Parameterization of Langmuir Circulation in the Ocean Mixed Layer Model Using LES and Its Application to the OGCM

YIGN NOH, HYEJIN OK, AND EUNJEONG LEE

Department of Atmospheric Sciences, Yonsei University, Seoul, South Korea

TAKAHIRO TOYODA
Meteorological Research Institute, Japan Meteorological Agency, Tsukuba, Ibaraki, Japan

NAOKI HIROSE
Research Institute of Applied Mechanics, Kyushu University, Fukuoka, Fukuoka, Japan

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ABSTRACT

The effect of Langmuir circulation (LC) on vertical mixing is parameterized in the ocean mixed layer model (OMLM), based on the analysis of large-eddy simulation (LES) results. Parameterization of LC effects is carried out in terms of the modifications of the mixing length scale as well as the inclusion of the contribution from the Stokes force in momentum and TKE equations. The performance of the new OMLM is examined by comparing with LES results, together with sensitivity tests for empirical constants used in the parameterization. The new OMLM is then applied to the ocean general circulation model (OGCM) Meteorological Research Institute Community Ocean Model (MRI.COM), and its effect is investigated. The new OMLM helps to correct too shallow mixed layer depths (MLDs) in the high-latitude ocean, which has been a common error in most OGCMs, without making the thermocline in the tropical ocean more diffused. The parameterization of LC effects is found to affect mainly the high-latitude ocean, in which the MLD is shallow in summer and stratification is weak in winter.

1. Introduction

The ocean mixed layer model (OMLM), or the parameterization of vertical mixing in the upper ocean, is a key element in the ocean general circulation model (OGCM) or in the climate model, as it determines the downward transport of heat in the ocean and thus controls air–sea interaction through the sea surface temperature (SST).

A large number of OMLMs have been suggested with an aim to reproduce the upper ocean more realistically in the OGCM (Pacanowski and Philander 1981; Rosati and Miyakoda 1988; Blanke and Delecluse 1993; Chen et al. 1994; Sterl and Kattenberg 1994; Large et al. 1997; Burchard and Bolding 2001; Noh et al. 2002, 2005; Qiao et al. 2004; Richards et al. 2009; Sasaki et al. 2012). In spite of significant progress through these efforts, there still remain many features in the OGCM that require the further improvement of the OMLM. For example, too shallow mixed layer depths (MLDs) in the high-latitude ocean are a common error in most OGCMs, especially in the Southern Ocean (Li et al. 2001; Kara et al. 2003; Gnanadesikan et al. 2006; Noh and Lee 2008; Belcher et al. 2012; Schiller and Ridgway 2013). It also causes the warm bias of SST in summer in the high-latitude ocean. Another common error is a too diffused thermocline in the tropical ocean (Halpern et al. 1995; Li et al. 2001, Large and Danabasoglu 2006; Tsujino et al. 2011), which is likely to cause weaker El Niño in the climate model (Meehl et al. 2001).

Langmuir circulation (LC), which appears in the form of an array of alternating horizontal roll vortices with axes aligned roughly with the wind, represents one of the most important characteristics of the ocean mixed layer (see, e.g., Leibovich 1983; Smith 2001; Thorpe 2004).
The prevailing theory of LC is made by Craik and Leibovich (1976), who described the formation of LC in terms of instability brought on by the interaction of the Stokes drift with the wind-driven surface shear current. The instability is initiated by an additional “vortex force” term in the momentum equation as \( \mathbf{u} \times \omega \), where \( \mathbf{u} \) is the Stokes drift velocity and \( \omega \) is vorticity.

It has been well established, that within the mixed layer, LC enhances vertical mixing greatly, resulting in the uniform profiles of temperature and velocity (McWilliams et al. 1997; Noh et al. 2004; Li et al. 2005; Skyllingstad 2005). On the other hand, the role of LC in the mixed layer deepening remains less clear. LC appears to contribute to the enhanced mixed layer deepening in some cases (Li et al. 1995; Sullivan et al. 2007; Grant and Belcher 2009; Kukulka et al. 2009) but not in other cases (Weller and Price 1988; Thorpe et al. 2003). Skyllingstad et al. (2000) suggested from the large-eddy simulation (LES) results that the effects of LC are consistent with bulk models (e.g., Niiler and Kraus 1977), by taking into account the characteristics of the free surface with wave breaking, based on the analysis by Craig and Banner (1994). The model has been applied to various ocean models (Hasumi and Emori 2004; Duan et al. 2008; Rascle and Ardhuin 2009; Tsujino et al. 2011), and parameterizations used in the model have been confirmed by LES results (Noh et al. 2004, 2009, 2010, 2011a; Noh and Nakada 2010). For the detailed description of the model, one can refer to Noh and Kim (1999).

In the present work, we propose a new OMLM, which includes the parameterization of LC effects based on the analysis of LES results by Noh et al. (2011a). We examine the performance of the new OMLM by comparing with LES results, together with sensitivity tests of empirical constants used in the parameterization. At the next step we apply the new OMLM to the OGCM and investigate how the inclusion of LC effects in the OMLM helps to reproduce the upper ocean more realistically from the OGCM.

### 2. The parameterization of LC effects in the OML model

The new OMLM is developed by the modification of the Noh model, which has been shown to reproduce the realistic upper-ocean structure well (e.g., Noh and Kim 1999; Noh et al. 2002, 2005, 2011b). The model is a turbulence closure model using eddy diffusivity and viscosity, similar to the Mellor–Yamada model (Mellor and Yamada 1982), but reproduces a uniform mixed layer, consistent with bulk models (e.g., Niiler and Kraus 1977), by taking into account the characteristics of the free surface with wave breaking, based on the analysis by Craig and Banner (1994). The model has been applied to various ocean models (Hasumi and Emori 2004; Duan et al. 2008; Rascle and Ardhuin 2009; Tsujino et al. 2011), and parameterizations used in the model have been confirmed by LES results (Noh et al. 2004, 2009, 2010, 2011a; Noh and Nakada 2010). For the detailed description of the model, one can refer to Noh and Kim (1999).

In the model, the eddy viscosity and diffusivity are calculated by

\[
K_m = S_m q l, \quad \text{and} \quad K_h = S_h q l, \quad (1)
\]

where \( S_m \) and \( S_h \) are empirical constants.

To calculate \( q \), the turbulent kinetic energy (TKE) equation is solved, in which the dissipation rate \( \varepsilon \) is calculated by

\[
\varepsilon = C q^3 / \Lambda, \quad (3)
\]
where $\Lambda$ is the dissipation length scale and $C$ is an empirical constant.

The TKE flux at the sea surface $F$ is given by $F = muw^3$, where $m = 40$ is obtained from LES by adjusting the TKE flux so as to produce the dissipation rate profile matching observation data (Noh et al. 2004). Although some analyses of observation data suggested larger values for $m$ (Craig and Banner 1994; Gerbi et al. 2009), Gemmrich and Farmer (1999) suggested a comparable value from the analysis including the effect of LC.

It is assumed that $l$ and $\Lambda$ are the same, as in Mellor and Yamada (1982), and the mixing length scale in the absence of stratification $l_0$ is prescribed by

$$\frac{1}{l_0} = \frac{1}{\kappa(z + z_0)} + \frac{1}{h},$$

(4)

where $z_0$ is the roughness length scale at the sea surface ($z = 0$ m) and $h$ is the MLD. The term $h$ is calculated in the model by $E(z = h) = 10^{-4}E(z = 0)$, where $E$ is TKE ($=q^2/2$), representing the depth to which vertical mixing actually occurs. The roughness length scale at the surface is given by $z_0 = 1$ m, following Craig and Banner (1994). Although $m$ and $z_0$ approach zero on a day when wave breaking does not occur any more under very strong solar radiation and weak wind (Burchard 2001; Noh et al. 2011b), their variation is not considered here, as its effect is negligible in monthly mean climatology.

Since TKE production is dominated by the TKE flux in the mixed layer, rather than shear production (Craig and Banner 1994; Noh et al. 2004, 2009, 2011a; Sullivan et al. 2007; Grant and Belcher 2009), the effect of stratification is parameterized in terms of $Rt$, instead of $Ri$; that is,

$$ll_0 = (1 + \alpha Rt)^{-1/2},$$

(5)

where $\alpha$ is a proportional constant. It is equivalent to say that $l$ approaches to $l \sim \alpha^{-1/2}l_0$ as $Rt$ becomes large, as in Blanke and Delecluse (1993) and D’Alessio et al. (1998), where $l_0$ is the buoyancy length scale, defined by $l_0 = q/N$. To represent convective mixing under unstable stratification ($Rt < 0$), $K_m$ and $K_h$ are assumed to be

$$K_m = \frac{K_h = K_{MAX}}{K_{MAX} = 1.0 m^2 s^{-1}},$$

following the suggestion by Klinger et al. (1996).

The recent analysis of LES results by Noh et al. (2011a) found that LC enhances vertical mixing, or increases $K_m$ and $K_h$, only if MLD is shallow and stratification is weak. It is also found that the increase of $K_m$ and $K_h$ is attributed predominantly to the increase of $l$ than $q$, although both $q$ and $l$ are increased by LC (Fig. 6 in Noh et al. 2011a). In other words, if $K_m^L$ and $K_h^O$ represent $K_m$ with and without LC, its ratio $K_m^L/K_m^O$, estimated by $(q^L q^O)(l^L/l^O)$ from (2), is largely determined by $l^L/l^O$. Here, the subscripts $L$ and $O$ of $q$ and $l$ represent the cases with and without LC, respectively.

The effect of LC on $l$ can be represented by the relationship between $ll_0$ and $Rt$, as shown in Fig. 1 (also Fig. 8 in Noh et al. 2011a). Here, data are obtained from LES experiments of the wind mixed layer deepening with no surface heat flux and $u_a = 0.02 m s^{-1}$, under the conditions of different $La$ [$=(ua/U_b)^{1/2}$] and initial stratification {initial stratifications as $N_0^2 = 10^{-5}, 5 \times 10^{-5},$ and $2 \times 10^{-4} s^{-2}$, referred to as N1, N2, and N3, and intensities of LC as $La = \infty, 0.64,$ and 0.32 with $\lambda = 40 m$, referred to as L0, L1, and L2}. Details of LES experiments can be found in Noh et al. (2011a).

LES provides the information on the fluxes of momentum and buoyancy as well as the mean velocity and buoyancy. It enables us to evaluate $K_m$ and $K_h$ directly from the relations such as $-\overline{uw}\partial U/\partial z - \overline{bw}\partial V/\partial z = K_m[(\partial U/\partial z)^2 + (\partial V/\partial z)^2]$ and $-\overline{bw} = K_h \partial B/\partial z$, where $(\overline{uw}, \overline{bw})$ is the momentum flux, $\overline{bw}$ is the buoyancy flux, $(U, V)$ is the mean horizontal velocity, and $B$ is the mean buoyancy. Once $K_m$ and $K_h$ are evaluated, it is possible to calculate $l$ directly from relations (1) and (2), since $q$ is also obtained directly from LES.
Remarkably, data $l/l_0$ versus $Rt$ collapse well at larger $Rt$ ($Rt > \sim 1$), thus confirming relation (5) regardless of the presence of LC; that is, $l \sim \alpha^{-1/2} l_b$ with $\alpha \approx 50$, as shown by a solid line. On the other hand, when $Rt \ll 1$, $l$ follows (4) for $l_b$ in the absence of LC, but it becomes much larger than it does in the presence of LC. The difference of $l/l_0$ between the cases with and without LC becomes larger for smaller $Rt$ and for stronger LC (smaller La) at the same $Rt$. For example, $l/l_0$ becomes as large as 10, as $Rt$ approaches 0 for $La = 0.32$. Figure 2 confirms that $l$ is much larger in the presence of LC within the mixed layer, where $Rt$ is very small (Sullivan et al. 2007; Noh et al. 2011a). It also shows that $l$ tends to increase with $\lambda$ at small $La$.

Note that small values of $Rt = (N_0 q)^2$ represent the conditions of shallow depth $l_0$, weak stratification $N$, and strong turbulence $q$. It can be understood physically by the fact that large eddies of LC, which are driven by the surface convergence, are essentially unforced and behave inertially, unlike convective eddies, which are continuously driven by the buoyancy force during the downward movement. Therefore, large eddies by LC are efficient for the vertical mixing only under weak stratification. Their contribution is also limited vertically by the fact that LC penetrates only to a certain depth.

If $q \sim \nu_L, l_0 \sim h, \Delta B \sim N^2 \delta$, and $\delta \approx h$ are assumed near the MLD ($z = h$), where $\delta$ is the thickness of the thermocline, $(N_0 q)^2 \propto h \Delta B / \nu_L^2$ at $z = h$. In this case, Fig. 1 becomes equivalent to the model suggested by Li et al. (1995), Li and Garrett (1997), and Smith (1998) where the additional deepening of the mixed layer occurs in the presence of LC when $h \Delta B / \nu_L^2$ becomes smaller than a critical value. For example, Smith (1998) suggested that further deepening occurs when $h \Delta B / \nu_L^2 < 9.8$ in addition to the mixed layer deepening by the PWP model (Price et al. 1986), which occurs under the condition $h \Delta B / (\Delta U)^2 < 0.64$, where $\Delta U$ is the velocity jump across the MLD.

The mixing length scale in the presence of LC, shown in Fig. 1, can be parameterized by modifying (4) to

$$\frac{1}{l_0} = \frac{1}{\kappa \Gamma (z + z_0)} + \frac{1}{h},$$

(6)

where $\Gamma$ is 1 in the absence of LC ($La = \infty$), and increases with decreasing $La$; for example, $\Gamma$ can become as large as 10 at $La = 0.32$. We can find from (5) and (6) that $l^2/l_0^2 \sim \Gamma$ at $Rt \ll 1$ and $z \ll h$, but $l^2 \sim l^2 / \alpha^{-1/2} l_b$.

Fig. 2. Profiles of $l$ with different $\lambda$, obtained from $K_h (N_0^2 = 10^{-5} s^2, t = 12 h)$ (black dotted lines are the profiles from LES with $La = \infty$) (violet: $\lambda = 20 m$, red: $\lambda = 40 m$, and green: $\lambda = 80 m$): (a) $La = 0.64$ and (b) $La = 0.32.$
at Rt ≫ 1, as shown in Fig. 1. For example, solid and dashed lines in Fig. 1 represent $l^0$ and $l^t$, obtained from (5) and (6) with $\Gamma = 1$ and 10, respectively. Note that both $l_0$ and Rt in (5) are now defined in terms of the new length scale (6); for example, $Rt = (NL_0^2/q)^2$ in the presence of LC.

The parameterization of mixing by LC in (6) can be interpreted as strong mixing ($l^t/l^0 \gg 1$) when $0 < Rt \ll 1$. It contrasts with the parameterization of convective mixing in terms of $K_{\text{MAX}}$, which can be interpreted as strong mixing ($l^C/l^0 \gg 1$) when $Rt < 0$, if the mixing length scale during convection is defined as $l^C = K_{\text{MAX}}/\delta h q$. It is also interesting to observe that the present parameterization has a similar effect on vertical mixing to that of surface wave–induced fluctuation, as suggested by Qiao et al. (2004). With an aim to parameterize its effect, they introduced an additional eddy viscosity and diffusivity proportional to $aU_s \exp(-4\pi z/\Lambda)$, where $a$ is the wave amplitude. In both cases, the additional eddy diffusivity is generated at shallow depths, whose magnitude increases with $U_s$. The additional mixing is suppressed by stratification in the present parameterization, however, unlike Qiao et al. (2004).

In the present model, the modification of $l_0$ is made only for the mixing length scale but not for the dissipation length scale $\Lambda$; that is,

$$\frac{1}{\Lambda_0} = \frac{1}{\kappa(z + z_0)} + \frac{1}{\hbar}, \quad \text{(7)}$$

regardless of the presence of LC, where $\Lambda_0$ is the dissipation length scale in the absence of stratification. It is based on the fact that the estimation of $\varepsilon$ can be obtained.
from the conventional length scale such as (4), only with the modification of \( q \) for small \( \text{La} \) (Grant and Belcher 2009). One can expect that the mixing process, dominated by large eddies, is enhanced radically by large eddies of LC, but the dissipation process, dominated by small eddies, is affected only by the increase of TKE due to LC that enhances the energy cascade. Mellor and Yamada (1982) also questioned the validity of \( l = \Lambda \), although they assumed it. Therefore, the values of \( \alpha \) in (5), representing the sensitivity to stratification, are not necessarily the same for \( l \) and \( \Lambda \). Thierry and Lacarrère (1983) and Gaspar et al. (1990) suggested that \( l \) must be more sensitive to stratification than \( \Lambda \) because vertical mixing is controlled by the TKE of vertical velocity, while dissipation is controlled by total TKE. Accordingly the suppression of \( \Lambda \) by stratification is parameterized as

\[
\Lambda/\Lambda_0 = (1 + \alpha_d \text{Rt}_d)^{-1/2},
\]  

where \( \text{Rt}_d = (N\Lambda_\alpha/q)^2 \), and the optimized value for \( \alpha_d \) is obtained as \( \alpha_d = 20 \) from the comparison with LES results. It will be discussed in section 4a. Note that \( \Lambda_0 \) is used on the unstable condition, and it also implies different length scales for mixing (\( l^* \)) and dissipation.

In addition, momentum and TKE equations should include the terms arising from the Stokes force in the presence of LC as (e.g., D’Alessio et al. 1998)

\[
\frac{\partial U}{\partial t} = \frac{\partial}{\partial z} \left( K_m \frac{\partial U}{\partial z} \right) + f(V + u_z), \tag{9}
\]

\[
\frac{\partial V}{\partial t} = \frac{\partial}{\partial z} \left( K_m \frac{\partial V}{\partial z} \right) - f(U + u_z), \tag{10}
\]
\[
\frac{\partial E}{\partial t} = \frac{\partial}{\partial z} \left( K_E \frac{\partial E}{\partial z} \right) + K_m \left( \left( \frac{\partial U}{\partial z} \right)^2 + \left( \frac{\partial V}{\partial z} \right)^2 \right) + K_m \left( \frac{\partial U}{\partial z} \frac{du}{dz} + \frac{\partial V}{\partial z} \frac{dv}{dz} \right) + K_h \frac{\partial B}{\partial z} - \varepsilon, \tag{11}
\]

where \( u_s = (u_s, v_s) \).

In the present parameterization, \( K_m \) and \( K_h \) increase under the influence of LC in terms of the increase of \( q \) and \( l \) in (1) and (2). The increase of \( l \) is prescribed as \( l^2/l^O \sim \Gamma \) at \( R_l \ll 1 \), but \( l^2/l^O \sim 1 \) at \( R_l \gg 1 \). The increase of \( q \) is realized through the inclusion of the contribution from the Stokes force in momentum and TKE equations [(9)–(11)]. It will be shown in section 4a that, under the present parameterization, the increase of \( K_m \) and \( K_h \) is mainly due to the increase of \( l \), rather than \( q \), in agreement with LES results (Noh et al. 2011a). Discussion on the value of \( \Gamma \) used in the OGCM will be made in section 3.

Various other approaches have been attempted to parameterize the effect of LC on vertical mixing so far, although consensus has yet to be achieved. As a new approach, we propose the parameterization of LC, which is simple with little additional computational cost but realizes the key effect of LC effectively, with an aim to apply it to the OGCM.

3. OGCM

The OGCM used in this study is the Meteorological Research Institute (MRI) Community Ocean Model, version 3 (MRLCOM3; Tsujino et al. 2011). This is a free-surface, depth-coordinate ocean–sea ice model that solves primitive equations using the Boussinesq and hydrostatic approximations. Fundamental equations are formulated on generalized orthogonal coordinates. The Arakawa B grid arrangement is employed, and the coastlines are created by connecting tracer points instead of velocity points. A split-explicit algorithm is used for the barotropic and baroclinic parts of the equations (Killworth et al. 1991). Several upper levels follow the undulation of the sea surface, as in \( \sigma \)-coordinate models (Hasumi and Emori 2004). A highly accurate advection scheme is adopted for momentum (Ishizaki and Motoi 1999) and tracers (Prather 1986).

The global model is a tripole grid model developed for the operational seasonal prediction in the Japan Meteorological Agency (Toyoda et al. 2013). Its horizontal grid spacing is 1° in the zonal and 0.3°–0.5° in the meridional directions. It has 52 vertical levels with an additional bottom boundary layer (Nakano and Sugino-hara 2002). The thickness of the surface layer is 2 m, and the upper layers above 500 m are resolved into 34 layers.
Mixing along neutral surfaces caused by eddy stirring was parameterized by a skew flux \((300 \text{ m}^2 \text{s}^{-2}/100 \text{ km} \times \text{grid size})\) (Gent and McWilliams 1990) and isoneutral mixing \((1000 \text{ m}^2 \text{s}^{-2})\) (Redi 1982). Smagorinsky (1963) viscosity is applied using an anisotropic tensor (Smith and McWilliams 2003). Vertical mixing is parameterized by the Noh model (Noh and Kim 1999; Noh et al. 2002). The low vertical resolution tends to underestimate \(N\) and thus \(R_t = (N_l/q)^2\), near the thermocline, and the low-frequency wind forcing tends to underestimate the TKE flux at the surface \((\mu^3 \ast)\). Therefore, the parameter values in the OMLM need to be modified, and the optimized values currently used in MRI.COM \((a = 52, \mu = 120, m = 100)\) (Tsujino et al. 2011) are used in the present work. The background vertical mixing for tracers was a horizontally uniform vertical profile with values of \(0.1 \times 10^{-4} \text{m}^2\text{s}^{-1}\) at the surface \((=nu^2_b)\). Therefore, the parameter values in the OMLM need to be modified, and the optimized values currently used in MRI.COM \((\alpha = \alpha_s = 120, m = 100)\) (Tsujino et al. 2011) are used in the present work. The background vertical mixing for tracers was a horizontally uniform vertical profile with values of \(0.1 \times 10^{-4} \text{m}^2\text{s}^{-1}\) at the surface and \(2.7 \times 10^{-4} \text{m}^2\text{s}^{-1}\) near the bottom.

The model is driven by the wind stress, downward shortwave and longwave radiation, precipitation, 2-m air temperature and specific humidity, and 10-m wind speed fields calculated from the 6-hourly data from the Japanese 55-yr reanalysis data (Kobayashi et al. 2015) during the 1958–2012 period with the initial condition taken from the end of the spinup period by using the climatological forcing field. Albedo at the sea surface is set as a function of latitude. Bulk fluxes are computed using the bulk formula described by Large and Yeager (2004), which is widely used in the OGCM (e.g., Danabasoglu et al. 2012). Here, upward longwave radiation is calculated using the surface temperature, and the emissivity factor is assumed to be unity. Integration continues until 2012, and the climatology of the last 20 yr (1993–2012) is used for analysis.

To realize the effect of LC, the information on La and \(\lambda\) is necessary, which determines \(\mathbf{u}_s\) and \(\Gamma\) in (6) and (9)–(11). For this purpose, it is desirable ultimately to couple the ocean wave model to the OGCM (e.g., Qiao et al. 2004; Moon 2005; Liu et al. 2011; Janssen 2012). In the present work, however, as a first step, we use constant values for \(\Gamma\) and \(\lambda\). The values of La has been found to be located predominantly in the range \(0.2 < La < 0.5\) (Smith 1992; Tamura et al. 2012; Belcher et al. 2012). We used a constant value for \(\Gamma\) as \(\Gamma = 10\) too, based on Fig. 1. As to the typical value of La, Li and Garrett (1993) estimated the e-folding depth \(D_e(=\lambda/4\pi)\) as \(D_e = 0.12U_{10}^2/g\), based on the Person–Moskovitz spectrum for fully developed seas, where \(U_{10}\) is the wind speed at \(z = 10\text{m}\), and the value used in the LES experiments \((\lambda = 40\text{m})\).
corresponds to $U_{10} = 16 \text{ m s}^{-1}$. The same value is used in the present OGCM. The justification of using constant values of $T$ and $\lambda$ will be discussed in section 4a. Note that the values of $m$ and $z_0$, which determine the surface boundary conditions for $F$ and $l$, should be evaluated by the wave model in principle too (e.g., Terray et al. 1996), although constant values are used as a useful approximation.
4. Results

a. Comparison with LES results

We examine first whether the new OMLM including the effect of LC (OMLM-L) reproduces the LES results properly before applying it to the OGCM. Figures 3 and 4 compare the evolutions of profiles of temperature $T$ and dissipation rate $\varepsilon$ from OMLM-L during the wind mixed layer deepening in both cases with and without LC ($La = 0.32, \alpha = \infty$). Here, LES results are the same as in Noh et al. (2011a) and correspond to L0 and L2 with the initial stratification $N_1$ ($N_1^2 = 10^{-5}$ s$^{-2}$). In the absence of LC ($La = \infty, \Gamma = 1$), OMLM-L is identical to the old OMLM (OMLM-O). For the case of OMLM-L, $\Gamma = 10$ is assumed at $La = 0.32$, based on Fig. 1. The initial profiles of the OMLM are set to those at $t = 1$ h from LES results in order to avoid the spinup time in LES ($\ell/hu \sim 10^3$ s). The vertical resolution of the OMLM is set to the same as LES ($\Delta z = 1.25$ m), and salinity is constant in both cases.

It is found that OMLM-L reproduces well the evolutions of $T$ and $\varepsilon$ profiles from LES results in both cases with and without LC. The OMLM reproduces successfully both the strong turbulence near the surface by wave breaking ($\sim 10^{-4}-10^{-3}$ m$^2$ s$^{-3}$) and the magnitude of $\varepsilon$ within the mixed layer ($\sim 10^{-7}-10^{-6}$ m$^2$ s$^{-3}$). Meanwhile, in the OMLM the mixing below the MLD, generated typically by internal gravity waves, is not calculated but treated by the background diffusivity. Therefore, $\varepsilon$ decreases abruptly to the background value below the MLD in the OMLM, and it makes the thermocline in the OMLM slightly sharper. Nonetheless, Figs. 3 and 4 indicate that OMLM-O can predict the mixed layer deepening well in the absence of LC, and OMLM-L can predict successfully the enhanced mixed layer deepening in the presence of LC.

Figure 5 compares the time series of MLD and $\Delta$SST from the OMLM with different values of $\alpha_d$ in the absence of LC ($La = \infty$) for both cases N1 and N3 ($N_1^2 = 10^{-5}, 2 \times 10^{-4}$ s$^{-2}$). Here, MLD is calculated by the depth of maximum stratification for both LES and OMLM results, and $\Delta$SST means the difference of SST from the initial value. Note that the MLD based on the maximum stratification is usually equivalent to the MLD based on the decrease of TKE, used in the mixed layer model, as one can notice from Figs. 3 and 4. It is found that the growth rate of MLD tends to increase with decreasing $\alpha_d$, as expected from smaller $\varepsilon$ under stratification. It is followed by the faster decrease of SST with decreasing $\alpha_d$. The best agreement with LES results is found with $\alpha_d = 20$, although the optimized value differs slightly depending on cases. The fact that $\alpha_d$ is smaller than $\alpha$ is consistent with the argument by Therry and Lacarrère (1983) and Gaspar et al. (1990) that $\ell$ is more sensitive to stratification than $\Lambda$. 

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**Fig. 8.** Zonal-mean MLD bias in (left) January and (right) July: (a) $\Delta \sigma_s = 0.1$ kg m$^{-3}$ and (b) $\Delta \sigma_s = 0.03$ kg m$^{-3}$ (red: OGCM-L – WOA, orange: OGCM-L – Argo, blue: OGCM-O – WOA, and green: OGCM-O – Argo).
When stratification is strong at the base of the mixed layer, LES tends to simulate the realistic turbulent heat flux, dominated by large-scale eddies, but simulate the unrealistically low dissipation rate, dominated by small-scale eddies, as shown by Skyllingstad et al. (1999). Therefore, we evaluate \( l \) directly from LES but evaluate \( \Lambda \) by comparing density profiles predicted by LES and the mixed layer model. Furthermore, both \( \ell_b \) and \( \delta_\alpha \) are assumed approximations, so the direct evaluations of both \( \alpha \) and \( \delta_a \) from LES do not necessarily lead to a more accurate model.

Figure 6 shows similar sensitivity tests to the value of \( \Gamma \) in OMLM-L in the presence of LC (La = 0.32) for both cases N1 and N3. Here, OMLM-L with \( \Gamma = 1 \) corresponds to the case where the additional terms arising from the Stoke force in momentum and TKE [(9)–(11)] are included, but the mixing length scale is not changed. The difference between the results from OMLM-O and OMLM-L with \( \Gamma = 1 \) is negligible. On the other hand, the substantial enhancement of mixed layer deepening is observed at OMLM-L with \( \Gamma = 5 \) and 10.

This result illustrates two important aspects of the parameterization of LC. First, the increase of \( K_b \) is mainly due to the increase of \( l \), rather than \( q \) (Noh et al. 2011a). Second, the results from OMLM-L with \( \Gamma = 5 \) and 10 indicate that the results are not so sensitive to the value of \( \Gamma \), as long as it is sufficiently larger than 1. From these results we can conclude that the effect of LC on vertical mixing is rather insensitive to the values of \( L \) and \( \lambda \), as long as \( L \) is sufficiently small so as to make \( \Gamma \) large. Note that the variation of \( \lambda \) affects only \( q \) in the OMLM.

Finally, it is necessary to mention that the contribution of LC in vertical mixing becomes less important if the mixed layer is strongly affected by the surface buoyancy flux. The mixed layer deepening is mainly controlled by convection under strong surface cooling and thus becomes independent of \( L \) (Li et al. 2005; Noh et al. 2010). On the other hand, LC itself is weakened under surface heating and is ultimately broken down if the intensity of surface heating becomes sufficiently strong (Weller and Price 1988; Min and Noh 2004; Goh and Noh 2013).

### b. OGCM results

Figure 7 shows the distributions of MLD bias from climatological data of the World Ocean Atlas (WOA) (http://www.nodc.noaa.gov/OC5/woa13/) in January and July from the OGCM using the OMLM with and without the inclusion of LC effects, which will be referred to as OGCM-L and OGCM-O, together with the difference between OGCM-L and OGCM-O. Here, MLD is calculated by the criterion of the density difference from the surface as \( \Delta \rho_o = 0.1 \text{ kg m}^{-3} \). The calculation of MLD based on the maximum buoyancy gradient, used in section 4a, is not suitable here because of low vertical resolution and various factors affecting the temperature profile other than vertical mixing. The OGCM is shown to reproduce MLD fairly well, regardless of the presence of LC effects (OGCM-O), as discussed in detail in Tsujino et al. (2011). On the other hand, systematic bias, common in most OGCMs, is still observed in the reproduction of the mixed layer, such as large deviations of MLD in the frontal zones of the western boundary currents, owing to the inaccurate simulation of circulation and the underestimation of MLD in the high-latitude ocean, especially in the Southern Ocean (Kara et al. 2003; Gnanadesikan et al. 2006; Noh and Lee 2008; Belcher et al. 2012; Danabasoglu et al. 2012; Schiller and Ridgway 2013).

The most prominent difference between OGCM-O and OGCM-L is the increase of MLD in OGCM-L in

<table>
<thead>
<tr>
<th>Compared to WOA Observations</th>
<th>( \Delta \rho_o = 0.1 \text{ kg m}^{-3} )</th>
<th>( \Delta \rho_o = 0.03 \text{ kg m}^{-3} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>OGCM-O</td>
<td>January</td>
<td>July</td>
</tr>
<tr>
<td>60°–50°S</td>
<td>36.36</td>
<td>118.19</td>
</tr>
<tr>
<td>30°–20°S</td>
<td>6.07</td>
<td>23.02</td>
</tr>
<tr>
<td>5°S–5°N</td>
<td>18.77</td>
<td>12.64</td>
</tr>
<tr>
<td>20°–30°N</td>
<td>21.91</td>
<td>5.21</td>
</tr>
<tr>
<td>50°–60°N</td>
<td>163.33</td>
<td>5.07</td>
</tr>
<tr>
<td>( \Delta \rho_o = 0.1 \text{ kg m}^{-3} )</td>
<td>( \Delta \rho_o = 0.03 \text{ kg m}^{-3} )</td>
<td></td>
</tr>
<tr>
<td>OGCM-L</td>
<td>January</td>
<td>July</td>
</tr>
<tr>
<td>60°–50°S</td>
<td>15.93</td>
<td>92.86</td>
</tr>
<tr>
<td>30°–20°S</td>
<td>6.33</td>
<td>24.44</td>
</tr>
<tr>
<td>5°S–5°N</td>
<td>17.18</td>
<td>12.49</td>
</tr>
<tr>
<td>20°–30°N</td>
<td>24.10</td>
<td>5.34</td>
</tr>
<tr>
<td>50°–60°N</td>
<td>156.77</td>
<td>5.86</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Compared to Argo Observations</th>
<th>( \Delta \rho_o = 0.1 \text{ kg m}^{-3} )</th>
<th>( \Delta \rho_o = 0.03 \text{ kg m}^{-3} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>OGCM-O</td>
<td>January</td>
<td>July</td>
</tr>
<tr>
<td>60°–50°S</td>
<td>38.42</td>
<td>127.18</td>
</tr>
<tr>
<td>30°–20°S</td>
<td>7.68</td>
<td>20.32</td>
</tr>
<tr>
<td>5°S–5°N</td>
<td>16.99</td>
<td>12.12</td>
</tr>
<tr>
<td>20°–30°N</td>
<td>16.48</td>
<td>5.59</td>
</tr>
<tr>
<td>50°–60°N</td>
<td>185.83</td>
<td>7.81</td>
</tr>
<tr>
<td>( \Delta \rho_o = 0.1 \text{ kg m}^{-3} )</td>
<td>( \Delta \rho_o = 0.03 \text{ kg m}^{-3} )</td>
<td></td>
</tr>
<tr>
<td>OGCM-L</td>
<td>January</td>
<td>July</td>
</tr>
<tr>
<td>60°–50°S</td>
<td>15.10</td>
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<td>30°–20°S</td>
<td>4.73</td>
<td>20.75</td>
</tr>
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<td>5°S–5°N</td>
<td>15.29</td>
<td>11.56</td>
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<tr>
<td>50°–60°N</td>
<td>176.24</td>
<td>3.91</td>
</tr>
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</table>
both summer and winter hemispheres in the high-latitude ocean ($\phi > 40^\circ$) (Fig. 7). Consequently, OGCM-L rectifies the underestimation of MLD in the high-latitude ocean in OGCM-O significantly. Especially too shallow MLDs in the Southern Ocean during summer, which is a common error in most OGCMs, almost disappear. The present result confirms the suggestion by Belcher et al. (2012) that the underestimation of MLD in the Southern Ocean may be caused by the lack of LC effects in the OMLM.

Fig. 9. (top) Distributions of SST from climatology (WOA), (middle top) SST bias of OGCM-O (OGCM-O − WOA), (middle bottom) SST bias of OGCM-L (OGCM-L − WOA), and (bottom) SST difference between OGCM-O and OGCM-L (OGCM-L − OGCM-O); (a) January and (b) July.
Meanwhile, significant improvement notwithstanding, the MLD is still underestimated in the high-latitude winter in OGCM-L. We can consider several reasons for this bias.

First, the determination of MLD based on $\Delta \sigma_\theta = 0.1 \text{ kg m}^{-3}$ tends to overestimate MLD from climatological data in the high-latitude winter with the small density difference across the thermocline because the density profile is diffused owing to the averaging of a large number of individual profiles (Noh and Lee 2008). Figure 8 compares the distributions of the zonal-mean MLD bias from two different density difference criteria ($\Delta \sigma_\theta = 0.03$ and $0.1 \text{ kg m}^{-3}$). It confirms that the shallow bias of MLD during winter in the high-latitude ocean is substantially reduced if the smaller value of $\Delta \sigma_\theta$ ($=0.03 \text{ kg m}^{-3}$) is used. Furthermore, Fig. 8 compares the distributions of the zonal-mean MLD bias from two different sources of observation data (WOA and Argo). Here, the MLD from the Argo data (Argo Science Team 2001) was calculated from an averaged climatological temperature and salinity profiles at each grid during the period 2001–14. It is found that the patterns of the MLD bias from the OGCM are not affected by observation data sources, although the MLD from Argo tends to be slightly deeper, probably reflecting the less diffused thermocline. The corresponding rms errors of MLD at different latitudinal bands are listed in Table 1.

Second, the prediction of MLD during winter is affected by many other processes as well, such as the parameterizations of convection, lateral mixing, and sea ice. It is possible that the inclusion of nonlocal mixing in the OMLM can enhance convective mixing, as in the case of the atmospheric boundary layer (Hong and Pan 1996). Its role is less clear, however, in the ocean, in which uniform temperature is maintained near the sea surface by wave breaking. The OGCM results from the OMLM using nonlocal mixing (KPP model; Large et al. 1997) still show the underestimation of MLD in the Southern Ocean (Danabasoglu et al. 2012). The overestimated lateral heat transport from the lower latitude can suppress convection, the inaccurate lateral mixing causes the unrealistic temperature and salinity structures of the deeper water in the high-latitude ocean, and the formation of sea ice can modify the momentum, heat, and salinity fluxes at the sea surface.

Third, it has been found that convective deepening is sensitive to the horizontal resolution and the frequency of atmospheric forcing of the OGCM because of the presence of cold storm events or mesoscale eddies can modify the convection process greatly (Holdsworth and Myers 2015; Oschlies 2002). Note that the MLD bias in OGCM-L almost disappears in the Southern Ocean summer (Fig. 7), in which the prediction of the MLD is not affected by the above-mentioned processes, unlike the winter MLD, and the density difference across the thermocline is much larger.

Since SST is mainly determined by the downward transport of surface heat flux in the high-latitude ocean during summer, the increase of MLD leads to the decrease of SST and thus helps reduce the warm bias in the high-latitude ocean during summer of both hemispheres (Fig. 9). The rms errors of SST at different latitudinal bands are listed in Table 2. The effect of increased MLD on SST is not so clear in winter, however, because the

![Fig. 10. Distributions of the zonally averaged zonal velocity of ACC (black: AVISO, red: OGCM-L, and blue: OGCM-O): (a) January and (b) July.](image-url)
heat budget of the mixed layer is affected by lateral heat transport and entrainment of the deeper layer as well. For example, OGCM-L produces a larger SST bias during winter in the North Pacific than OGCM-O, in spite of the deeper MLD. It is due to the presence of temperature inversion appearing in the North Pacific subarctic region, or the mesothermal structure (Ueno and Yasuda 2000), which is often overestimated in the OGCM as shown in Fig. 12 (below) [see also Fig. 5 in Tsujino et al. (2011)].

More realistic simulation of the MLD in the Southern Ocean also helps to predict the more realistic Antarctic Circumpolar Current (ACC) by controlling the downward flux of momentum. Figure 10 compares the distributions of the surface zonal velocity of ACC from the OGCM with the AVISO data (http://www.aviso.altimetry.fr/duacs/) during the period 1993–2012. It shows clearly that the overestimation of the surface zonal velocity from OGCM-O in January is rectified in OGCM-L.

One remarkable aspect in the difference between OGCM-O and OGCM-L is that the low-latitude ocean, especially the tropical ocean, is hardly affected by the inclusion of the LC effects, as illustrated in annual-mean temperature in the zonal cross section along the Pacific equator (Fig. 11).

The latitudinal dependence is clearly identified in the distribution of temperature at the meridional cross section (Fig. 12). In the high-latitude ocean (\(\phi > 40^\circ\)), MLD is indeed significantly deeper in OGCM-L in both summer and winter, as indicated in Fig. 7, and therefore closer to WOA. On the other hand, in the low-latitude ocean (\(\phi < 20^\circ\)), the difference between OGCM-O and OGCM-L is much smaller.

The latitudinal dependence of the LC effects can be explained by the fact that the parameterization of the LC effects in term of the modification of the length scale, shown in (6), enhances vertical mixing only when \(R_t \ll 1\) (Fig. 1). Small \(R_t\) appears under the condition of weak stratification or shallow MLD, if the variation of \(q\) is not considered. The distributions of temperature in Fig. 12 reveal that MLD is shallow in summer and stratification is weak in winter in the high-latitude ocean (\(\phi > 40^\circ\)). Therefore, either shallow MLD in summer or weak stratification in winter makes \(R_t\) small near the MLD. On the other hand, in the low-latitude ocean (\(\phi < 20^\circ\)), the mixed layer is always deeper than 50 m and is bounded below by strong stratification, thus making \(R_t\) large near the MLD throughout the year. Figure 13 indeed shows that \(K_h\) is larger and penetrates deeper in OGCM-L in the high-latitude ocean, while \(K_h\) profiles are virtually the same between two cases in the tropical ocean.

Another persistent problem associated with the OMLM in the OGCM is a too diffused thermocline,
especially in the tropical ocean (Halpern et al. 1995; Li et al. 2001; Tsujino et al. 2011; Large and Danabasoglu 2006). If vertical mixing is enhanced at all latitudes, it may help increase MLD in the high-latitude ocean, but it is likely to make the thermocline in the tropical ocean even more diffused. The present parameterization of the effect of LC provides a desirable feature of increasing the MLD in the high-latitude ocean, without making the thermocline more diffused in the tropical ocean. Small values of La (La < 0.35) are dominant in the high-latitude ocean, but larger values of La appear more frequently in the low-latitude ocean (Belcher et al. 2012). According to the present OGCM results, however, the low-latitude ocean is not affected by LC effects regardless of La. Furthermore, sensitivity tests of the OMLM (Fig. 6) suggest that the enhancement of vertical
mixing is not so sensitive to the values of La, as long as La is small enough to make $\Gamma$ sufficiently large ($\Gamma > 5$). One sensitivity test of the OGCM also shows that the global averages of the MLD difference between two OGCM-L with different $\lambda$ (13 and 40 m) in January and July are only 6.2% and 7.2% of those between OGCM-O and OGCM-L. Therefore, the assumption of $\Gamma = 10$ over the whole ocean, based on $La = 0.32$ and $\lambda = 40$ m, may still provide a useful approximation for the realization of LC effects in the global ocean.

The vertical distributions of zonally averaged temperature differences from the annual mean in the meridional plane for January and July clearly illustrate the improvement of the OGCM results with the introduction
Fig. 14. Zonally averaged temperature difference from the annual mean as a function of depth and latitude for (left) January and (right) July: (a) WOA, (b) OGCM-O, (c) OGCM-L, and (d) contours of $|\Delta T| = 3^\circ$ (solid line) and $1^\circ$C (dashed line), where $|\Delta T|$ is the magnitude of temperature difference (black: WOA, blue: OGCM-O, and red: OGCM-L).
of LC parameterization (Fig. 14). It shows that the penetration depths of temperature difference from OGCM-L become deeper at the high-latitude ocean in both summer and winter hemispheres in better agreement with WOA, while they remain invariant with OGCM-O in the low-latitude ocean. The improvement from OGCM-L is also supported in the seasonal variation of temperature anomaly profiles at 40°N (Fig. 15). It is worthwhile to mention that the penetration depths of zonally averaged temperature difference, such as shown in Figs. 14 and 15, are often substantially underestimated in OGCM results when other OMLMs are used (Li et al. 2001).

Finally, we investigated how sensitively the MLD responds to the fluctuation of surface forcing by comparing the variance $\sigma_h^2$ of the interannual variation of the monthly mean MLD from the OGCM to Argo data during the period 2001–12, where both OGCM and Argo data are available (Fig. 16). It is found that $\sigma_h$ is smaller in OGCM results than in Argo data in general. It may be due to the filtering of high-frequency atmospheric forcing in the OGCM to a certain degree.

It is found that $\sigma_h$ from OGCM-L becomes larger than that of OGCM-O during summer in the high-latitude ocean. It helps to alleviate the underestimation of $\sigma_h$ from OGCM-O in the Southern Ocean during summer. The deeper MLD from OGCM-L may make the variation of MLD larger during summer. On the other hand, the variation of MLD during winter tends to decrease slightly in OGCM-L. Since the MLD is maintained at a deeper depth, the increase of MLD by convection can be smaller in OGCM-L.

5. Conclusions

The present paper proposed a new OMLM by including the effect of LC on the vertical mixing in the upper ocean, based on the analysis of LES results. The new OMLM not only reproduces the enhanced vertical mixing by LC in good agreement with LES results, but also helps to reproduce the more realistic upper-ocean structure from the OGCM by rectifying too shallow MLDs in the high-latitude ocean, which has been a common error in most OGCMs.

Parameterization of LC effects is carried out in terms of the modifications of the mixing length scale in addition to the inclusion of the additional terms arising from the Stokes force in momentum and TKE equations. The former plays a more important role in enhancing vertical mixing. The new parameterization contributes to enhance vertical mixing significantly only in the high-latitude ocean, characterized by weaker stratification in winter and shallower MLD in summer. Consequently,

![Figure 15](attachment:image.png)
it helps to increase MLD in the high-latitude ocean, without making the thermocline in the tropical ocean more diffused. It also helps to reproduce more realistic zonal velocity of ACC and the SST increase during summer in the Southern Ocean.

Various attempts have been made to parameterize the effect of LC in the vertical mixing process so far, but its effect is not yet taken into account in most OGCMs. Here, we propose a new parameterization as one approach. The new parameterization is based on the LES...
results and requires only minor modification of the existing OMLM with little additional computational cost. Nonetheless, it is shown to improve the performance of the OGCM substantially by rectifying too shallow MLDs in the high-latitude ocean without making the thermocline in the tropical ocean more diffused. Although it is desirable to develop a more sophisticated parameterization of LC in the OMLM ultimately, the present parameterization is likely to help improve the performance of the OGCM without much difficulty.

The present paper mainly focuses on showing how the new parameterization of the effect of LC can improve the OGCM performance. Accordingly, we used the constant values of $La$ and $\lambda$, representing the global-mean values, as a first step. Nonetheless, the OMLM results are found to be insensitive to the magnitude of $La$ and $\lambda$, as long as $La$ is small enough to make $\Gamma$ larger than 5, whose condition is generally satisfied in the high-latitude ocean. Larger values of $La$ appear more frequently in the low-latitude ocean, but the effect of LC does not appear there in the OGCM. Therefore, the present parameterization of LC based on constant $La$ and $\lambda$ may still provide a useful improvement of the current OGCM, without the extensive effort to couple it with the wave model. Note that constant values $m$ and $z_0$ are also used as useful approximations, although their accurate estimations require the coupling with the wave model in principle.

Nonetheless, it is desirable ultimately to couple the OGCM with the wave model and to evaluate directly $La$ and $\lambda$, and consequently $\Gamma$, as a function of them. The wave model can also help us to calculate $m$ and $z_0$ directly. Further examination of the OGCM under various conditions, including resolution and forcing frequency, will help improve the performance of the OMLM through the optimization of model parameters. The improvement in the parameterizations of convection and lateral mixing can also improve the predictability of MLD in the high-latitude winter.

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