Effects of symmetric instability in the Kuroshio Extension region in winter

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Abstract

Submesoscale symmetric instability (SI) in the ocean surface mixed layer (SML) plays a significant role in the forward energy cascade and vertical material transports. Due to its small spatial scales of typically less than 1 km, SI is not resolved by current climate ocean models and most regional models. This work investigates the SI effects in the Kuroshio Extension (KE) region by using a SI parameterization scheme implemented in the Coastal and Regional Ocean Community Model (CROCO). Compared to an SI-lacking model at the same spatial resolution of 500 m, an SI-parameterized model more closely resembles an ultra-high resolution (20 m) non-hydrostatic model that is intended to resolve SI directly on the basis of evaluating the energy source of SI: geostrophic shear production. As SI restratifies the SML, the SI-parameterized model simulates a shallower SML depth in the KE region compared to the SI-lacking one, which is also closer to the observed state. The potential vorticity is also modulated and a reduction of negative PV is found due to parameterized SI. The results here demonstrate the capability of the SI scheme in the representation of the SI impacts and an improvement in the simulation of the KE due to the SI scheme.
The ocean surface mixed layer (SML) is the layer where air-sea exchanges are captured and communicated to the deeper ocean. These exchanges (such as momentum and buoyancy fluxes) lead to intense dynamic processes in the SML. These processes generate strong turbulence in the SML and modulate the air-sea exchanges in turn. Due to the thinness of the SML, its dynamical processes, such as effects from surface waves, Langmuir circulation, boundary layer turbulence structures, and submesoscales usually have small spatial scales in both the vertical and horizontal direction (from tens of meters to kilometers). These scales are hardly resolved by current ocean climate and regional models. As a result, including their effects requires parameterization in most ocean models (e.g., Fox-Kemper et al., 2008, 2011; Li et al., 2016; Noh et al., 2016; Qiao et al., 2016; Dong et al., 2021b). As one kind of active process in the SML, submesoscale currents are defined as those with Rossby number and Richardson number around 1 (i.e., $Ro \sim 1$, $Ri \sim 1$) translating into spatio-temporal scales of $O(0.01 \sim 10)$ km, $O(1)$ day (Thomas et al., 2008; McWilliams, 2016). The importance of submesoscales in the ocean transport (Su et al., 2018; Wenegrat et al., 2020) and the ocean energy cascade (D’Asaro et al., 2011; Sasaki et al., 2017; Dong et al., 2020a; Cao et al., 2021) has been widely reported.

Mixed layer instability (MLI) and symmetric instability (SI) are two major instabilities of mesoscale and submesoscale currents generating smaller submesoscales in the SML (Haine et al., 1998; Boccaletti et al., 2007; Thomas et al., 2013). According to Dong et al. (2020b; 2021a), SI tends to have spatial scales one order of magnitude smaller compared to MLI globally. MLI is (partially) resolved by in-situ observations, high-resolution regional models, and even global models (e.g., MITgcm LLC4320 with a nominal resolution of 2 km; Su et al., 2018). As a result, MLI has been intensively studied in the past decade, and its role in vertical transport, inverse cascade, and SML restratification have been investigated (e.g., Capet et al., 2008; Fox-Kemper et al., 2008; Callies et al., 2016; Schubert et al., 2020; Zhang et al., 2021; Cao et al., 2021; Cao and Jing, 2022). By contrast, SI is hardly directly observed or simulated due to its smaller
spatial scales [around O(100) m; Dong et al., 2021a]. The existence of SI from observations is usually demonstrated by inference from its conditions for existence--
the presence of negative potential vorticity (PV)--and co-located elevated turbulence dissipation rates (e.g., D’Asaro et al., 2011; Thomas et al., 2013; Buckingham et al.,
2019; Yu et al., 2019; Zhang et al., 2021).

MLI feeds on the release of available potential energy and tends to transfer its kinetic energy inversely to larger scales (Callies et al., 2016; Dong et al., 2020; Schubert et al., 2020). By contrast, SI grows by extracting energy through geostrophic shear production (GSP) and leads to a downscale transfer of kinetic energy (i.e., forward cascade; Taylor and Ferrari, 2009; Bachman et al., 2017; Dong et al., 2021). SI can occur when the Ertel PV \( q = (f \mathbf{k} + \nabla \times \mathbf{u}) \cdot \nabla b \) is anticyclonic, i.e., has an opposite sign to the local Coriolis parameter (Hoskins, 1974; Haine and Marshall, 1998; Haney et al., 2015). For a strong geostrophic front in the SML, the baroclinic component can shift the PV to anticyclonic, as the stratification is weak so the vertical component is easily overcome. Once anticyclonic PV arises, SI can quickly restore PV to a neutral state (i.e., zero) through mixing highly cyclonic PV (and other tracers) from the pycnocline. **Forced** SI persists under continued surface buoyancy flux injecting anticyclonic PV (Thomas and Taylor, 2010).

Although SI significantly affects SML dynamics, including the energy cascade and material transports, its quantitative effects are difficult to evaluate in the real ocean due to its small spatial scales. Parameterizing SI in ocean models is a less costly choice to evaluate its impacts. However, SI parameterizations are not commonly in use (Buckingham et al., 2019; Dong et al., 2021), although many parameterizations have been proposed (e.g., Lindstrom and Nordeng, 1992; Balasubramanian and Yau, 1994; Fei et al., 2011; Bachman et al., 2017).

The Kuroshio is one of the strongest western boundary currents and the Kuroshio Extension (KE) region is a place where strong air-sea and multi-scale interactions occur due to this strong, warm current. Previous works have demonstrated the KE contains active submesoscales which deeply modulate the local mesoscale variability (Sasaki et
al., 2017; Cao et al., 2021), mixing (D’Asaro et al., 2011), and air-sea interaction (Yang et al., 2021). According to Dong et al. (2021), the KE is anticipated to be a hot spot for SI occurrence. Nevertheless, the SI effects in the region have yet to be quantified. Given that an evaluation of SI effects by directly resolving it is impractical (the spatial resolution should be as fine as tens of meters; cf. Section 3), using a parameterization scheme becomes an alternate choice for the evaluation. In this work, the effects of SI in the KE are quantitatively investigated by implementing the SI parameterization scheme proposed by Bachman et al. (2017) in the Coastal and Regional Ocean Community Model (CROCO). Dong et al. (2021b) have evaluated the performance of the SI scheme by investigating its performance on idealized frontal configurations. The performance of the scheme in the realistic regional ocean will be evaluated in this work.

The paper is organized as follows: Section 2 gives a brief introduction of the SI parameterization and describes the model setup; Section 3 presents the main results of the evaluation of SI effects in the KE. The last section is the discussion and conclusions.

### 2 Symmetric instability parameterization and model setup

#### 2.1 Symmetric instability parameterization

This section gives a brief description of the parameterization scheme proposed by Bachman et al. (2017). An anticyclonic PV is a precondition for SI, it is convenient to define an anticyclonic-PV layer using a bulk PV formula,

\[
\tilde{f}q_{\text{bulk}} = f \left( f \Delta b + \langle \zeta \rangle \Delta b + \Delta u \left( \frac{\partial b}{\partial y} \right) - \Delta v \left( \frac{\partial b}{\partial x} \right) \right).
\]

The layer thickness, \(H\), is calculated as the depth below which \(\tilde{f}q_{\text{bulk}}\) becomes positive, i.e., \(\tilde{f}q_{\text{bulk}} < 0\), the deepest penetration depth of the unstable SI modes (Haney et al., 2015; Dong et al., 2020b). Here, \(\zeta\) is the vertical relative vorticity, \(\Delta\) denotes the difference between the surface value and the value at a given depth, and the angle bracket denotes a depth average over the same depth range. For a high-resolution model (such as the model used here), the vertical vorticity tends to be comparable to the local
planetary vorticity, so we include the impacts of the vertical vorticity as suggested by Dong et al. (2020b) unlike the suggestion of B17.

The injection of PV from surface forcing is necessary for forced SI, (Thomas, 2005), which includes sea surface heat and freshwater exchanges combined into the buoyancy forcing,

$$B_0 = B_T + B_S = g\alpha \frac{Q_{\text{net, heat}}}{\rho_0 C_p} + g\beta (EP) S,$$

combined with the Ekman Buoyancy Flux driven by along-front wind at the surface,

$$EBF = \frac{\tau_w \times k}{\rho_0 f} \cdot \nabla B.$$  \(\text{(3)}\)

Here, $\alpha$ (unit: $^\circ \text{C}^{-1}$) is the thermal expansion coefficient, $Q_{\text{net, heat}}$ (unit: W m$^{-2}$) is the net surface heat flux, $C_p = 3800$ J kg$^{-1} \cdot ^\circ \text{C}^{-1}$ is the seawater specific heat capacity, $\beta$ (unit: PSU$^{-1}$) is the saline contraction coefficient, $EP$ (unit: m s$^{-1}$) is the net freshwater exchange due to evaporation and precipitation (vertical convection occurs when the ocean loses heat, indicated with positive $Q_{\text{net, heat}}$ or freshwater with positive $EP$), and $S$ is the sea surface salinity, and $\tau_w$ is the wind stress. PV tends to be shifted toward anticyclonic values when $F_{SI} = B_0 + EBF > 0$.

Two distinct sublayers exist in anticyclonic-PV layer when $B_0 > 0$ and $EBF > 0$: a convective layer near the surface and a deeper SI-dominated layer below (e.g., Taylor and Ferrari, 2010). The convective layer thickness, $h$ is determined by solving a quartic equation (Taylor and Ferrari, 2010; Thomas et al., 2013),

$$\left(\frac{h}{H}\right)^4 - c^3 \left(1 - \frac{h}{H}\right)^3 \left[\frac{w^2}{U^2} + \frac{u^2}{U^2} \cos \theta_w\right] = 0.$$  \(\text{(4)}\)

Here, $c = 14$ is an empirical constant, $w_\ast = (B_0 H)^{1/3}$ is the convective velocity by surface buoyancy loss, $u_\ast = (|\tau_w|/\rho_0)^{1/2}$ is the friction velocity, $U$ is the velocity difference over the thickness of SI layer $H$, and $\theta_w$ is the angle between wind vectors and frontal currents.

Once $H$ and $h$ are known, the energy source for SI, GSP is parameterized as a piecewise linear function (Thomas et al., 2013),
The corresponding vertical viscosity and diffusivity are calculated as,

\[ \nu_{SI} = \frac{f^2}{|\nu_b|} GSP, \quad (6) \]

and

\[ \kappa_{SI} = \frac{2\nu_{SI}}{1 + (10 \max(0, Ri_b))^{0.8}}. \quad (7) \]

assuming mixing takes on a simple form of the turbulent Prandtl number (i.e., \( Pr = \frac{\nu_{SI}}{\kappa_{SI}} \)). Here, \( Ri_b \) is the balanced Richardson number. Based on the parameterized variables, the scheme is easily applied in the CROCO.

Note that the SI scheme is active only within frontal regions under destabilizing surface forcing. So, the SI parameterization is implemented alongside another default boundary layer turbulence closure, such as the nonlocal K-Profile Parameterization (KPP; Large et al., 1994) scheme used here, so that when the SI conditions are not met and SI is stable, then the other scheme is used. In the practice as used here, the SI scheme will be activated only when \( B_0 > 0 \) and \( EBF > 0 \) and \( h < 0.95H \) (Dong et al., 2021b). Strictly speaking, forced SI requires only \( B_0 + EBF > 0 \). It is still unclear about the SI dissipation when \( B_0 \) and EBF are of opposite sign and further studies are needed in the future to improve the parameterization. When the SI scheme is activated, the vertical flux, \( F_{wc} \) for tracers with nonzero surface fluxes in the convective layer needs to be considered, which can be parameterized as a linear profile,

\[ F_{wc} = \left\{ \begin{array}{ll}
0, & z = 0 \\
F_0 \frac{z + h}{h}, & -h \leq z < 0 \\
0, & z < -h
\end{array} \right. \quad (8) \]

Here, \( F_0 \) is the surface flux for a given tracer, \( C \). Steps to apply the SI scheme in the CROCO have also been summarized by Dong et al. (2021b). In conclusion, the KPP
is the default scheme in this work, and the viscosity and diffusivity values will be replaced when the SI scheme is activated.

Finally, the momentum and tracer equations that include the SI scheme in the z-coordinate can be rewritten as,

\[
\begin{align*}
\frac{\partial u}{\partial t} + \nabla \cdot (uv) - fu &= -\frac{\partial \phi}{\partial x} + \frac{\partial \nu_{sl}}{\partial z} \frac{\partial u}{\partial z} + F_u + D_u \\
\frac{\partial v}{\partial t} + \nabla \cdot (vv) + fu &= -\frac{\partial \phi}{\partial y} + \frac{\partial \nu_{sl}}{\partial z} \frac{\partial v}{\partial z} + F_v + D_v \\
\frac{\partial C}{\partial t} + \nabla \cdot (VC) &= \frac{\partial \nu_{sl}}{\partial z} \frac{\partial C}{\partial z} + \nabla \cdot (K_{sl}C) - \frac{\partial F_{wc}}{\partial z} + F_c + D_c.
\end{align*}
\]

Here, \( \nabla = \frac{\partial}{\partial x} \mathbf{i} + \frac{\partial}{\partial y} \mathbf{j} + \frac{\partial}{\partial z} \mathbf{k} \) is the three-dimensional gradient operator, \( \mathbf{v} \) is the three-dimensional velocity field, \( \phi \) is the pressure potential, \( F \) and \( D \) are the corresponding forcing and horizontal diffusive terms, respectively. Here, \( K_{sl} \) is a symmetric tensor in Redi (1982) form in the z-coordinate representing the along-isopycnal diffusivity (Equation 6).

### 2.2 Model Setup

The parameterization and simulation are built with the CROCO (http://www.croco-ocean.org). The CROCO is a new model built upon the ROMS-Agrif (Jullien et al., 2019). The model covers the KE and has a spatial resolution of 500 m and 60 vertical layers with refined vertical level thicknesses in the SML (the averaged layer thickness is 4 m in the upper 50 m; Figure 1). The climatological daily-mean surface atmospheric forcing is applied for this model simulation. The wind stress is provided by the Quick Scatterometer (QuikSCAT) dataset with a spatial resolution of 0.25°, while the buoyancy fluxes including heat and freshwater are from the monthly climatology of the International Comprehensive Ocean–Atmosphere Data Set (ICOADS) with a resolution of 1° (Woodruff et al., 2011). The initial condition and boundary fluxes are derived from successively downscaled one-way nested models, i.e., from a parent model with a 7.5-km resolution to an intermediate model with a 1.5-km resolution (Figure 1a).
The boundary and initial information driving the parent domain are from the monthly averaged Simple Ocean Data Assimilation (SODA) ocean climatology outputs with a spatial resolution of 0.5° (Carton and Giese 2008). The parent model is spun up from its initial state for 20 years and then run for 2 years to provide daily boundary information for the intermediate model which is run for one year. The models have been validated using in-situ and satellite observations and have been widely used for nesting to investigate submesoscales (e.g., Luo et al., 2020; Cao et al., 2021; Jing et al., 2021; Cao and Jing, 2022). More details about the model setup can also be found in Luo et al. (2020) and Jing et al. (2021). As SI is more active in winter in the KE (Dong et al., 2021a), the nested 500 m model is run for one month (February).

The 500 m resolution is believed to resolve MLI in the SML of the KE according to an estimate of the local MLI scale (Dong et al., 2020b). However, the SI scale is below 1 km (Dong et al., 2021a) which is hardly resolved by the model. To investigate the effects of SI, two contrast simulations are conducted. The two cases share the same configurations, except for the turbulence closure. One is simulated by solely using the KPP for the subgrid vertical mixing (hereinafter, KPP500), while the SI scheme is activated alongside the default KPP scheme in the other one (hereinafter, SI500). Moreover, to directly evaluate SI and the performance of the SI scheme, an ultra-high-resolution non-hydrostatic model with a spatial resolution of 20 m is built in a still smaller nested domain (hereinafter, NBQ20; Figure 1b). Note that this domain is too small to capture much of the SI statistics across the region, so it is in a combination of the SI500 and NBQ20 models versus the KPP500 model to reveal the regional SI effect.

To meet the requirement of $\frac{\partial v}{\partial z} \sim \frac{N}{f}$ for a high-resolution model to resolve small-scale geophysical processes (SI here; Menesguen et al., 2018), the NBQ20 case has 150 vertical layers with finer resolution in the SML. The NBQ20 is nested from a 100 m resolution model which is downscaled from the KPP500 case. Due to computational limits, the domain of the NBQ20 is 12 km $\times$ 12 km. Due to their small spatial scale, the domain is sufficient to contain and resolve SI partially [theoretically, SI has no fastest growing mode according to Stone (1966), and no smallest unstable scale, but the larger
scales are expected to dominate the geostrophic energy extraction and transport]. The KPP scheme is also applied in the NBQ20. The NBQ20 is firstly run for 0.5 days in a hydrostatic mode and then run for 1 hour with non-hydrostatic mode turned on in the early morning on 3 Feb with output every 5 min. The early morning conditions ensure buoyancy loss of the ocean and favor forced SI.

![Figure 1](image.png)

Figure 1 The setup of the nested models. Domains of the 7.5 km, 1.5 km and 500 m resolution models are shown in (a), while domains of the 100 m and 20 m resolution models are shown in (b). The color contours are the instantaneous Rossby number (i.e., the vertical relative vorticity normalized by the local planetary vorticity) calculated from the corresponding model cases on the same day (the Rossby number in panel b is from the KPP500). The Rossby number in the tropical region, e.g., between 10°S and 10°N, is excluded here.

### 2.3 Observations

Hydrographic profiles including temperature, salinity and pressure from the Argo program in Februaries from 2010 to 2019 over the KE (the 500-m model domain) are used as an observational basis for the SML depth comparison. The data are provided by the NOAA National Centers for Environmental Information (https://www.nodc.noaa.gov/argo/basins_data.htm). Each Argo float is equipped with sensors of conductivity, temperature, and pressure, ascends to the sea surface every 10 days, and transmits the measured parameters to the satellites (http://www.argo.ucsd.edu). The profiles are filtered with vertical grid spacings less than 2 m and the shallowest level above 5 m depth. Then a meridional section is obtained by averaging these profiles zonally.
3 Results

In this section, the SI effects are evaluated from three different aspects, including the GSP, SML depth and PV. According to the Rossby number (Ro) distribution from the KPP500 shown in Figure 1b, active submesoscales with Ro~1 are frequently observed. The activity of submesoscales and its role in the KE has been well studied (e.g., Rocha et al., 2016; Dong et al., 2020a; Cao et al., 2021). However, these resolved submesoscales are not the focus of this work, as they derive from the larger-scale processes of MLI and frontogenesis. In this work, these larger submesoscales are not discussed and the SI effects are focused on.

3.1 Geostrophic shear production

The NBQ20 is taken as the “truth” case which is believed to resolve SI and other smaller submesoscales and larger boundary layer turbulence directly. These phenomena, called SI perturbations from here on for simplicity, are obtained by applying a highpass filter with a cutoff scale of 1 km in the NBQ20 results. The chosen cutoff scale is near the Nyquist sampling rate of the coarse-resolution cases (i.e., 500 m cases), indicating that the perturbations are not resolved by the coarse-resolution cases. The original fields at the surface from the NBQ20 show fronts and strong velocity shear are observed at the fronts (Figure 2a, c, e). The velocity and temperature perturbations are mainly concentrated in the frontal regions, with magnitudes up to 0.03 m s\(^{-1}\) and 0.1 °C (Figure 2b, d, f). The horizontal velocity perturbations show remarkable alternating patterns typical of SI (Haney et al., 2015; Dong et al., 2021b). The vertical modes of the velocity perturbations along the two sections are shown in Figure 3. Outcropping of isopycnals in the SML is explicitly observed, indicating strong SML fronts along the sections. Furthermore, alternating stripes of the velocity perturbations accompany the fronts. The vertical mode of these perturbations is contained within the SML, and it slants across the front, all suggesting that the perturbations are dominated by SI.
Figure 2 Snapshots of (a, b, c, d) surface currents (m s$^{-1}$) and (e, f) temperature ($^\circ$C) from the NBQ20. The upper panels show the original fields while the lower for the perturbations. The first and second columns show the zonal and meridional velocities, respectively. The black lines denote two sections, S1 (meridional) and S2 (zonal) for further analysis.

Figure 3 The horizontal velocity perturbations (m s$^{-1}$) along (a, b) S1 and (c, d) S2. The upper panels show the zonal component while the lower for the meridional. The black lines denote the isopycnals.

Taking the NBQ20 as the “truth”, the SI energy source, GSP, is investigated and compared. First, the velocity and buoyancy fields are decomposed into the background, SI components, namely,
\[
\begin{align*}
\mathbf{u} &= \mathbf{u}_b + \mathbf{u}', \\
w &= w_b + w', \\
b &= b_b + b'.
\end{align*}
\]

Here, the SI component is calculated by a highpass filter with a cutoff scale of 1 km (this scale is below the scale that the coarse-resolution cases can resolve, so the SI component is expected to be zero for those cases), the background component is the residual. The geostrophic component is determined based on the thermal wind balance \( f \mathbf{k} \times \frac{\partial \mathbf{u}_g}{\partial z} = -\nabla h b_b \).

Based on the decomposition, the GSP is calculated as (with overbar denoting spatio-temporal averages, and the whole equation is calculated as an integral over the horizontal),

\[
GSP = K_v \left( \frac{\partial \mathbf{u}_g}{\partial z} \right)^2 - \overline{\mathbf{u}' \mathbf{w}'} \frac{\partial \mathbf{u}_g}{\partial z},
\]

where \( K_v \) is the model output vertical viscosity (the parameterized GSP is implicitly expressed in \( K_v \) in the SI500; cf. Equation 6). It’s noteworthy that in the coarse-resolution simulations KPP500 and SI500 this viscosity is intended to capture the full GSP by SI, and the second production term should be near zero (Bachman et al., 2017; Dong et al., 2021b). The domain-averaged (the same domain as NBQ20) GSP from the three simulations is shown in Figure 4. All the GSP profiles vary similarly in vertical: increase first as the depth increase but then decrease to null below the SML base. However, the magnitudes are different. Compared to the KPP500, the GSP from the SI500 is strengthened near the surface but weakened below 20 m (black and red lines in Figure 4). In addition to the quantitative difference, the SI500 GSP resembles more the NBQ20 one. The NBQ20 GSP is also generally weak compared to the KPP500, but strong near the surface. This GSP correction due to SI resembles Bachman et al. (2017). However, note the differences between the SI500 and NBQ20. The peak of the NBQ20 GSP is in the shallow layer above 20 m. The peak of the SI500 GSP is still below 20 m, despite that the magnitude of the SI500 GSP is weakened. Possible reasons for the bias are discussed in the last section.
Figure 4 GSP profiles from the three simulation cases, KPP500 (black line), SI500 (red line), and NBQ20 (blue line). The GSP from the KPP500 and SI500 is averaged over the same region and period as the NBQ20.

### 3.2 Surface mixed layer depth

SI tends to restore the PV to a neutral state by mixing the cyclonic PV into the SML, leading to SML restratification. In other words, a model that includes SI effects tends to simulate a shallower SML depth compared to a no-SI model. The SML depth is calculated as the depth at which the surface potential temperature decreases by a value of 0.2 °C (de Boyer Montégut et al., 2004). The temporally-averaged SML depth in February from the KPP500 and SI500 is shown in Figure 5a and b. The KE has deep SML depths up to 300 m. In general, the pattern of the SML depth from the SI500 is close to the one from KPP500, indicating that the change of the SML depth due to SI restratification is not very significant when compared to the background spatiotemporal variability of the SML depth. The change is highlighted by their differences (Figure 5c). The change of the SML depth is up to 50 m in certain regions. It is also noted that the
differences are not always negative due to SI, and positive values (i.e., deepening the
SML) are frequently observed over the domain. The change of the SML depth is a
systemic result in the simulation, since SI can not only restratify the SML, but also
modulate the SML dynamics (such as enhancing the SML dissipation, altering fronts
and modulating the MLI; Dong et al., 2021a). These effects also potentially lead to SML
depth change.

Nevertheless, the change of the SML due to SI is mainly concentrated in the frontal
regions (simply taken as the regions where strong variabilities of the SML depth are
observed). Also, an overall shoaling of the SML is demonstrated by spatially averaging.
Considering that the SML in the KE shoals poleward, the simulated SML is averaged
zonally (Figure 6). Compared to the observed SML depth (black line in Figure 6), both
the KPP500 and SI500 simulate deep SML up to 200 m in the south KE but shallow in
the north KE less than 60 m (blue and red lines in Figure 6). However, it is found that
the model tends to simulate the KE axis slightly southward with a bias of ~0.5° (dash
lines in Figure 6). To make the SML depth more comparable, the SML depth from the
simulations is shifted northward by 0.5° (solid lines in Figure 6). After the shift, both
cases simulate similar meridional variabilities of the SML depth to the observations.

Overall, the SML depth from the simulations is deeper compared to the
observations. The bias between the simulation and observation is generally around tens
of meters. The restratification effect from MLI has been discussed due to a release of
the available potential energy (Boccaletti et al, 2007; Fox-Kemper et al., 2008). The
500 m resolution of the model is believed to resolve MLI fairly well. It can be inferred
that some restratifying processes are still missed in the model. Nevertheless, the SI500
tends to simulate a shallow SML compared to the KPP500 as the effect of SI is
parameterized, which may indicate biases in the KPP scheme or consequences of the
high frequency and diurnal cycle of forcing which are not used in the simulations here.
Quantitatively, the SML depth is shoaled by up 15 m due to SI restratification. Despite
that biases of the SML still exist compared to the observation, the SI parameterization
improves the SML simulation and makes it closer to the observations. The remaining
bias between the SI500 and the observation also suggests that our understanding of the
SML turbulence is still limited and further studies are needed in the future.

Figure 5 The SML depth (m) averaged over the simulation period (February) from (a) the KPP500 and (b) the SI500, and (c) their difference ($H_{SI500} - H_{KPP500}$).

3.3 Potential vorticity

SI occurs when PV is anticyclonic and trends to eliminate anticyclonic PV. PV is restored to a neutral state due to SI. As a result, the PV statistics are a direct indicator of the SI effect. Based on the model outputs, the PV ($q = (f \mathbf{k} + \nabla \times \mathbf{u}) \cdot \nabla b$) is calculated and analyzed. The temporally-averaged PV at 25 m depth is shown in Figure 7a and b. On average, the mean PV over the period is positive and the patterns from the
two cases, KPP500 and SI500 are similar. However, slight differences are still observed
between the KPP500 and SI500. Due to the SI restratification, the PV from the SI500
is slightly stronger than that from the KPP500. This enhancement is more pronounced
in the shallow-SML region, such as the core of the southwest cyclonic eddy and the
north KE region. The modulation on PV is confirmed by the differences between the
two cases (Figure 7c). Despite that alternating positive and negative changes are
observed over the domain, positive values are dominant over the negative ones.

Figure 7 PV ($\times 10^9$ s$^{-3}$) at 25 m depth averaged over the simulation period (February)
from (a) the KPP500 and (b) the SI500, and (c) their difference ($\text{PV}_{\text{SI500}} - \text{PV}_{\text{KPP500}}$).

In addition to the analysis of the mean PV, the percentiles of PV at 25 m depth also
show the modulation by SI (Figure 8a, b, c). Due to the SI parameterization, the 10th
and median PV percentiles during the simulation are both shifted to larger values
(Figure 8a, b). This shift is consistent with the SI effect, which can reduce negative PV
in the SI500 case. By contrast, the change of the 90th percentile due to SI is in contrast
to the naive expectation. The magnitude of the 90th percentile from the SI500 case is
weakened in comparison to the KPP500. To further investigate the SI effect on PV, the
probability density function (PDF) of the PV at 25 m during the simulation period is
calculated (Figure 8d). Compared to the KPP500, the PDF of PV from the SI500 is
generally shifted to more cyclonic values, indicating a reduction of anticyclonic PV
values due to the SI scheme. The modulation is applicable for PV in the range below
$8 \times 10^{-10}$ s$^{-3}$. For PV above $8 \times 10^{-10}$ s$^{-3}$, the SI500 has smaller PDFs compared to the
KPP500. This modulation needs further investigation on the PV budget to figure out
the mechanism, potentially involving feedbacks between the resolved flow and the parameterization effects which will not be discussed in this work.

Figure 8 Time evolution of the (a) 90th percentile, (b) median, (c) 10th percentile values of PV (s$^{-3}$) and (d) the probability density functions at 25 m depth.

4 Discussion and Conclusions

4.1 Discussion

As the energy source for SI, it should be noted that GSP is not necessarily increased in a model with SI parameterized or resolved compared to a model with no SI effects. The KPP closure may overestimate the viscosity at the front leading to a stronger GSP—indeed KPP has no awareness of the presence or absence of fronts. No attempt is made here to retune the KPP coefficients to be more in line with the predicted parameterized viscosity and diffusivity in the SI scheme (Equations 6 and 7). In this work and Bachman et al. (2017; cf. Figure 5 in their work), the magnitudes of the GSP are all weakened as SI is included compared to the cases with solely KPP scheme. In contrast, based on idealized fronts, Dong et al. (2021b) found strengthened GSP due to SI. According to the SI parameterization, the GSP is not only determined by the front intensity, but also by the surface buoyancy flux. So, whether SI can enhance GSP or not may be tightly related to the surface buoyancy flux.
Biases of GSP are observed between the SI500 and NBQ20. Except for the SI scheme performance capability, another reason may also account for the bias. Increasing the horizontal resolution of a model requires an increase in the vertical resolution. A requirement of $\frac{dx}{dz} \sim \frac{N}{f}$ for a high-resolution model to resolve small-scale geophysical processes is suggested by Menesguen et al. (2018). As stated in Section 2, the NBQ20 case has 150 vertical layers, which is 3 times as many as the KPP500 and SI500. The more vertical layers potentially simulate more fine structures and more accurate stratification in the SML, which potentially modulate the SI structures and further its effects. The sensitivity of the vertical layers needs further simulations.

Due to the computational limitation and the SI scheme, only 1 hour outputs under a positive buoyancy flux (i.e., ocean loses buoyancy) from the NBQ20 are analyzed. On one hand, the domain size is small and we have no idea how big the effect of boundary fluxes is. Due to the coarse resolution, the air-sea flux actually becomes almost constant over the whole NBQ20 domain. It is clear that the coarse-resolution air-sea flux does not match the ultra-high model, and it is still unknown if it is suitable to directly use the coarse-resolution to drive the ultra-high resolution model. It may be another factor that restricts us to get an accurate GSP. On the other hand, the 1 hour period is not enough to capture several life cycles of SI, and a longer simulation period will be helpful to better evaluate the SI in the NBQ20 case.

### 4.2 Conclusions

The performance of the SI scheme proposed by Bachman et al. (2017) in the CROCO has been evaluated by Dong et al. (2021b), which shows improvement of the model simulation capability in terms of several aspects. In this work, the scheme is further implemented in a realistic region, and the SI effects on GSP, SML depth, and PV in the KE region are investigated.

According to the simulation results, the KPP tends to overestimate the energy source of SI, GSP magnitude in the KE. By contrast, GSP is weakened by the SI scheme
in the SI500 compared to the KPP500 and is closer to the non-hydrostatic case NBQ20.

Although both KPP500 and SI500 cases simulate deeper SML depths compared to the
observations, the SI scheme improves the SML simulation by restratifying the SML by
up to 15 m, which tends toward the observations. Lastly, a primary role of SI is to
eliminate negative PV, which is reproduced by the SI scheme in the simulation. It is
found that the negative PV is reduced by the SI scheme in the SI500 compared to
KPP500, and the PV distribution is shifted toward positive PV by the SI scheme.

The primary goal of this work is to demonstrate the capability of the SI scheme in
a realistic region. The results in the KE here suggest that the SI scheme performs well
in the representation of the SI impacts, in that it approaches the behavior of the much
higher resolution non-hydrostatic model. According to Dong et al. (2021a), the western
boundary currents and the Antarctic Circumpolar Current are regions with active SI. As
SI is significant in SML dissipation (Dong et al., 2021a) and also the SML thickness,
this scheme is recommended in regional simulations to include SI effects and improve
the simulation results.

Author contributions

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Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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