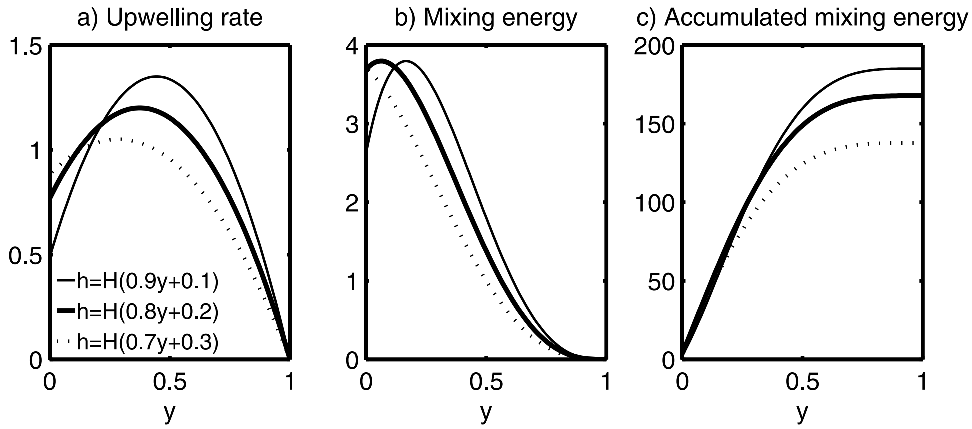

Figure 4. Meridional structure of the solution for three different mixed layer depth profiles (in non-dimensional units). (a) The upwelling rate; (b) energy required for supporting diapycnal mixing; and (c) meridionally accumulated energy required for sustaining mixing.

2.7. Sensitivity of the Circulation to the Mixed Layer Properties

[23] To explore the sensitivity of MOC and poleward heat flux to changes in mixed layer properties in the model ocean, we studied three cases with different linear MLD profiles (Figure 3a). As stated in the previous section, MLD in our simple model should be equal to the depth H of the ocean at the northern boundary, so that the difference between these three cases appears at lower latitudes. It is readily seen that for these cases, if MLD is larger at low and middle latitudes, the circulation rate is also larger there, but it remains unchanged at the northern boundary (Figure 3b). At the same time, poleward heat flux is enhanced, and the position of maximum poleward heat flux drifts to lower latitudes with the deepening of MLD at lower latitudes (Figure 3c).

[24] On the other hand, the upwelling rate at the base of the mixed layer at middle and high latitudes for the case with the deepest mixed layer $h = H(0.7y + 0.3)$ is the lowest, although it is the highest near southern boundary (Figure 4a). Similarly, energy required for sustaining the subsurface diapycnal mixing is the lowest at middle and high latitudes, and the corresponding meridionally accumulated energy for sustaining subsurface diapycnal mixing is the smallest among all three cases (Figures 4b and 4c). Thus this series of cases demonstrate that with a deep mixed layer at lower latitudes, the meridional overturning cell can transport more water at low and middle latitudes, and more heat poleward; at the same time, the circulation may require less mechanical energy for sustaining diapycnal mixing in the ocean interior.

[25] In contrast, for the case with a shallow mixed layer at lower latitudes, $h = H(0.9y + 0.1)$, MOC transports less water and heat; at the same time, the circulation may require more energy for sustaining the subsurface diapycnal mixing, as shown in Figures 3 and 4.

3. Results Obtained From an Oceanic General Circulation Model

3.1. Models

[26] The analytical model discussed above is highly simplified. Many important dynamic processes were omitted in such a two-dimensional model, such as the Ekman transport in the surface ocean, the quasi-horizontal wind-driven gyre, and buoyant forcing induced by surface freshwater flux. In addition, other important dynamics in the oceans, such as the rotation of the Earth and the β effect, were omitted as well. Thus we carried out a series of numerical experiments in order to further explore the dynamical roles of the mixed layer structure in regulating the thermohaline circulation, in particular, their connection with the meridional circulation through the dynamical constraint of mechanical energy required to sustain mixing in the subsurface oceans.

[27] The Hallberg isopycnal model (HIM [*Hallberg and Rhines*, 1996]) has been used in our numerical experiments. HIM is an oceanic general circulation model based on the isopycnal coordinate. The major advantage of using an isopycnal coordinate model is that such a model can avoid artificial diapycnal mixing associated with the vertical-horizontal advection in the z-coordinate model [*Griffies et al.*, 2000]. In addition, MLD in HIM is a diagnostic variable calculated from a bulk mixed layer model in the upper ocean, in which MLD is regulated through a combination of wind energy input and surface buoyancy forcing [*Hallberg*, 2003].

[28] The model ocean is a $60° \times 60°$ basin in spherical coordinates, with a southern boundary along the equator. The horizontal resolution is $1° \times 1°$, and the depth is 4000 m uniformly. The model has 30 layers in the vertical direction, including three modified bulk mixed layers of *Kraus and Turner* [1967] in which velocity shear is permitted, and a layer underneath these layers plays the role as a buffer layer between the mixed layers and the isopycnal layers below [*Hallberg*, 2003]. The horizontal viscosity coefficient is set to 2.5×10^4 m^2 s^{-1}, and the vertical viscosity coefficient is 10^{-3} m^2 s^{-1}. Both the thickness diffusion coefficient and horizontal diffusion coefficient are set to 1000 m^2 s^{-1} [*Gent and McWilliams*, 1990], and the diapycnal mixing coefficient is 2.0×10^{-5} m^2 s^{-1}. Wind stress profile is the same as used by *Weaver and Sarachik* [1990] (Figure 5a). Sea surface temperature is relaxed toward the a priori specified reference temperature with a relaxation time of 30 days (for a mixed layer of 25 m; Figure 5b). The time steps are

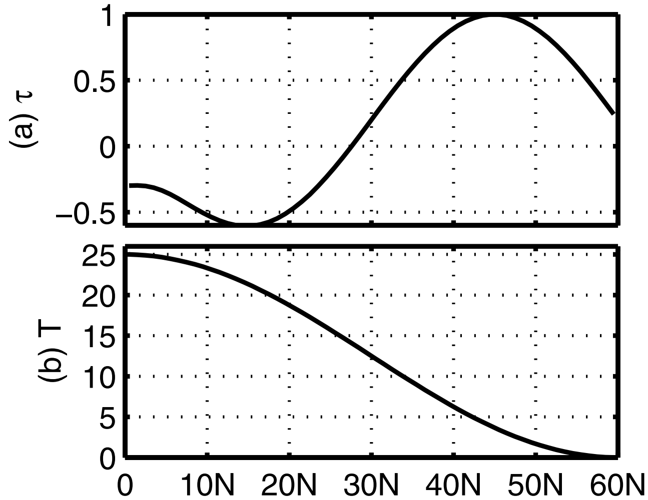

Figure 5. (a) Wind stress profile (in units of 0.1 Pa) and (b) reference temperature profile (in °C) used in the numerical experiments.

60 s, 1800 s, and 8 hours for the barotropic, baroclinic, and thermodynamic stepping, respectively.

3.2. Experiments

[29] MLD in the ocean is primarily controlled by wind stress stirring and surface buoyancy forcing. As discussed by *Wang and Huang* [2004a], the total wind energy input to surface waves in the world oceans is estimated as 60 TW, and it is postulated that a substantial portion of this energy is lost in the upper ocean through wave breaking, turbulence, and other processes. These processes are very complicated, and our understanding of their dynamical structure remains rudimentary in spite of much effort devoted to the in situ observations and theory. In general, wave breaking has an important effect on MLD, and a larger amount of wind energy input associated with wave breaking can result in deeper MLD [*Mellor and Blumberg*, 2004]. Moreover, other processes, such as the Langmuir circulation, also have important effects on MLD [*Li and Garrett*, 1997].

[30] In most existing models, wind energy input to the mixed layer is parameterized in terms of energy input to the turbulence in simple formulae. For example, in HIM, wind energy input is parameterized as

$$e = \rho m^* u^{*3} \tag{19}$$

where m^* is an empirical constant in nondimensional units and u^* is (the air side) the frictional velocity. Dynamical impact of wind stress to the surface ocean is complicated. Obviously, the contribution of wind energy input to the mixed layer cannot be accurately simulated through such a simple formula. It is speculated that a direct coupling between the traditional ocean general circulation model (OGCM) and a surface wave model (SWM) may provide a much better dynamical picture of the oceanic circulation [e.g., *Noh et al.*, 2002; *Qiao et al.*, 2004].

[31] A fully SWM-OGCM coupled model is beyond the scope of this study; instead of such a complicated model, we explore the possible impact of different MLD distribution in terms of changes in the empirical constant m^*. No doubt such an approach has some limitations. However, this approach will enable us to explore the fundamental issue in a simple way: How does the mixed layer deepening at low latitudes due to a larger amount of wind energy input affect MOC and poleward heat flux?

[32] In previous studies, the empirical constant m^* has been chosen over a quite large range. In OGCMs, m^* is usually set to 1.25 [*Oberhuber*, 1993; *Hallberg*, 2003]. However, quite different values of m^* have also been used in previous studies. *Gaspar* [1988] took a range of $0.4 \le m^* \le 0.8$ to model the seasonal cycle of the upper ocean at Ocean Station Papa. *Noh et al.* [2004] estimated turbulent kinetic energy at the surface with $m^* = 1.40$, but m^* is taken as 3.50 in the study of *Craig and Banner* [1994]. (The air side frictional velocity is used in our study, so there is a coefficient of about 28.6 relative to the studies using the water side frictional velocity.) However, *Stacey* [1999] analyzed the data from Knight Inlet in southwestern Canada and concluded that using $m^* = 5.25$ can provide the best fit to the data.

[33] In our experiments, the mode was spun up from a state of rest for 1000 years, in which m^* is set to 1.25. On the basis of this run, five experiments were run for 400 years afterward, in which m^* are set to 0.4, 1.25, 3.75, 7.5, and 12.5, respectively. The experiment with $m^* = 1.25$ is taken as the standard case in this study. The higher value of m^* implies that a larger amount of wind energy input is used to deepen MLD under the same wind stress.

[34] The reason of using a seemingly very high value of m^* in our numerical experiment is the following. At present time, most of ocean models are driven by wind stress data with coarse spatial and temporal resolutions; thus contribution due to strong nonlinear events, such as tropical cyclones and midlatitude storms, is eliminated through smoothing of the wind stress data. As a result, simple turbulence parameterization scheme, such as equation (19), can greatly underestimate contribution from strong wind events. Since wind stress data with high spatial and temporal resolutions are not currently available, a much larger value of m^* may be used in OGCMs to test the potential effects of strong wind events in the oceans.

[35] The basic idea behind this approach is the following. The contributions of wind stress to the oceanic circulation are rather complicated, and at least they can be separated into two categories: the large-scale mean wind stress, which drives the Ekman transport and the wind-driven gyre in the upper ocean, and the source for the turbulent kinetic energy in the upper ocean. Simply changing the "mean" wind stress profile in the ocean model cannot represent the complicated contribution from wind accurately. As a compromise in our conceptual study here, we choose to alternate this single parameter m^* in HIM to simulate contributions to surface turbulence in the upper ocean from the wind stress components with relatively high spatial and temporal resolution.

3.3. Experiment Results

[36] Increase in m^* has a profound effect on mixed layer deepening at low and middle latitudes, where MLD is primarily controlled by the turbulence kinetic energy input from the winds (Figure 6). It is generally accepted that increasing m^* can increase MLD in the bulk mixed layer

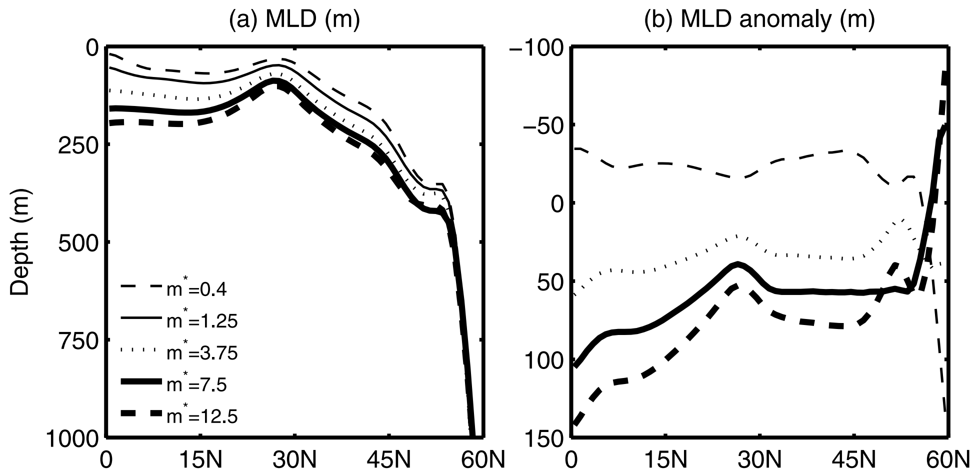

Figure 6. (a) Zonal-mean mixed layer depth in the three-dimensional ocean model and (b) its departure from the standard case (in m).

model [*Gaspar*, 1988]. Since a bulk mixed layer model is used in this study, choosing a higher value of m^* can lead to deeper MLD.

[37] However, it is controversial in the turbulence closure mixed layer model because it is believed that the effect of wave breaking is mainly limited to the near-surface zonal [*Craig and Banner*, 1994; *Noh et al.*, 2004]. In general, ocean circulation models based on the turbulence closure scheme without wave breaking underestimate MLD [*Martin*, 1985], while the simulation of MLD is improved to a certain extent when wave breaking is included in the model [*Noh et al.*, 2002]. Recently, *Mellor and Blumberg* [2004] pointed out that a larger value of m^* can result in deeper MLD in the turbulence closure model as long as MLD is controlled by wind stirring rather than cooling-induced convection. On the other hand, *Burchard* [2001] reported that wave breaking had almost no effect on MLD. Thus further study is needed to assess the effect of m^* (or wave breaking) on MLD in the turbulence closure model.

[38] Our results contradict with results from *Noh* [2004] at low latitudes. In his study, wave breaking leads to a substantial increase of MLD in the extratropical ocean, but its effect is insignificant in the equatorial region. This may be due to the following two reasons. First, a bulk mixed layer is used in our study, which is sensitive to the mechanical energy input. A larger amount of wind energy input entrains more water from the subsurface to the mixed layer, deepening the mixed layer. However, the mixed layer model used by *Noh* [2004] is based on a turbulence closure scheme; in the model, the effect of wave breaking is incorporated through modifying the surface boundary conditions. Second, the basinwise dynamical processes induced by the change of MLD at middle latitudes also contribute to MLD deepening at low latitudes. In fact, when tropical cyclones move over the oceans at low latitudes, they can deepen MLD greatly through a huge amount of mechanical energy input to the oceans [*Price*, 1981; *Emanuel*, 2001, 2005].

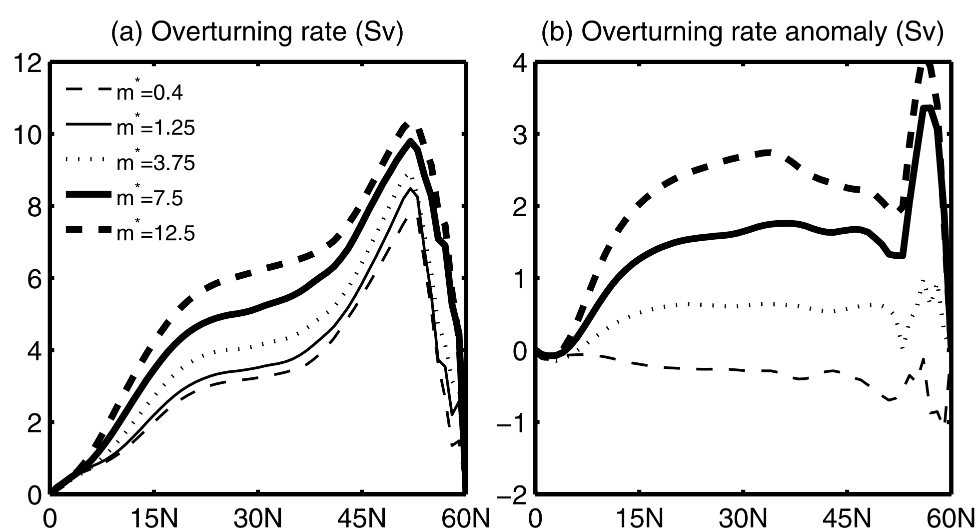

Figure 7. (a) Overturning rate obtained from a three-dimensional ocean model and (b) its departure from the standard case (in Sv).

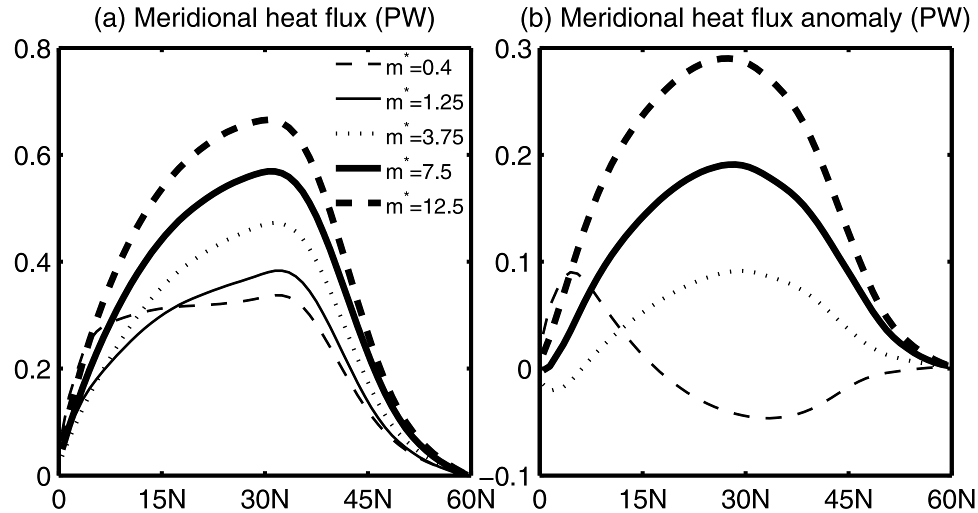

Figure 8. (a) Poleward heat flux in the three-dimensional ocean model and (b) its departure from the standard case (in units of 10^{15} W).

[39] It should be noted that the zonal-mean mixed layer depth at the high-latitude edge of the model is reduced with m^* increasing; such a decline of zonal-mean mixed layer depth is not directly related to the intensification of wind energy input into the model ocean; instead, it is due to complicated basinwise dynamical processes, which will be explained shortly.

[40] An immediate consequence of a deeper mixed layer at low and middle latitudes is the increase of the meridional pressure gradient in the ocean, and this leads to an intensification of MOC. For example, when $m^* = 7.5$, MOC increases nearly 1.6 Sv, compared to the case with $m^* = 1.25$ for the latitudinal bands of 20–45°N (Figure 7). Because of the increase in the meridional mass transport, the circulation can carry more heat poleward. In fact, the maximal poleward heat flux increases from 0.38 PW for the case of $m^* = 1.25$ to 0.57 PW for the case of $m^* = 7.5$ (Figure 8).

[41] At high latitudes, MLD is primarily governed by cooling-induced convection, and the influence of turbulence induced by wind is of lesser importance. The increase of MLD at low and middle latitudes induces more poleward heat flux, resulting in weaker cooling-induced convection at high latitudes. Thus the mixed layer becomes shallow at high latitudes with m^* increasing. On the other hand, more poleward heat flux reduces sea surface temperature at low latitudes, and this is a positive feedback loop which promotes a further deepening of MLD at lower latitudes.

[42] While the sensitivity of MOC and poleward heat flux on diapycnal diffusivity in the ocean interior has been studied extensively in previous studies [e.g., *Bryan*, 1987; *Park and Bryan*, 2000], the role of the mixed layer and the external source of mechanical energy for sustaining the thermohaline circulation have not received enough attention. In our study, a specific attention is focused on the effect of the mixed layer on MOC and poleward heat flux. In departure from the strategy of varying the diapycnal

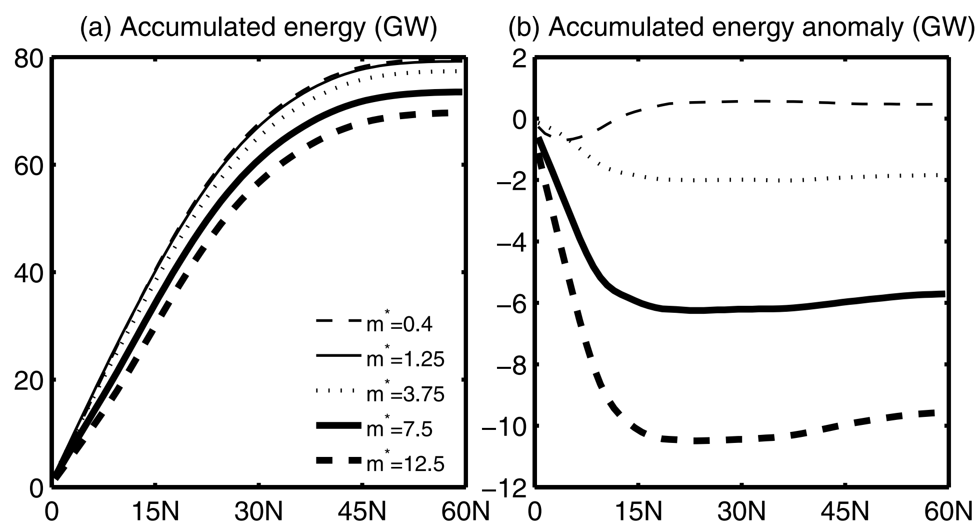

Figure 9. (a) Meridionally accumulated energy required for sustaining subsurface diapycnal mixing in the three-dimensional ocean model and (b) its departure from the standard case (in units of 10^9 W).

diffusivity used in the previous studies, we choose to fix the diapycnal diffusivity in the ocean interior and focus on the role of mixed layer and the energetics of the thermohaline circulation.

[43] From the view of mechanical energy balance, diapycnal mixing in the ocean interior is controlled by mechanical energy input from the winds and tides [*Munk and Wunsch*, 1998; *Huang*, 1999]. A larger amount of mechanical energy input can result in stronger diapycnal mixing in the ocean interior, which may enhance MOC and poleward heat flux. On the other hand, more mechanical energy input can increase MLD. This process also plays an important role in regulating MOC and poleward heat flux. For example, tropical cyclones can input a large amount of mechanical energy into the ocean which not only enhances diapycnal mixing in the ocean interior but also deepens the mixed layer greatly [*Price*, 1981], so tropical cyclones play an important role in driving MOC and poleward heat flux [*Emanuel*, 2001]. However, there is no simple linear relationship between the amount of mechanical energy input and the strength of MOC and the related poleward heat flux. The complicated nature of such connections remains a critically important issue for further study.

[44] Another interesting result from this series of experiments is that the energy required for sustaining subsurface diapycnal mixing is actually reduced for the cases with a deeper mixed layer at lower latitudes (Figure 9). This can be explained as follows. A deeper mixed layer implies a smaller density difference between the bottom water and the base of the mixed layer. Since the amount of energy required for sustaining subsurface diapycnal mixing is proportional to this density difference, it is reduced with this density difference decreasing. Thus this series of OGCM experiments confirms the theoretical analysis based on the simple two-dimensional model that deeper mixed layer induced by larger turbulence kinetic energy can give rise to stronger circulation and larger poleward heat flux, and, at the same time, less mechanical energy input is required to sustain subsurface diapycnal mixing. However, our two-dimensional model is an idealization of the ocean, and it has excluded many important dynamical factors, such as the wind stress and the complication involved with the three-dimensional circulation in the ocean; nevertheless, the basic idea seems robust.

4. Conclusion

[45] There are many factors that can contribute to the regulation of MOC and poleward heat flux. Although most recent studies on the energetics of the steady state circulation in the ocean have been focused on the important role of subsurface diapycnal mixing in regulating MOC, our study here indicates that mixed layer properties, such as MLD at low and middle latitudes, which is mostly regulated by turbulent kinetic energy input from wind stress, can also play an important role in regulating the meridional transports of mass and heat.

[46] As discussed above, large MLD at low and middle latitudes can give rise to strong MOC and poleward heat flux in a steady circulation system. Since MLD at low and middle latitudes is controlled primarily by wind energy input to the ocean, a huge amount of mechanical energy input to the surface waves and ageostrophic currents can contribute to the maintenance and regulation of MOC and poleward heat flux in the steady state of the ocean circulation. Most importantly, the amount of external mechanical energy required for sustaining diapycnal mixing in the subsurface interior ocean is reduced with MLD increasing (i.e., with a larger amount of wind energy input). The substantial difference between diapycnal mixing in the subsurface layers and in the mixed layer seems an important issue to be explored in further studies.

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References

Alford, M. H. (2003), Improved global maps and 54-year history of wind-work done on the ocean inertial motions, *Geophys. Res. Lett.*, *30*(8), 1424, doi:10.1029/2002GL016614.
Bissett, W. P., M. B. Meyers, J. J. Walsh, and F. E. Müller-Karger (1994), The effects of temporal variability of mixed layer depth on primary productivity around Bermuda, *J. Geophys. Res.*, *99*(C4), 7539–7553.
Bryan, F. (1987), Parameter sensitivity of primitive equation ocean general circulation models, *J. Phys. Oceanogr.*, *17*, 970–985.
Burchard, H. (2001), Simulating the wave-enhanced layer under breaking surface waves with two-equation turbulence models, *J. Phys. Oceanogr.*, *31*, 3133–3145.
Craig, P. D., and M. L. Banner (1994), Modeling wave-enhanced turbulence in the ocean surface layer, *J. Phys. Oceanogr.*, *24*, 2546–2559.
Deser, C., M. A. Alexander, and M. S. Timlin (1996), Upper-ocean thermal variations in the North Pacific during 1970–1991, *J. Clim.*, *9*, 1840–1855.
Emanuel, K. (2001), Contribution of tropical cyclones to meridional heat transport by the oceans, *J. Geophys. Res.*, *106*(D14), 14,771–14,781.
Emanuel, K. (2005), Increasing destructiveness of tropical cyclones over the past 30 years, *Nature*, *436*, 686–688.
Gaspar, P. (1988), Modeling the seasonal cycle of upper ocean, *J. Phys. Oceanogr.*, *18*, 161–180.
Gent, P. R., and J. C. McWilliams (1990), Isopycnal mixing in ocean circulation models, *J. Phys. Oceanogr.*, *20*, 150–155.
Griffies, S. M., R. C. Pacanowski, and R. W. Hallberg (2000), Spurious diapycnal mixing associated with advection in a z-coordinate ocean model, *Mon. Weather Rev.*, *128*, 538–564.
Hallberg, R. (2003), The ability of large-scale ocean models to accept parameterizations of boundary mixing, and a description of a refined bulk mixed-layer model, *Proceedings of the 2003 Aha Hulikoa Hawaiian Winter Workshop*, pp. 187–203, U. Hawaii.
Hallberg, R., and P. Rhines (1996), Buoyancy-driven circulation in an ocean basin with isopycnals intersecting the sloping boundary, *J. Phys. Oceanogr.*, *26*, 913–940.
Huang, R. X. (1999), Mixing and energetics of the thermohaline circulation, *J. Phys. Oceanogr.*, *29*, 727–746.
Huang, R. X. (2004), Ocean, energy flow, in *Encyclopedia of Energy*, edited by C. J. Cleveland, Vol. 4, pp, 497–509, Elsevier, New York.
Huang, R. X., W. Wang, and L. L. Liu (2006), Decadal variability of wind-energy input to the world ocean, *Deep Sea Res., Part II*, *53*, 31–41.
Kara, A. B., P. A. Rochford, and H. E. Hurlburt (2003), Mixed layer depth variability over the global ocean, *J. Geophys. Res.*, *108*(C3), 3079, doi:10.1029/2000JC000736.
Kraus, E. B., and J. S. Turner (1967), A one-dimensional model of the seasonal thermocline: II. The general theory and its consequences, *Tellus*, *19*, 98–106.
Ledwell, J. R., E. T. Montgomery, K. L. Polzin, L. C. St. Laurent, R. W. Schmitt, and J. M. Toole (2000), Evidence for enhanced mixing over rough topography in the abyssal ocean, *Nature*, *403*, 179–182.
Li, M., and C. Garrett (1997), Mixed layer deepening due to Langmuir circulation, *J. Phys. Oceanogr.*, *27*, 121–132.
Li, M., P. G. Myers, and H. Freeland (2005), An examination of historical mixed layer depth along Line P in the Gulf of Alaska, *Geophys. Res. Lett.*, *32*, L05613, doi:10.1029/2004GL021911.
Martin, P. J. (1985), Simulation of the mixed layer at OWS November and Papa with several models, *J. Geophys. Res.*, *90*(C1), 903–916.

Mellor, G., and A. Blumberg (2004), Wave breaking and ocean surface layer thermal response, *J. Phys. Oceanogr.*, *34*, 693–698.
Monterey, G. L., and S. Levitus (1997), *Climatological Cycle of Mixed Layer Depth in the World Ocean*, U. S. Gov. Printing Office, NOAA NESDIS, 5 pp.
Munk, W., and C. Wunsch (1998), Abyssal recipes II. Energetics of tidal and wind mixing, *Deep Sea Res.*, *45*, 1977–2010.
Noh, Y. (2004), Sensitivity to wave breaking and the Prandtl number in the ocean mixed layer model and its dependence on latitude, *Geophys. Res. Lett.*, *31*, L23305, doi:10.1029/2004GL021289.
Noh, Y., H. S. Min, and S. Raasch (2004), Large eddy simulation of the ocean mixed layer: The effects of wave breaking and Langmuir circulation, *J. Phys. Oceanogr.*, *34*, 720–735.
Noh, Y., C. J. Jang, T. Yamagata, P. C. Chu, and C. H. Kin (2002), Simulation of more realistic upper-ocean processes from an OGCM with a new ocean mixed layer model, *J. Phys. Oceanogr.*, *32*, 1284–1307.
Oberhuber, J. M. (1993), Simulation of the Atlantic circulation with a coupled sea ice-mixed layer-isopycnal general circulation model. Part I: Model description, *J. Phys. Oceanogr.*, *23*, 808–829.
Paparella, F., and W. R. Young (2002), Horizontal convection is non turbulent, *J. Fluid Mech.*, *466*, 205–214.
Park, Y. G., and K. Bryan (2000), Comparison of thermally driven circulation from a depth-coordinate model and an isopycnal-layer model. Part I: Scaling-law sensitivity to vertical diffusion, *J. Phys. Oceanogr.*, *30*, 590–605.
Polovina, J. J., G. T. Mitchum, and G. T. Evans (1995), Decadal and basin-scale variation in mixed layer depth and the impact on biological production in the Central and North Pacific, 1960–88, *Deep Sea Res.*, *42*, 1701–1716.
Polzin, K. L., J. M. Toole, J. R. Ledwell, and R. W. Schmitt (1997), Spatial variability of turbulent mixing in the abyssal ocean, *Science*, *276*, 93–96.
Price, J. F. (1981), Upper ocean response to a hurricane, *J. Phys. Oceanogr.*, *11*, 153–175.
Qiao, F. L., Y. L. Yuan, Y. Z. Yang, Q. A. Zheng, C. S. Xia, and J. Ma (2004), Wave-induced mixing in the upper ocean: Distribution and application to a global ocean circulation model, *Geophys. Res. Lett.*, *31*, L11303, doi:10.1029/2004GL019824.
Sandström, J. W. (1908), Dynamicsche Versuche mit Meerwasser, *Ann. Hydrogr. Marit. Meteorol.*, *36*, 6–23.
Sandström, J. W. (1916), Meteorologische Studien im schwedischen Hochgebirge, *Goteborgs K. Vetenskaps-och Vitterhetssamhalles Handl.*, Ser. 4, 22(2), 48 pp.
Stacey, M. W. (1999), Simulations of the wind-forced near-surface circulation in Knight Inlet: A parameterization of the roughness length, *J. Phys. Oceanogr.*, *29*, 1363–1367.
Wang, W., and R. X. Huang (2004a), wind energy input to the surface waves, *J. Phys. Oceanogr.*, *34*, 1276–1280.
Wang, W., and R. X. Huang (2004b), wind energy input to the Ekman layer, *J. Phys. Oceanogr.*, *34*, 1267–1275.
Wang, W., and R. X. Huang (2005), An experimental study on thermal circulation driven by non-uniform horizontal heating, *J. Fluid Mech.*, *540*, 49–73.
Watanabe, M., and T. Hibiya (2002), Global estimate of the wind-induced energy flux to the inertial motion in the surface mixed layer, *Geophys. Res. Lett.*, *29*(8), 1239, doi:10.1029/2001GL014422.
Weaver, A. J., and E. S. Sarachik (1990), On the importance of vertical resolution in certain ocean general circulation models, *J. Phys. Oceanogr.*, *20*, 600–609.
Wunsch, C. (1998), The work done by the wind on the oceanic general circulation, *J. Phys. Oceanogr.*, *28*, 2332–2340.
Wunsch, C., and R. Ferrari (2004), Vertical mixing, energy and the general circulation of the oceans, *Annu. Rev. Fluid Mech.*, *36*, 281–314.

C. J. Huang, Key Laboratory of Marine Science and Numerical Modeling, The First Institute of Oceanography, State Oceanic Administration, Qingdao, Shandong 266061, China. (cjhuang@fio.org.cn)

R. X. Huang, Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA.

W. Wang, Physical Oceanography Lab, Ocean University of China, Qingdao, Shandong 266003, China.