Langmuir Mixing Effects on Global Climate: WAVEWATCH III in CESM

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Abstract

Large-Eddy Simulations (LES) have shown the effects of ocean surface gravity waves in enhancing the ocean boundary layer mixing through Langmuir turbulence. Neglecting this Langmuir mixing process may contribute to the common shallow bias in mixed layer depth in regions of the Southern Ocean and the Northern Atlantic in most state-of-the-art climate models. In this study, a third generation wave model, WAVEWATCH III, has been incorporated as a component of the Community Earth System Model, version 1.2 (CESM1.2). In particular, the wave model is now coupled with the ocean model through a modified version of the K-Profile Parameterization (KPP) to approximate the influence of Langmuir mixing. Unlike past studies, the wind-wave misalignment and the effects of Stokes drift penetration depth are considered through empirical scalings based on the rate of mixing in LES. Wave-Ocean only experiments show substantial improvements in the shallow biases of mixed layer depth in the Southern Ocean. Ventilation is enhanced and low concentration biases of pCFC-11 are reduced in the Southern Hemisphere. A majority of the improvements persist in the presence of other climate feedbacks in the fully coupled experiments. In addition, warming of the subsurface water over the majority of

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global ocean is observed in the fully coupled experiments with waves, and the cold subsurface ocean temperature biases are reduced.

Keywords: Langmuir mixing, Mixed layer, KPP, Climate model

1. Introduction

The climate system is a multi-scale system. Processes at different spatial scales frequently interact with each other, especially in the ocean surface boundary layer. Biases in the mixed layer depth (MLD) of the ocean surface boundary layer lead to biases in dynamically important quantities such as temperature and salinity, as well as in the transport of passive tracers such as the chlorofluorocarbons (CFCs) and nutrients and other biogeochemical tracers (Wang et al., 2012; Long et al., 2013; Moore et al., 2013). MLD biases of both signs are commonly observed in the state-of-the-art climate models (Fox-Kemper et al., 2011; Danabasoglu et al., 2012; Belcher et al., 2012), particularly a persistent shallow bias in distinct regions of the Southern Ocean (Sallée et al., 2013). Accurate modeling of the climate therefore requires better representation of processes in the ocean surface boundary layer from large to small scales.

A recent parameterization of submesoscale eddies in the ocean mixed layer substantially improves the simulation of the upper ocean climate (Fox-Kemper et al., 2008, 2011). While the restratification of ocean surface layer by the submesoscale eddies reduces or eliminates many deep MLD biases, it exacerbates the shallow MLD bias in the Southern Ocean. Hence, even smaller scale phenomena, such as ocean surface gravity waves, are believed to contribute to this longstanding model bias (Belcher et al., 2012). Langmuir mixing in particular is one potential mechanism that typically deepens the MLD, even in the presence of mixed layer eddy restratification nearby (Hamlington et al., 2014). Since Langmuir mixing and submesoscale eddy restratification respond to very different forcings, the deepening and shoaling effects do not necessarily cancel out. Instead the deepening and shoaling induced by the two mechanisms will occur separately depending on the ocean state and forcings. This study there-

fore attempts to assess the impact of surface gravity waves on the MLD and global climate by parameterizing Langmuir mixing in a climate model that also uses a submesoscale eddy restratification parameterization (Fox-Kemper et al., 2008, 2011).

It is well known that ocean surface gravity waves can influence the climate through a variety of processes at the ocean-atmosphere interface (Cavaleri et al., 2012). However, few of them are explicitly represented in the state-of-the-art climate models at present. Efforts to model the bulk wave influences have focused mainly on two processes: (a) wave breaking (Sullivan et al., 2004) and (b) Langmuir circulation (Langmuir, 1938) and its turbulent form, Langmuir turbulence (McWilliams et al., 1997; Sullivan and McWilliams, 2010). While the mixing and dissipation induced by wave breaking decay rapidly away from the surface, Langmuir turbulence can elevate the turbulent kinetic energy and mixing throughout the mixed layer and is therefore believed to be a major mechanism in deepening the mixed layer (Kantha and Clayson, 2004; D’Asaro et al., 2014). Langmuir turbulence has been studied extensively with Large Eddy Simulations (LES) of the wave-averaged or Craik-Leibovich equations (e.g. McWilliams et al., 1997; McWilliams and Sullivan, 2000; Harcourt and D’Asaro, 2008; Van Roekel et al., 2012; Hamlington et al., 2014). A diagnostic study by Belcher et al. (2012) suggests that turbulent energy available for mixing in the ocean surface boundary layer is underestimated without forcing by the surface waves, which is verified by Sutherland et al. (2014a) with summer observations in the North Atlantic. Using a second moment closure model (Harcourt, 2013), D’Asaro et al. (2014) found roughly a factor of two greater turbulent kinetic energy within the mixed layer, and deepening of MLD by 15-20% on average at high latitudes when Langmuir turbulence was included.

Efforts have been made to estimate the bulk effects of Langmuir mixing by scaling the vertical turbulent kinetic energy (VKE) with parameters such as the Langmuir number¹, and inspire a practical parameterization of Langmuir

¹A bulk measure of the relative importance of shear and Stokes forcing, with smaller values

56 mixing by modifying the K-profile parameterization (KPP; Large et al., 1994).
 57 McWilliams and Sullivan (2000, hereafter MS2K) first attempted to account for
 58 Langmuir mixing in KPP by introducing an enhancement factor to the turbulent
 59 velocity scale based on the turbulent Langmuir number La_t (McWilliams et al.,
 60 1997). Smyth et al. (2002, hereafter SS02) refined the MS2K scheme by adding a
 61 modulation to the enhancement factor by stable or convective stratification con-
 62 ditions. However, both of these schemes assume simplistic relationships between
 63 wind and waves, e.g., assuming wind and waves are aligned, which are uncom-
 64 mon in the real world (Hanley et al., 2010; Webb and Fox-Kemper, 2015). In
 65 addition, the Stokes drift magnitudes integrated over the wavenumber-direction
 66 wave spectrum typically decay faster than exponentially with depth (Webb and
 67 Fox-Kemper, 2011). Since La_t only uses surface magnitudes, it may not cor-
 68 rectly estimate the relative effect of Stokes forcing to shear forcing. To account
 69 for the penetration depth of Stokes drift versus the surface layer depth, Har-
 70 court and D’Asaro (2008) proposed a VKE scaling predicted by a surface layer
 71 averaged Langmuir number, La_{SL} . To account for the effect of misaligned wind
 72 and waves, Van Roekel et al. (2012, hereafter VR12) proposed a new VKE scal-
 73 ing predicted by the surface layer averaged and projected Langmuir number,
 74 $La_{SL,proj}$. The basic idea is that when the wind and waves are misaligned,
 75 both the friction and Stokes drift velocities must be projected into the Lang-
 76 muir cell direction, which lies in between, to represent the important portion of
 77 the wind induced shear and Stokes forcing. It generally reduces the influence of
 78 Langmuir turbulence when misaligned wind and waves occur.

79 These scalings were tuned to best match the statistics of LES, but rarely
 80 have they been tested in a climate model. To our knowledge, only one study

indicating stronger wave effects. It has various definitions to account for different aspects of the
 wind wave relation. The simplest one is the turbulent Langmuir number, La_t , first introduced
 by McWilliams et al. (1997) and defined in (5). Other definitions include the surface layer
 averaged Langmuir number, La_{SL} , introduced by Harcourt and D’Asaro (2008) and defined
 in their (27), and the surface layer averaged and projected Langmuir number, $La_{SL,proj}$,
 introduced by Van Roekel et al. (2012) and defined in (11).

81 (Fan and Griffies, 2014) has implemented and tested MS2K and SS02 scalings
 82 in a climate model, the Geophysical Fluid Dynamics Laboratory (GFDL) Cli-
 83 mate Model (CM2M). SS02 is their preferred scheme, improving the MLD and
 84 sea surface temperature (SST) most. However, as will be shown later, both
 85 schemes introduce too much mixing when implemented in the National Center
 86 for Atmospheric Research (NCAR) Community Earth System Model, version
 87 1.2 (CESM1.2). This is one of our motivations to use the newer scalings, as well
 88 as a desire to quantify the novel physical processes. It should be noted that
 89 the phenomenology and behavior of Langmuir turbulence are better described
 90 in LES work (See, e.g., McWilliams et al., 1997; Harcourt and D’Asaro, 2008;
 91 Grant and Belcher, 2009; Van Roekel et al., 2012). Our primary goal is to test
 92 the newly proposed VR12 scaling with $La_{SL,proj}$ in CESM, thereby assessing
 93 the impact of Langmuir mixing on global climate. This is the first time this
 94 scaling has been tested in a state-of-the-art climate model.

95 A third generation wave model WAVEWATCH III is coupled with CESM1.2
 96 to provide the wave spectrum for calculating the Stokes drift velocity and Lang-
 97 muir number. Here, Langmuir mixing is parameterized with different VKE
 98 scalings in KPP and its impacts on the MLD, ocean ventilation, as well as
 99 ocean temperature and salinity, are assessed in this coupled climate model. It
 100 should be noted that this work represents a first step towards more compre-
 101 hensive studies on the influences of ocean surface gravity waves on the global
 102 climate. In this study the effects of Langmuir mixing are parameterized in the
 103 climate model through KPP in the most straightforward way. Wave breaking is
 104 not considered here, nor the wave influences on the atmosphere from roughening
 105 the ocean surface beyond the ensemble average effects related to wind speed.
 106 The set of parameterizations explored here for Langmuir mixing in KPP could
 107 easily be implemented in any other climate models using KPP and a coupled
 108 wave model to provide the Stokes drift velocity.

109 The details of coupling WAVEWATCH III in CESM1.2 and parameterizing
 110 Langmuir mixing in KPP are described in Section 2, followed by their impact
 111 on MLD, water mass ventilation, and ocean temperature and salinity in Section

112 3. We will summarize in Section 4 and close with a brief discussion in Section
113 5.

114 2. Methods

115 2.1. Coupling WAVEWATCH III in CESM1.2

116 A modified version of WAVEWATCH III (version 3.14; Tolman, 2009), an
117 operational third generation wave model developed by the National Oceanic
118 and Atmospheric Administration (NOAA) and the National Centers for Envi-
119 ronmental Prediction (NCEP), has been incorporated into the NCAR Commu-
120 nity Earth System Model, version 1.2 framework (CESM1.2; Gent et al., 2011;
121 Vertenstein et al., 2013) as a new active model component. All model physics
122 and settings in this component are based on the work by Rascle et al. (2008)
123 and Ardhuin et al. (2010)².

124 A schematic diagram of WAVEWATCH III coupling in CESM1.2 is displayed
125 in Fig. 1. As illustrated, WAVEWATCH III is now two-way coupled with the
126 ocean component, the Parallel Ocean Program version 2 (POP2; Smith et al.,
127 2010), through the central coupler (CPL7; Craig et al., 2011), but only one-way
128 coupled with the atmosphere and sea ice components. WAVEWATCH III gener-
129 ates and evolves the wave action density spectrum in a coupled spatial-spectral
130 domain; it is forced by sea surface temperatures (SSTs) and currents from the
131 ocean model, and by surface air temperatures, 10 m wind velocities, and ice
132 fraction concentrations, from either an active atmosphere and sea ice model or
133 from forcing datasets such as the inter-annually varying Coordinated Ocean-ice
134 Reference Experiments phase II atmospheric data sets (CORE-II; Large and
135 Yeager, 2009). WAVEWATCH III passes back to the ocean model the surface
136 Stokes drift (see next paragraph), Langmuir number (various definitions follow-
137 ing McWilliams et al., 1997; Harcourt and D’Asaro, 2008; Van Roekel et al.,

²The relevant compile switches are ‘ST3’, ‘STAB3’, and ‘FLX0’; all others are default.
Further discussion on the choice of source terms will be delayed until Section 5.

2012) and the angles between wind and Langmuir cells (as part of the enhance-
 ment factor following Van Roekel et al., 2012, see Section 2.2.3 for detail). The
 boundary layer depth diagnosed in the ocean model is also an input variable for
 WAVEWATCH III to calculate the surface layer averaged Langmuir number,
 La_{SL} , and the angle between wind and Langmuir cells.

It should be noted that two forms of Stokes drift are used for calculating
 the Langmuir number and are diagnosed within WAVEWATCH III using the
 frequency-direction spectrum and a parametric f^{-5} tail assumption (Webb and
 Fox-Kemper, 2011; Webb, 2013). It is necessary to pass the first, the surface
 Stokes drift, to the ocean model in order to estimate the wave induced entrain-
 ment from below the ocean surface boundary layer (see Section 2.2.4 for more
 detail). The second, a surface layer averaged Stokes drift, relies on the mixed
 layer depth from the ocean model and is used when the effects of misaligned
 wind and waves and Stokes drift penetration depth are considered (see (11)).
 Since this latter form has been depth-integrated analytically, there is no con-
 cern that the faster than exponential decay in Stokes drift is not resolved by the
 coarse vertical resolution of the climate ocean model grid.

The coupling frequency of WAVEWATCH III with other model components
 is the same as that for the atmosphere and sea ice components – the wave-
 atmosphere coupling interval is 30 minutes if running active atmosphere and
 sea ice models, and 6 hours if using CORE datasets. Since the ocean model
 communicates with the coupler once per day, the wave-ocean coupling is not as
 frequent as the wave-atmosphere coupling. However, the short-time misalign-
 ment of wind and waves may be captured here since it depends on the wave-
 atmosphere coupling frequency, not the wave-ocean coupling frequency. Further
 work will assess the importance of the diurnal cycle of wave-forced turbulence
 versus other high-frequency ocean coupling effects, such as near-inertial shear
 driven mixing (Jochum et al., 2013); for now the coupling frequency to POP2
 is too low in the base CESM1.2 configuration to examine such phenomena.

Running WAVEWATCH III over the global ocean with high resolution (e.g.
 1° spatial resolution) is computationally expensive. To capture the wave clima-

169 tology well while reducing the cost as much as possible, we need to choose an
 170 optimal spatial and spectral resolution for WAVEWATCH III before incorpo-
 171 rating it into CESM. Four different grids, G1F25T24, G1F13T12, G2F25T24
 172 and G3F25T24, have been tested in this study, which are summarized in Table
 173 1. We have tested each grid by running WAVEWATCH III alone with CORE-
 174 II interannual forcing data for an arbitrarily chosen month, here the month of
 175 August, 2000. Surface Stokes drift magnitudes for three coarser resolutions are
 176 compared with a higher control resolution, G1F25T24, and root mean square
 177 errors (RMSEs) and the relative errors are tabulated in Table 1. Coarsening the
 178 spectral grid resolution increases the error more so than coarsening the spatial
 179 grid resolution. In addition, the computational cost is not significantly reduced
 180 on the coarser spectral grid; in contrast, WAVEWATCH III performs much
 181 faster on the coarser spatial grid both in CPU hours and wall clock measures
 182 (Webb, 2013). G3F25T24 therefore appears to be the best choice of those con-
 183 sidered, with a relative error of 11.2% for surface Stokes drift (corresponding to
 184 a relative error in the enhancement factor of no greater than 16.8%) versus the
 185 much more expensive G1F25T24 model. It should be noted that polar ice caps
 186 were used in this grid, and as such, we are assuming that there are no wave
 187 effects north of $\sim 78^\circ\text{N}$ in the Arctic.³ While future longer simulations would
 188 need to address this shortcoming, it is not expected to significantly affect lower
 189 latitude Stokes drift velocities, which are the dynamical emphasis of this study.

190 2.2. *Parameterizing Langmuir mixing in KPP*

191 Parameterizing Langmuir mixing in the K-profile parameterization (KPP;
 192 Large et al., 1994) involves two processes to account for the effects of Langmuir
 193 turbulence: first, enhanced vertical turbulent kinetic energy (VKE) within the
 194 ocean surface boundary layer and second, deeper boundary layer penetration.

³The ice lines within WAVEWATCH III were chosen to maximize the latitudinal regions covered without degrading model performance by using smaller time steps to account for shrinking grid cell widths.

195 The former is parameterized by an enhancement factor applied to the turbulent
 196 velocity scale, for which different scalings have been proposed. The second ac-
 197 counts for eddies formed near the surface, or within a Langmuir cell being more
 198 energetic due to the additional Stokes, or Langmuir velocity, so that they are
 199 able to penetrate deeper into the stratification below. Five different experiments
 200 are designed to explore the impact of different aspects of Langmuir mixing on
 201 global climate through these effects, which are described below and summarized
 202 in Table 2 for quick reference.

203 2.2.1. *McWilliams and Sullivan (2000) parameterization of VKE*

204 Following the work of McWilliams and Sullivan (2000), we introduce an
 205 enhancement factor $\mathcal{E}$ to the turbulent velocity scale in KPP

$$W = \frac{ku^*}{\phi} \mathcal{E}, \quad (1)$$

206 with $k = 0.4$ the von Kármán constant, $u^* \equiv \sqrt{|\boldsymbol{\tau}|/\rho_0}$ the magnitude of surface
 207 friction velocity ($\boldsymbol{\tau}$ is the surface wind stress and ρ_0 is the surface ocean density)
 208 and ϕ the dimensionless flux profile.

209 The enhanced turbulent velocity scale affects the vertical viscosity, tracer
 210 diffusivity and nonlocal flux in KPP. In particular, the eddy diffusivity profile

$$\kappa_v = WhG(\sigma), \quad (2)$$

211 with h the boundary layer depth, $G(\sigma)$ the shape function, and $\sigma = -z/h$ the
 212 normalized depth, is directly amplified. The boundary layer depth is determined
 213 as the smallest depth at which the bulk Richardson number reaches a critical
 214 value,

$$Ri_b(h) = \frac{h [b_r - b(h)]}{|\mathbf{u}_r - \mathbf{u}(h)|^2 + U_t^2} = Ri_{cr}, \quad (3)$$

215 with b_r the surface layer averaged buoyancy and $\mathbf{u}_r$ the surface layer averaged
 216 velocity. The critical bulk Richardson number $Ri_{cr} = 0.3$ is empirically de-
 217 termined from observations. Therefore, the boundary layer depth is deepened
 218 as a result of the enhanced turbulent velocity scale through the parameterized

unresolved vertical shear U_t^2 (which is proportional to W , Equation 23 in Large et al. (1994)) in the bulk Richardson number.

McWilliams and Sullivan (2000) propose the enhancement factor as

$$\mathcal{E} = \sqrt{1 + 0.08La_t^{-4}} \approx \sqrt{1 + (1.88La_t)^{-4}} \quad (4)$$

to best fit their LES results.

$$La_t = \sqrt{\frac{u^*}{|\mathbf{u}_s(0)|}} \quad (5)$$

is the turbulent Langmuir number (McWilliams et al., 1997), with $\mathbf{u}_s(0)$ the surface value of Stokes drift velocity. Hereafter we will refer to the parameterization scheme described by (3), (4), and (5) as MS2K.

2.2.2. Smyth et al. (2002) parameterization of VKE

Instead of using a constant coefficient in the enhancement factor (4), Smyth et al. (2002) propose a modified version

$$\mathcal{E} = \sqrt{1 + \mathcal{C}_w(u^*, w^*)La_t^{-4}}, \quad (6)$$

in which

$$\mathcal{C}_w(u^*, w^*) = 0.15 \left(\frac{u^{*3}}{u^{*3} + 0.6w^{*3}} \right)^2 \approx \left(1.61 \left(\frac{u^{*3} + 0.6w^{*3}}{u^{*3}} \right)^{1/2} \right)^{-4}. \quad (7)$$

Here, w^* is the convective velocity scale defined as

$$w^* = (-B_f h)^{1/3}, \quad (8)$$

and B_f is the surface buoyancy flux. This modification enhances the effect of Langmuir turbulence in a stable (wind-forced, positive surface buoyancy flux) condition and reduces it in a convective (negative surface buoyancy flux) condition. Heuristically, this is similar to the effect achieved by the Harcourt and D’Asaro (2008) La_{SL} below. Hereafter, we refer to this parameterization described by (6) and (7) together with (3) and (5) as SS02.

237 *2.2.3. Van Roekel et al. (2012) parameterization of VKE*

238 In both MS2K and SS02, the relative importance of wind induced shear and
 239 Stokes forcing are represented by La_t . Both studies assume that Stokes drift
 240 is aligned with the surface wind stress, which is not always the case. In fact,
 241 the ocean is rarely in wind-wave equilibrium (Hanley et al., 2010). Van Roekel
 242 et al. (2012) propose a projected Langmuir number,

$$La_{proj} = \sqrt{\frac{u^* \cos(\alpha)}{|\mathbf{u}_s(0)| \cos(\theta_{ww} - \alpha)}}, \quad (9)$$

243 to account for the influence by the misalignment of wind and waves. Here α
 244 is the angle between wind and Langmuir cells, and θ_{ww} is the angle between
 245 Stokes drift and wind. Note that wave orientation depends on depth (Webb
 246 and Fox-Kemper, 2015), but here only one wave direction is calculated, which
 247 tends to be more aligned with wind for the La_t definition (9) than the La_{SL}
 248 definition (as defined in (11) below), as surface Stokes drift is more aligned with
 249 wind than depth-integrated Stokes drift (Webb and Fox-Kemper, 2015).

250 Van Roekel et al. (2012) also suggest that a suitable estimation of α could
 251 be made by

$$\alpha_{LOW} \approx \tan^{-1} \left[\frac{\sin(\theta_{ww})}{\frac{u^*}{u_s(0)k} \ln(|H_{ML}/z_1|) + \cos(\theta_{ww})} \right], \quad (10)$$

252 in which H_{ML} represents a relevant depth scale, as discussed below, and z_1
 253 is the onset depth of the law of the wall, which could be taken as four times
 254 the significant wave height (Thorpe, 2007). There are at least two ways of
 255 conceptualizing the relevant depth scale H_{ML} . The first is to estimate the
 256 *mixing depth* to which active turbulence is expected under the present forcing
 257 (as the KPP boundary layer depth). The second is to evaluate a *mixed layer*
 258 *depth*, based on the present stratification of active tracers and thus the historical
 259 depths of turbulent mixing. As noted by Van Roekel et al. (2012), it might
 260 be more realistic to take H_{ML} as the mixing depth (the KPP boundary layer
 261 depth as implemented in our simulations). In our model, the mixed layer depth
 262 is consistently deeper than the mixing depth by as much as 20m in summer

263 and up to 200m differences in winter convection regions. So using mixing depth
 264 here gives a higher estimate of α . However, the relation between the mixing
 265 layer depth and the mixed layer depth could be complicated in other models
 266 and observations (e.g. Noh and Lee, 2008; Sutherland et al., 2014b). When the
 267 Langmuir cells are aligned with wind (i.e., $\alpha \rightarrow 0$), La_{proj} reduces to La_t .

268 Combined with the influence of Stokes drift penetration depth (Harcourt and
 269 D’Asaro, 2008), Van Roekel et al. (2012) propose another Langmuir number,
 270 the surface layer averaged and projected Langmuir number,

$$La_{SL,proj} = \sqrt{\frac{u^* \cos(\alpha)}{|\langle \mathbf{u}_s \rangle_{SL} \cos(\theta_{ww} - \alpha)}}. \quad (11)$$

271 Here $\langle \rangle_{SL}$ denotes an average over the surface layer, which is defined as the
 272 upper 20% of the mixed layer ⁴ by Harcourt and D’Asaro (2008), as confirmed
 273 by Van Roekel et al. (2012).

274 The VKE scaling by Van Roekel et al. (2012) suggests a different formula
 275 for the enhancement factor,

$$\mathcal{E} = |\cos \alpha| \sqrt{1 + (c_1 La_x)^{-2} + (c_2 La_x)^{-4}}, \quad (12)$$

276 in which La_x represents either La_t , La_{proj} or $La_{SL,proj}$. The value of constants
 277 c_1 and c_2 depends on the choice of Langmuir number. Van Roekel et al. (2012)
 278 proposed that for La_{proj} , $c_1 = 3.1$, $c_2 = 5.4$ and for $La_{SL,proj}$, $c_1 = 1.5$, $c_2 = 5.4$.
 279 Reiterate that all parameterization scaling factors are taken from LES results,
 280 and are *not* adapted based on the outcomes in the coarse resolution general
 281 circulation models (GCMs). Thus, no additional tuning occurs to optimize the
 282 GCM results, and indeed tuning the coefficients based on GCM results would
 283 be likely to confuse different model biases, such as Langmuir mixing versus wave
 284 breaking and near-inertial mixing.

285 The parameterization described by (10), (11), and (12), with $La_x = La_{SL,proj}$,
 286 will be referred to as VR12-MA. As a comparison to assess the effects of mis-

⁴Again, we are using the KPP boundary layer instead of the mixed layer in our experiments to be consistent with the choice in α .

aligned wind and waves and Stokes drift penetration depth, an aligned case, VR12-AL, is also tested. It assumes $\alpha = 0$ in the enhancement factor formula in (12) with $La_x = La_{proj}$, which reduces to La_t . In this case the only difference from MS2K is the enhancement factor scaling.

2.2.4. Parameterization of boundary layer depth and entrainment

The boundary layer depth, h , determined from (3), represents the maximum depth that turbulent eddies formed in the boundary layer can penetrate into the stratification given by buoyancy profile, $b(z)$. It is assumed these most deeply penetrating eddies form in the surface layer, which gives them the average buoyancy, b_r , and velocity, $\mathbf{u}_r$, of the layer. For present purposes the Langmuir and Stokes velocities are both assumed to scale with $\mathbf{u}_s(0)$, but are not distinguished, and their effects are represented by including the square of surface Stokes drift $|\mathbf{u}_s(0)|^2$ in the denominator of the bulk Richardson number as

$$Ri_b(z) = \frac{z [b_r - b(z)]}{|\mathbf{u}_r - \mathbf{u}(z)|^2 + U_t^2 + |\mathbf{u}_s(0)|^2}. \quad (13)$$

The boundary layer depth is then determined by matching (13) to the critical Richardson number, Ri_{cr} , as in (3). This parameterization of boundary layer entrainment with the Van Roekel et al. (2012) VKE scaling will be referred to as VR12-EN.

Note that this might overestimate the influence of Stokes shear due to the “anti-Stokes” effect, i.e., in some cases, part of the Eulerian and the Stokes velocity will in fact cancel, leaving a residual velocity whose vertical shear impacts the bulk Richardson number (McWilliams and Fox-Kemper, 2013; Haney et al., 2014). Another source of overestimation is the use of surface Stokes drift instead of a surface layer averaged value, as used for b_r and $\mathbf{u}_r$. In addition, wave effects were present in the observations used to determine $Ri_{cr} = 0.3$ in (3); therefore using the same value in (13) would also tend to overestimate h . However, a compensating factor is that perhaps the surface Stokes drift velocity should be added to $\mathbf{u}_r$ in (13) and then sum squared.

As mentioned before, this set of empirical parameterizations, to include Langmuir mixing in a climate model, is intended as a starting point toward more comprehensive parameterizations (Harcourt, 2013, Grant and Belcher, per. comm.). With these choices, it is easy to turn on the effects of misaligned wind and waves and Stokes drift penetration depth on amplifying VKE, and the effects of Stokes shear on the boundary layer depth in steps to estimate their influences individually. Note that some recent studies, especially McWilliams et al. (2014), propose more sophisticated scalings for entrainment rate. Exploring these alternatives will be left to future work.

2.3. *Experimental setup*

Two sets of experiments are conducted in this study: 1) Wave-Ocean only experiments, with all the parameterization schemes described in the previous section, and 2) fully coupled experiments for CTRL and VR12-MA.

In all the simulations presented here, the wave model WAVEWATCH III is integrated on the G3F25T24 grid, described in Table 1. The ocean model POP2 is integrated on a nominal 3° horizontal grid with 60 vertical levels (referred to as gx3v7 in the CESM terminology). The horizontal resolution is enhanced in the Tropics and high latitudes. The vertical grid spacing is 10 m in the upper 160 m, increasing to 250 m by the depth of around 3500 m, then remaining constant to the bottom at 5500 m.

All of the Wave-Ocean experiments have been integrated for four CORE-II interannual forcing cycles (248 years)⁵. Unless otherwise noted, we define the mean states as the 48-year time-means for model years 200-247, corresponding to year 1961-2008, avoiding the large adjustments associated with the unphysical jump in the forcing from 2009 back to 1948 (Danabasoglu et al., 2014).

In the fully coupled experiments, the Wave-Ocean model is coupled to an

⁵One cycle of CORE-II forcing represents a 62-year period, corresponding to 1948-2009. Note that the CORE-II protocol calls for five forcing cycles. For our purpose here four forcing cycles are adequate.

341 atmosphere model (CAM4) and a land model (CLM4) running on the T31 grid,
 342 and a sea ice model (CICE4) running on the same nominal 3-degree grid as
 343 the ocean model, with a wave-atmosphere-ice-land coupling interval of half an
 344 hour. The wave-ocean coupling frequency remains once per day. Preliminary
 345 fully coupled experiments are integrated for 70 years. The average of the last
 346 20 years are analyzed in this study.

347 **3. Results**

348 *3.1. Vertical turbulent kinetic energy*

349 To illustrate the impact of misaligned wind and waves and the penetration
 350 depth of Stokes drift on scaling the bulk effects of Langmuir mixing, Fig. 2
 351 shows the mixed layer mean VKE normalized by the friction velocity squared,
 352 calculated from the scalings with La_t , La_{proj} (to account for misaligned wind
 353 and waves), La_{SL} (to account for the effect of Stokes drift penetration depth),
 354 and $La_{SL,proj}$ (to account for both), following Van Roekel et al. (2012), for
 355 arbitrarily chosen winter and summer months (Jan. and Jul. of model year
 356 247). In the no wave limit ($La \rightarrow \infty$), the normalized VKE is 0.6, determined
 357 using simulations that only include shear turbulence (see more in Van Roekel
 358 et al., 2012). It is also important to note that misaligned wind and waves could
 359 potentially reduce the normalized VKE in comparison with the no wave case
 360 (leading to a normalized VKE less than 0.6) when the angles between wind
 361 and waves, and therefore the angles between wind and Langmuir cells, are large
 362 (due to the $\cos \alpha$ term in (12)). However, this is rarely seen here presumably as
 363 a result of the relatively coarse spatial resolution used in the ocean and wave
 364 models, which limits refraction by coastal and current features.

365 The effects of misaligned wind and waves and the penetration depth of Stokes
 366 drift could be seen, respectively, by comparing the La_{proj} case and La_{SL} case
 367 with the La_t case. Generally, VKE scaled by La_{proj} is smaller than that scaled
 368 by La_t almost everywhere (Fig. 2b,f versus 2a,e), suggesting that misalignment
 369 of wind and waves prevails in the open ocean, and that scaling by La_t assum-

ing alignment overestimates the Langmuir mixing. Scaling by La_{SL} (Fig. 2c,g
 versus 2a,e) show the enhanced VKE is more vigorous in the summer hemi-
 spheres where the mixed layer depth (MLD) is shallow and less vigorous in the
 winter hemispheres where the MLD is deep. This is because the VKE is sen-
 sitive to the ratio of the penetration depth of Stokes drift to the MLD which
 varies seasonally. The VKE increases monotonically with this ratio, with small
 values occurring for deep MLD (Harcourt and D’Asaro, 2008). These two ef-
 fects are partly canceled in the summer hemisphere but reinforced in the winter
 hemisphere. Therefore, the amount of enhancement in VKE with or without
 the influences by the wind-wave misalignment and the Stokes drift penetration
 depth are similar in the summer hemisphere but very different in the winter
 hemisphere, with much smaller VKE enhancement observed in the $La_{SL,proj}$
 scaling. As in Fan and Griffies (2014), it will be shown that overmixing in the
 winter hemisphere can be problematic.

3.2. Mixed layer depth

The direct effect of including Langmuir mixing in KPP is the deepening of
 the MLD through enhanced surface ocean mixing. In order to directly compare
 the model results with observations, we calculate the MLD following the density
 threshold method in de Boyer Montégut et al. (2004) as the depth where the
 potential density (referenced to surface) changes by 0.03 kg m^{-3} from its surface
 value, rather than using the default MLD definition implemented in POP2 with
 the buoyancy gradient criterion (Large et al., 1997)⁶. It should be noted that
 both definitions are tested and the qualitative effects of Langmuir mixing shown
 here do not depend on the diagnostic definition of MLD. The main reason to
 choose the density criterion definition is to have a matched observation for
 comparison. Different definitions for MLD agree in the extratropical regions⁷,

⁶Note that for both definitions MLD is calculated online at each time step, then averaged
 to output the monthly mean.

⁷For clarity regions poleward of 30°N/S are referred to as extratropical regions, while
 regions between 30°S and 30°N are referred to as equatorial regions.

396 but MLD in the equatorial regions varies among definitions.

397 Fig. 3 and 4 show the summer and winter mean MLD, respectively, from
398 different experiments versus the MLD observed climatology (de Boyer Montégut
399 et al., 2004) with the same density criterion, updated to include the ARGO data
400 to 2012. The subfigures 3b and 4b summarize the latitudinal distribution of root
401 mean square errors (RMSEs) for each experiment. In addition, the percentage
402 change of the summer and winter mean MLDs in VR12-MA and VR12-EN
403 versus CTRL are shown in Fig. 5 to highlight the net effects of Langmuir
404 mixing on MLD.

405 A shallow bias of summer mean MLD in the Southern Ocean is clearly seen in
406 CTRL (Fig. 3a,c and black line in 3b) with no Langmuir mixing effect included.
407 By simply implementing an enhancement factor to the turbulent velocity scale
408 in KPP scaled by La_t as in (4), MS2K generally introduces too much mixing,
409 greatly increasing the RMSE. While the enhanced mixing alleviates the shallow
410 biases in the Southern Ocean, it also introduces a modest deep bias in the equa-
411 torial regions (Fig. 3d and solid gray line in 3b). Accounting for the modulation
412 to the enhancement factor by stratification conditions, i.e., to magnify the en-
413 hancement in stable conditions and restrain it in convective conditions, SS02
414 seems to correct the over-deepening of MLD in the equatorial regions in MS2K,
415 but it causes even more overmixing than MS2K in the extratropical regions
416 (Fig. 3e and dashed gray line in 3b). In the sense of RMSEs (Table 3), MS2K
417 improves the summer MLD simulation in the Southern Ocean but worsens it
418 elsewhere. SS02 generally increases the RMSE of summer MLD everywhere.

419 Van Roekel et al. (2012) considered the effects of both the misaligned wind
420 and waves and the penetration depth of Stokes depth (Harcourt and D’Asaro,
421 2008) in their recently proposed scaling of VKE with $La_{SL,proj}$. Following the
422 idea of applying an enhancement factor to the turbulent velocity scale in KPP,
423 one interesting experiment could be just to use the new VKE scaling, assuming
424 aligned wind and waves (VR12-AL). In VR12-AL, La_{proj} reduces to the La_t ,
425 but the enhancement factor differs from MS2K. As shown in Fig. 3f and the blue
426 line in 3b, the shallow biases of MLD in the Southern Ocean are significantly

reduced by 22% relative to CTRL (Table 3). The over-deepened MLD in the equatorial regions observed in MS2K is also improved in this case. Over the global ocean the RMSE in MLD is reduced by about 15% in comparison with CTRL.

Further accounting for the effects of both the misaligned wind and waves and the penetration depth of Stokes drift (VR12-MA), as shown in Fig. 3g, we get similar results as that in VR12-AL, which is not surprising given the similar Summer Hemisphere VKEs in these two experiments (Fig. 2a,e and 2d,h, see the Southern Hemisphere in Jan. and the Northern Hemisphere in Jul.). There are, however, notable improvements in RMSEs over VR12-MA as expected since more physical realism is included. The MLD bias reductions in VR12-MA relative to CTRL are 27% in the Southern Ocean and 18% globally (Table 3). This change has the same spirit as the entrainment scaling proposed by Li and Garrett (1997), although differs in detail. Uncertainties in RMSEs are also given in Table 3 for CTRL and VR12-MA, by calculating the RMSEs of 1000 bootstrap estimates of the 48-year mean MLD⁸. The fact that the 90% confidence intervals of the summer MLD RMSEs for CTRL and VR12-MA do not overlap indicates that the reductions in summer MLD bias in VR12-MA are significant versus the uncertainty of the mean state estimated by the 48-year time-mean.

Though the exact physics by which Langmuir mixing enhances entrainment is not yet clear, one way to estimate how large this effect could be is to treat Stokes shear as an additional source of the unresolved vertical shear in the definition of bulk Richardson number (VR12-EN; (13)). The resulting summer MLD is shown in Fig. 3h. This is rather an estimation of the upper limit neglecting the “anti-Stokes” Eulerian flow that opposes the Stokes drift (McWilliams

⁸For each bootstrap estimate, 48 years of monthly mean MLD are used. Bootstrapping occurs by choosing 48 versions of each month with replacement, followed by averaging to give an estimate of the mean annual cycle. 1000 versions of this cycle were used to generate the confidence interval.

453 and Fox-Kemper, 2013). It should be noted that the Langmuir mixing induced
 454 deepening of extratropical MLD, especially in the Southern Ocean, is significant
 455 in comparison with the natural variability of MLD, as shown by the dot shading
 456 in Fig. 5. Equatorial deepening is statistically significant in VR12-EN, but not
 457 VR12-MA or VR12-AL.

458 Compared with the observations, substantial biases for the winter mean
 459 MLD in CTRL occur in the extratropical regions in both hemispheres (Fig.
 460 4b). There are shallow biases in the Gulf Stream and Kuroshio Extension re-
 461 gions, the Labrador Sea, and the Pacific and Atlantic sectors of the Southern
 462 Ocean (Fig. 4a,c and black line in 4b). Similar to the summer MLDs, the
 463 winter MLDs get too deep in MS2K and SS02, especially in the extratropical
 464 regions. The improvements brought by SS02, compared with MS2K, is that
 465 SS02 handles the equatorial regions better. Though the reductions of winter
 466 MLD biases in the three versions of the Van Roekel et al. (2012) parameteri-
 467 zation are not as substantial as that for the summer MLD, we do get smaller
 468 RMSEs of winter MLD in both the global ocean and the Southern Ocean, and
 469 these reductions are significant versus the estimated uncertainty of the mean
 470 state (Table 3). These improvements mainly come from the deepening winter
 471 MLD in the Pacific and Atlantic sectors of the Southern Ocean and the North
 472 Pacific, though considerable biases still remain in the North Atlantic, especially
 473 the Labrador Sea. However, unlike MS2K and SS02, the Van Roekel et al.
 474 (2012) parameterizations do not come at a cost of introducing new statistically
 475 significant biases.

476 One interesting result is the similarity in winter mean MLD between VR12-
 477 AL and VR12-MA, given the large differences in the winter hemisphere VKEs
 478 (Fig. 2a,e and 2d,h, see the Southern Hemisphere in Jul. and the Northern
 479 Hemisphere in Jan.). Moreover, the differences between VR12-EN and VR12-
 480 MA are much greater than those between VR12-MA and VR12-AL. This might
 481 be because, as winter MLD is deep, it may be more sensitive to the enhanced
 482 entrainment, which directly deepens the ocean boundary layer, rather than en-
 483 hanced mixing within the already mixed layer. This suggests that improving

484 the representation of entrainment may be critical in reducing the winter MLD
485 biases.

486 The percentage change of the summer and winter mean MLDs in VR12-
487 MA and VR12-EN relative to CTRL is shown in Fig. 5, illustrating the net
488 effects of Langmuir mixing on MLD. By introducing Langmuir mixing in KPP,
489 the MLD is increased globally. Predominantly, the summer mean MLD in the
490 Southern Ocean and winter mean MLD in the North Atlantic and North Pa-
491 cific increase by 20% (VR12-MA) to 60% (VR12-EN). All these changes are
492 significant in terms of the MLD standard deviation ⁹ in CTRL. The degree of
493 relative increase in MLD induced by Langmuir mixing is surprisingly consistent
494 with other studies that use totally different schemes (e.g. D’Asaro et al., 2014),
495 which gives us confidence in the parameterization of Langmuir mixing in KPP
496 in the climate model. D’Asaro et al. (2014) had less summertime deepening,
497 the reasons for which are beyond the scope of the present work.

498 3.3. *pCFC-11*

499 The Langmuir enhanced surface boundary layer mixing should directly affect
500 ventilation to deeper waters with indirect effects on air-sea gas exchange and
501 energy transfer. Ventilation can be quantified by the response of passive tracers
502 with known atmospheric sources, such as the chlorofluorocarbons (CFCs). CFCs
503 enter the ocean through air-sea interaction at the ocean surface and are inert
504 in the ocean. Their concentration history in the atmosphere is relatively well
505 known, and they have been widely measured in the global ocean during the
506 Global Ocean Data Analysis Project (GLODAP; Key et al., 2004). Therefore,
507 they are good indicators of how mixing effects in the upper ocean affect deeper
508 water masses, which makes them ideal for assessing integrated model biases.
509 Here we use the CFC-11 partial pressure distribution (pCFC-11), as it largely
510 eliminates the influences of temperature biases in the CFC-11 concentration.

511 To set up the CFC tracers experiment, CFC concentrations in the atmo-

⁹Based on monthly mean MLD data.

sphere were set near the end of the third CORE-II cycle, at model year 170 (corresponding to data year 1931), and ran through the end of the fourth cycle. The annual mean pCFC-11 of model year 233 (corresponding to data year 1994) is compared with the GLODAP pCFC-11 data.

Fig. 6 shows the zonal mean pCFC-11 (patm) in the Southern Hemisphere for the GLODAP data and zonal mean pCFC-11 anomaly from the observation for different experiments. In comparison with the observations, substantially low concentration biases are found in CTRL, reflecting insufficient ventilation and mixing in the upper ocean. When Langmuir mixing is parameterized in KPP, MS2K and SS02 actually introduce too much mixing, so pCFC-11 concentrations are too high (Not shown). The corresponding RMSEs increase significantly as shown in Table 4. As expected, VR12-AL, VR12-MA, and VR12-EN improve the surface ocean mixing, enhancing the ventilation in the upper ocean and reducing the pCFC-11 concentration biases. The low concentration biases of pCFC-11 in the Southern Hemisphere are reduced by 8%, 6%, and 19% in VR12-AL, VR12-MA, and VR12-EN, respectively. Similar improvements are observed globally (Table 4).

Though it appears from RMSE that VR12-EN is better than VR12-AL and VR12-MA, VR12-EN introduces a high concentration bias at the surface, presumably by working away at the distinction between mixed layer and pycnocline waters. Similar but smaller high concentration biases also exist in VR12-AL, presumably during winter when VR12-AL mixes more strongly than VR12-MA (Fig. 2). Therefore, VR12-MA appears to be the most skillful in terms of enhancing ocean ventilation and reducing the pCFC-11 biases. Overall the Langmuir effects on ventilation and the bias pattern are small, suggesting that boundary layer mixing is not the dominant source of bias; with possible candidates including the surface forcing, the general circulation, the mesoscale eddy parameterization and neglect of anisotropy in particular (Fox-Kemper et al., 2013; Reckinger and Fox-Kemper, 2015; Abernathey et al., 2015).

541 3.4. MLD in the fully coupled experiment

542 The influences of parameterizing Langmuir mixing in KPP on the MLD
543 described above are further examined in the fully coupled experiments. Fig. 7
544 illustrates the summer and winter mean MLD averaged over the last 20 years of
545 the 70-year runs without Langmuir mixing (CTRL) and with Langmuir mixing
546 (VR12-MA). Though the wave effect is much smaller than that in the Wave-
547 Ocean only experiments, the deepening of the summer MLD in the Southern
548 Ocean and winter MLD in the extratropical regions in both hemispheres are
549 clearly seen when Langmuir mixing is included. Similar with analyses in the
550 Wave-Ocean only experiments, we also calculated the RMSEs of summer and
551 winter mean MLD in CTRL and VR12-MA compared with the observations in
552 different regions, as recorded in parentheses in Table 3. In comparison with
553 CTRL, the RMSEs of summer and winter MLD in VR12-MA are reduced by
554 17% and 10% for the Southern ocean, 12% and 8% for the global ocean.

555 In the fully coupled experiments, the wave-atmosphere-ice coupling fre-
556 quency is much higher than that in the Wave-Ocean only experiments (once
557 per 30 minutes versus once per 6 hours). In theory, more short timescale pro-
558 cesses and more Langmuir mixing effects should be captured. However, a smaller
559 impact of Langmuir mixing on MLD is observed. Note that much of the global
560 wave field results from storminess, which elevates the turbulent kinetic energy
561 and generates waves in the surface ocean. Here the wind forcing is from an
562 active atmosphere model (CAM4) running on the T31 grid, and at T31 CAM4
563 resolution storminess will be much weaker than in CORE-II. Therefore, smaller
564 responses in the fully coupled experiments are actually expected. More promi-
565 nent effects of Langmuir mixing should result when the atmosphere model is
566 running on a higher resolution grid (e.g., Kirtman et al., 2011; McClean et al.,
567 2011).

568 Similar with Fan and Griffies (2014), we also find a strengthening of the
569 Atlantic meridional overturning circulation (AMOC) by about 1 Sv (1 Sv $\equiv$
570 $10^6 \text{ m}^3 \text{ s}^{-1}$) with waves, presumably due to enhanced winter deep convection in
571 the polar North Atlantic (Fig. 7c,d). As a result, the Atlantic meridional heat

transport is increased by 0.015 PW ($1 \text{ PW} \equiv 10^{15} \text{ W}$) on average. However, the shallow MLD biases in the Labrador Sea remain in our experiments with Langmuir mixing.

3.5. Temperature and salinity in the fully coupled experiment

Enhanced surface ocean mixing by Langmuir turbulence also affects the surface and subsurface ocean temperature. Fig. 8 shows the maps of errors in simulations of annual mean SST and temperature at the depth of 100 m for CTRL and VR12-MA. The errors are computed as model minus observation, where the observations are from the Polar Science Center Hydrographic Climatology (PHC3) dataset. The PHC3 dataset represents a blending of Levitus et al. (1998) data and Steele et al. (2001) data for the Arctic Ocean. In this section, we define the southern extratropical regions, the equatorial regions and the northern extratropical regions as south of 30°S , 30°S to 30°N and north of 30°N , respectively.

For SST in CTRL, cold biases are clearly observed in the northern extratropical regions, while biases of both signs are observed in the Southern Ocean. Unlike SST, the subsurface temperature shows cold biases over most part of global ocean except in the northern equatorial Pacific and regions in the Southern Ocean, presumably induced by less than enough ventilation over the global ocean. When Langmuir mixing is included (VR12-MA), SST in the extratropical regions get warmer, reducing cold biases in the Northern Hemisphere (RMSE of 2.90°C in VR12-MA versus 3.01°C in CTRL; see Table 5) while increasing both the warm and cold SST biases in the Southern Hemisphere (RMSE of 1.04°C in VR12-MA versus 0.90°C in CTRL). The reduction of cold SST biases in the Northern Hemisphere is mostly related to the deepening of winter MLD there, while the increasing SST biases in the Southern Ocean may come from the biases of the position of winter MLD maximum, which is too far poleward than the observation (Fig. 4 and 7c,d). Although the winter MLD is deepened due to the enhance mixing, the errors in the position of winter MLD maximum remain unfixed. Overall, the impact of Langmuir mixing on the global mean

602 SST is small, with RMSE of 1.54 °C in VR12-MA versus 1.53 °C in CTRL (Ta-
 603 ble 5). However, enhanced surface ocean mixing increases the subsurface ocean
 604 temperature (as illustrated by the temperature at a depth of 100 m), reducing
 605 the cold biases over the global ocean (RMSE of 1.75 °C in VR12-MA versus 1.96
 606 °C in CTRL). The RMSEs in the southern extratropical regions, the equatorial
 607 regions and the northern extratropical regions are reduced by 7%, 17% and 4%,
 608 respectively (Table 5). These improvements come from the warming effects of
 609 Langmuir mixing on the subsurface water, presumably through enhanced ocean
 610 surface mixing and ventilation discussed above.

611 To illustrate the impact of Langmuir mixing on the subsurface ocean tem-
 612 perature, Fig. 9 shows the RMSEs of temperature versus depth computed over
 613 the global ocean (black), the southern extratropical regions (red), the equatorial
 614 regions (blue) and the northern extratropical regions (green) for CTRL (line)
 615 and VR12-MA (dash). The monthly mean ocean temperature in both simu-
 616 lations are compared against the PHC3 monthly climatology. Thus, errors in
 617 seasonality as well as mean state are detectable in Fig. 9. Note that this is
 618 slightly different from that shown in Table 5 and Fig. 8, in which the annual
 619 mean ocean temperature is compared against the observation. Overall, the RM-
 620 SEs of temperature are smaller in VR12-MA than that in CTRL at all depths
 621 except near the surface for the southern extratropical regions and the equato-
 622 rial regions. The most significant improvements appear at the depth of around
 623 100 m. These reductions of RMSE could be divided into reduction of errors
 624 in the annual mean temperature and reduction of errors in the seasonal cycle
 625 of temperature (not shown). It is found that most of the reductions of RMSE
 626 come from the improvement in the annual mean temperature. However, small
 627 improvements in the seasonal cycle are also observed near the surface, especially
 628 for the southern extratropical regions, though in the southern extratropical re-
 629 gions the RMSE of the annual mean temperature is greater when Langmuir
 630 mixing is included.

631 Unlike the warming effects of Langmuir mixing on the subsurface temper-
 632 ature over the global ocean, Langmuir mixing appears to have little influence

on the distribution of salinity (not shown). The errors in the simulations of salinity therefore seems to be unrelated to the missing Langmuir mixing in the climate model. The mechanisms that lead to the errors of salinity in the model are beyond the scope of this study.

4. Summary

In this study we have incorporated a third generation wave model, WAVEWATCH III, into CESM1.2 as an active component model. WAVEWATCH III is two-way coupled with the ocean model POP2 through a modified version of KPP to account for the effects of Langmuir mixing. The impact of Langmuir mixing on the global climate was first assessed in a Wave-Ocean coupled system. The Langmuir mixing is parameterized in KPP by applying an enhancement factor to the turbulent velocity scale with three different scalings following McWilliams and Sullivan (2000), Smyth et al. (2002) and Van Roekel et al. (2012), and subsequently considering the additional physics in the Van Roekel et al. (2012) scalings.

It is found that both MS2K and SS02 introduce too much mixing and over-deepen the MLD in CESM1.2. Though the modulation of the enhancement factor by stratification conditions in SS02 alleviates the over-deepening of MLD in the equatorial regions, it actually degrades in the extratropical regions.

The use of all of the VR12 scalings improves the simulation of both summer and winter MLD. Comparison of the aligned (VR12-AL) and misaligned cases (VR12-MA) shows modest differences in MLD in the Wave-Ocean only simulations. However, smaller RMSEs of MLD are observed in VR12-MA than in VR12-AL. VR12-MA may capture transient winds and waves better than VR12-AL since considerably more physics and sensitivity have been included to agree with LES results of Harcourt and D’Asaro (2008) and Van Roekel et al. (2012), though the responses of the climate model may not appear to be significantly different.

Further estimating the enhanced entrainment at the base of ocean surface

662 boundary layer (VR12-EN) gives prominent deepening of MLD compared with
 663 VR12-MA, highlighting the importance of enhanced entrainment by Langmuir
 664 mixing. However, the estimation of the Langmuir turbulence induced entrain-
 665 ment is only intended as an upper limit. A proper treatment of this effect
 666 requires more theoretical support.

667 The performance of CESM1.2 in terms of surface ocean mixing is improved
 668 by parameterizing Langmuir mixing in KPP, particularly when the effects of
 669 misaligned wind and waves and the penetration depth of Stokes drift are in-
 670 cluded. Specifically in the Wave-Ocean experiments,

- 671 • The shallow biases of MLD in the Southern Ocean are reduced by 27% in
 672 summer and 9% in winter;
- 673 • Ventilation is enhanced and the low concentration biases of pCFC-11 are
 674 reduced in the Southern Hemisphere, though significant low biases persist
 675 with a very similar pattern.

676 Preliminary fully coupled experiments confirm these improvements when
 677 other climate feedbacks are included, though at a reduced level. In addition,
 678 the cold biases in the subsurface ocean temperature are reduced when Langmuir
 679 turbulence is included, presumably through enhanced surface ocean mixing.
 680 Thus, it appears that climate model simulations can be improved by including
 681 a Langmuir mixing parameterization.

682 5. Discussion and Conclusion

683 Parameterizing Langmuir mixing in KPP through a Langmuir number re-
 684 lated enhancement factor to the turbulent velocity scales highly depends on a
 685 LES based VKE scaling, which estimate the bulk effects of Langmuir turbulence
 686 on mixing within the ocean surface boundary layer. Our coupled Wave-Ocean
 687 experiments show that by switching from the MS2K scaling to the VR12 scal-
 688 ing alone, the unrealistic over-deepening of MLD in MS2K and SS02 could be
 689 eliminated. It is worth noting that the over-deepening of MLD by implementing

690 MS2K and SS02 in the GFDL model (Fan and Griffies, 2014) are not as severe
691 as shown here, suggesting some model dependence of parameterizing Langmuir
692 mixing in this way.

693 Unsurprisingly, VR12-MA, in which the effects of the misaligned wind and
694 waves and the penetration depth of Stokes drift are accounted for, improves
695 the surface ocean mixing simulation in the climate model the most. However,
696 it is interesting to note that as more and more physical processes are consid-
697 ered in the scaling (e.g. Harcourt and D’Asaro, 2008; Van Roekel et al., 2012),
698 the resulting amount and distributions of the enhanced surface ocean mixing,
699 and hence deepening in MLD, tend to converge. Newer and more complicated
700 VKE scalings with different Langmuir numbers may be more accurate in the
701 extreme cases, when approaching the upper and lower limits of the parameter
702 space. However, the parameters in the climate model rarely approach those
703 limits. This suggests that if no other new feedback mechanism is discovered,
704 the climate model may be approaching diminishing returns for further improve-
705 ments at greater computational cost in the VKE scaling. There is a large degree
706 of consistency between our results with that of D’Asaro et al. (2014) in terms
707 of the Langmuir turbulence induced deepening in MLD using a completely dif-
708 ferent parameterization of mixing (Harcourt, 2013), which further confirms this
709 convergence.

710 Uncertainty still exists in how to best estimate the depth of the boundary
711 layer, which is a key to capturing the entrainment fluxes. As we have seen
712 in the comparisons among VR12-AL, VR12-MA and VR12-EN, enhanced en-
713 trainment introduces substantial potential deepening in MLD, more so than the
714 effect due to misaligned wind and waves and Stokes drift penetration depth.
715 The surface wave induced entrainment therefore requires more study and better
716 representation. A possible upper limit estimate is given by VR12-EN.

717 Modest variations in the impact of Langmuir mixing shown here might result
718 from using different source terms in WAVEWATCH III. In wave-only testing, a
719 17% greater average enhancement factor results from using Stokes drift gener-
720 ated by WAVEWATCH III using the Tolman and Chalikov (1996) source terms

721 versus the Arduin et al. (2010) source terms, as the latter has La_t 20% larger
 722 on average for the same conditions. These enhancement factor discrepancies
 723 are small in comparison to the difference between the VR12-AL and VR12-MA
 724 parameterization forms shown in Figure 2a,e versus Figure 2d,h, and near the
 725 level of discrepancy between modeled and satellite-inferred Stokes drift magni-
 726 tude (Webb and Fox-Kemper, 2011). The Arduin et al. (2010) source terms
 727 were chosen as they report better significant wave height in swell-dominated
 728 regions. This estimate predicts modest additional mixing in the climate model
 729 if the other source term formulation was chosen.

730 This work focused mainly on the influence of Langmuir mixing on the simu-
 731 lation of physical variables such as the MLD and ocean temperature. However,
 732 Langmuir deepening of MLD may have potential downstream impact on the
 733 biogeochemical modeling. The shallow winter MLD affects the formation of
 734 water masses (Weijer et al., 2012), which leads to weak mixing and ventilation
 735 of the intermediate waters. This may partly contribute to the biases in the
 736 simulated subsurface oxygen in CESM1(BGC)¹⁰ (Moore et al., 2013), and the
 737 low biases of the simulated total oceanic anthropogenic carbon inventory (Wang
 738 et al., 2012; Long et al., 2013). The first hint of potential improvement in such
 739 errors were demonstrated here with CFCs. In addition, the shallow summer
 740 MLD may cause an underestimation of light limitation in a biogeochemistry
 741 model and affect the simulated oceanic carbon uptake by limiting the biological
 742 productivity (Moore et al., 2013; Rodgers et al., 2014). In a global ocean bio-
 743 geochemistry model, Rodgers et al. (2014) found reduced uptake of CO₂ by the
 744 Southern Ocean, and modified seasonal cycle of carbon and other ocean bio-
 745 geochemical tracers, when wind stirring is parameterized and tuned to achieve
 746 best consistency with the observed summer MLD. Similar effects should result
 747 here as the summer MLD is improved through parameterizing Langmuir mix-
 748 ing, though quantitative estimation of these potential biogeochemical impacts
 749 remains outstanding.

¹⁰CESM, version 1, with an ocean biogeochemistry module.

750 This work represents a first step towards more comprehensive parameteri-
751 zations of ocean surface gravity waves in the climate models. There are other
752 wave related processes that require better representation. These include more
753 than just the wind speed related influences of wind waves on the momentum
754 flux, both for the atmosphere through a roughened ocean surface and for the the
755 ocean through the redistribution of wind-driven momentum across the mixed
756 layer (e.g., Sullivan et al., 2004; Garfinkel et al., 2011), the influences of wave
757 induced bubbles on the tracer gas exchange at the air-sea interface and on the
758 ocean surface buoyancy flux (Liang et al., 2011), and the interaction of Langmuir
759 turbulence with the submesoscale eddies (Hamlington et al., 2014).

760 Parameterization of these processes will likely require parameters from a
761 well simulated wave field in the climate model. However, an active wave model
762 can be expensive. In our experiments, the model cost increased by 36% in the
763 Wave-Ocean only experiments and 28% in the fully coupled experiments when
764 Langmuir mixing was included in KPP.¹¹ Since the cost for WAVEWATCH
765 III is constant with a specified grid resolution and coupling frequency (which
766 determines the time step) and since increasing the wave model resolution does
767 not affect Stokes drift significantly, one would expect this number to decrease
768 when the low resolution WAVEWATCH III is coupled with a higher resolution
769 ocean or atmosphere model. Nevertheless, a more efficient but accurate wave
770 model is highly desirable for climate applications.

¹¹The model cost in the Wave-Ocean only experiments are 3.02 and 4.09 pe-hours/simulated-year for CTRL and VR12-MA, respectively. With the nominal 3-degree resolution grid for POP2 and the G3F25T24 grid (see Table 1 for details) for WAVEWATCH III; performed using 16 processors on Yellowstone at NCAR. The model cost in the fully coupled experiments are 40.12 and 51.35 pe-hours/simulated-year for CTRL and VR12-MA, respectively. Note that in the fully coupled experiments, the time step of WAVEWATCH III is one fourth of that in the Wave-Ocean coupled experiments.

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Table 1: Summary of grids for WAVEWATCH III with different spatial and spectral resolutions. N_x is the total number of spatial grid points, N_f and N_θ are the number of frequency and directional bins, respectively.

Grid name	Spatial resolution (N_x)	Latitude range	N_f	Frequency range (Hz)	Increment factor	N_θ	RMSE ($m\ s^{-1}$) ^a	Relative error ^b
G1F25T24	$1^\circ \times 1.25^\circ$ (45216)	$78^\circ S - 78^\circ N$	25	0.0418 - 0.411	1.1	24	-	-
G1F13T12	$1^\circ \times 1.25^\circ$ (45216)	$78^\circ S - 78^\circ N$	13	0.0412 - 0.414	1.212	12	0.049	55.2%
G2F25T24	$1.9^\circ \times 2.5^\circ$ (12096)	$78.85^\circ S - 78.85^\circ N$	25	0.0418 - 0.411	1.1	24	0.0095	10.6%
G3F25T24	$3.2^\circ \times 4^\circ$ (4500)	$78.4^\circ S - 78.4^\circ N$	25	0.0418 - 0.411	1.1	24	0.010	11.2%

^a RMSE of surface Stokes drift, compared with G1F25T24

^b Against the root mean square surface Stokes drift in G1F25T24

Table 2: Summary of parameterization schemes.

Case	Enhancement factor ($\mathcal{E}$)	Unresolved vertical shear ^a	Scaling	Description
MS2K	$\sqrt{1 + (1.88La_t)^{-4}}$	U_t^2	MS2K	Aligned wind and waves
SS02	$\sqrt{1 + \left(1.61 \left(\frac{u_*^3 + 0.6u_*^3}{u_*^3}\right)^{1/2} La_t\right)^{-4}}$	U_t^2	MS2K	MS2K + Stable/Convective conditions
VR12-AL	$\sqrt{1 + (3.1La_t)^{-2} + (5.4La_t)^{-4}}$	U_t^2	VR12	Aligned wind and waves
VR12-MA	$ \cos\alpha \sqrt{1 + (1.5La_{SL,proj})^{-2} + (5.4La_{SL,proj})^{-4}}$	U_t^2	VR12	Misaligned wind and waves, Stokes drift penetration depth
VR12-EN	as VR12-MA	$U_t^2 + \mathbf{u}_s(0) ^2$	VR12	VR12-MA + Enhanced entrainment

^a U_t^2 is the parameterized unresolved vertical shear in the bulk Richardson number definition (Eq. 23 in Large et al., 1994), with greater value indicating stronger entrainment from below the ocean surface boundary layer.

Table 3: Root mean square errors (RMSE, m) of summer and winter mean mixed layer depth in comparison with observation (de Boyer Montégut et al. (2004), updated to include the ARGO data to 2012).

Case	Summer			Winter		
	Global	South of 30°S	30°S-30°N	Global	South of 30°S	30°S-30°N
CTRL	10.62±0.27 ^a (13.40±0.19) ^b	17.24±0.48 (21.73±0.32)	5.38±0.14 (6.71±0.09)	43.85±0.38 (45.50±0.40)	57.19±0.76 (56.53±0.59)	12.57±0.28 (16.16±0.29)
MS2K	15.37	15.47	17.03	119.91	171.92	40.31
SS02	36.79	63.83	7.54	99.32	164.34	17.39
VR12-AL	9.06	13.47	6.49	40.45	50.33	14.52
VR12-MA	8.73±0.30 (11.83±0.29)	12.65±0.47 (18.13±0.62)	6.61±0.22 (7.52±0.16)	40.99±0.37 (42.02±0.39)	51.78±0.65 (50.78±0.67)	14.23±0.30 (15.67±0.35)
VR12-EN	8.95	10.52	8.91	41.94	52.98	19.58

^a Numbers with $\pm$ sign give the 90% confidence interval, estimated from the RMSEs of $n_b = 1000$ bootstrap estimates of the 48-year (for Wave-Ocean only experiments) and 20-year (for fully coupled experiments) mean mixed layer depth.

^b Numbers shown in the parentheses are for the fully coupled experiments.

Table 4: Root mean square difference (patm) of zonal averaged pCFC-11 compared with observation (GLODAP; Key et al. (2004)).

Case	Global	Southern Hemisphere
CTRL	23.90	20.20
MS2K	29.89	30.99
SS02	34.16	41.90
VR12-AL	22.14	18.53
VR12-MA	23.23	18.90
VR12-EN	20.67	16.44

Table 5: Root mean square errors (RMSE, °C) of the 20-year averaged ocean temperature at the surface and at the depth of 100 m in the fully coupled experiment, compared with observation (PHC3, Steele et al., 2001).

Case (depth)	Global	90°S - 30°S	30°S - 30°N	30°N - 90°N
CTRL (0 m)	1.53±0.02 ^a	0.90±0.02	1.10±0.02	3.01±0.05
VR12-MA (0 m)	1.54±0.01	1.04±0.03	1.14±0.01	2.90±0.04
CTRL (100 m)	1.96±0.02	1.39±0.01	1.92±0.02	2.88±0.04
VR12-MA (100 m)	1.75±0.02	1.29±0.01	1.60±0.02	2.76±0.04

^a Numbers with $\pm$ sign give the 90% confidence interval, estimated from the RMSEs of $n_b = 1000$ bootstrap estimates of the 20-year mean surface and subsurface temperature.

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Figure 1: Schematic diagram of WAVEWATCH III coupling in CESM1.2. The arrows in color show the variable flows passing between WAVEWATCH III and other model components, with $\mathbf{U}_0$ and T_0 the surface wind and temperature in the atmosphere; $\mathbf{u}(0)$ the surface current in the ocean; H_{BL} the ocean surface boundary layer depth; La_x the Langmuir number (La_t or $La_{SL,proj}$); α_{LOW} the angle between the wind and Langmuir cells and $\mathbf{u}_s(0)$ the surface Stokes drift. The black dash arrows indicate all other variable flows passing between the ocean component and other model components. The coupling frequency is labeled in parentheses for each model component. Note the different coupling frequencies of wave-atmosphere coupling and wave-ocean coupling. See text for more detail.

Figure 2: The mixed layer averaged vertical turbulent kinetic energy (VKE) normalized by the friction velocity squared, predicted from the scalings with different Langmuir numbers, showing the effects of misaligned wind and waves and Stokes drift penetration depth. The normalized VKEs are calculated from scalings with (a)(e) La_t ; (b)(f) La_{proj} ; (c)(g) La_{SL} and (d)(h) $La_{SL,proj}$, following Van Roekel et al. (2012), averaged over Jan. (a) - (d) and Jul. (e) - (h) of the model year 247. In the limit of no waves ($La \rightarrow \infty$), the normalized VKE is 0.6, determined from a shear turbulence only simulation (Van Roekel et al., 2012). Note that (a)(e) and (d)(h) are corresponding to experiment VR12-AL and VR12-MA, respectively.

Figure 3: Impact of Langmuir mixing on the summer mean mixed layer depth (MLD; m) for both hemispheres. (a) shows the observation from de Boyer Montégut et al. (2004), updated to include the ARGO data to 2012. (c) shows the control run without Langmuir mixing. (d) - (h) are results with Langmuir mixing implemented in different parameterization schemes (See Table 2 for description). MLDs are averaged over Jul., Aug. and Sep. (JAS) for the Northern Hemisphere (NH) and Jan., Feb. and Mar. (JFM) for the Southern Hemisphere (SH). (b) shows the latitudinal distribution of root mean square errors.

Figure 4: As Fig. 3, but for the winter mixed layer depth. Averaged over JFM for the NH and JAS for the SH.

Figure 5: Langmuir mixing induced (a)(c) summer and (b)(d) winter mean mixed layer depth (MLD) deepening in (a)(b) VR12-MA and (c)(d) VR12-EN, shown by the percentage change (%) in comparison with CTRL. Regions where MLD changes are significant (i.e. greater than the MLD standard deviation based on monthly data in CTRL) are shaded by dots.

Figure 6: Impact of Langmuir mixing on the zonal mean pCFC-11 (patm) in the Southern Hemisphere. (a) shows the observation (Key et al., 2004). (b) - (e) show the anomaly from the observation for the cases CTRL, VR12-AL, VR12-MA and VR12-EN, respectively. Note the continuous reduction of low concentration biases as more physics are included and the emergence of high concentration biases at the surface in VR12-AL and VR12-EN.

Figure 7: As Fig. 3 and Fig. 4, but for the fully coupled experiments. (a)(b) Summer and (c)(d) winter mean mixed layer depth (m) for experiments: (a)(c) without Langmuir mixing (CTRL); (b)(d) with Langmuir mixing accounting for the misaligned wind and waves and the Stokes drift penetration depth (VR12-MA).

Figure 8: Maps of errors in the simulations of annual mean SST (upper panels) and ocean temperature at depth of 100 m (lower panels) in CTRL and VR12-MA. The errors are computed as model minus observation, where the observations are from the PHC3 dataset.

Figure 9: RMSEs of temperature versus depth. RMSEs are computed over the global ocean (black), south of 30°S (red), 30°S to 30°N (blue) and north of 30°N (green) for CTRL (line) and VR12-MA (dash). Numbers in parentheses are total RMSEs over the upper 500 m. The monthly mean ocean temperature in both experiments are compared against the PHC3 monthly climatology.