Evaluating the Roles of Wind- and Buoyancy Flux-Induced Mixing on Phytoplankton Dynamics in the Northern and Central South China Sea

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Key Points:
- There is a stable SCM and a weak seasonality of surface CHL in the central SCS, while there is a shoaling SCM and a clear winter bloom in the north.
- In the northern SCS, buoyancy flux-induced mixing plays a dominant role in controlling winter mixing and seasonal phytoplankton dynamics.
- In the central SCS, both buoyancy flux- and wind-induced mixing control the winter mixing.

Abstract: Observations from two Bio-Argo floats deployed in the northern and central South China Sea (SCS) show distinct seasonal patterns of vertical chlorophyll distribution. There is a permanent subsurface chlorophyll maximum (SCM) located between 60 and 80 m throughout the year and a weak seasonality of surface chlorophyll in the central SCS. In the northern SCS, the SCM shoals to the upper mixed layer in winter and surface phytoplankton shows a clear winter bloom pattern. The mechanism driving the spatial and seasonal differences in phytoplankton dynamics in the euphotic zone remains unclear. Here a coupled physical-biological model is developed and applied to the northern and central SCS. With model and satellite data, we show that the contrasting patterns in chlorophyll are induced by the spatial variability in winter mixing dynamics. In the northern SCS, the buoyancy flux-induced mixing plays a dominant role in controlling the seasonal variability of vertical nutrient transport and phytoplankton production, which leads to the peak of surface chlorophyll and the significant shoaling of the SCM in winter. In the central SCS, the intensity of the buoyancy flux is reduced and both buoyancy flux- and wind-induced mixing control the winter mixing dynamics. However, the combination of these two mixing processes is weaker than in the northern SCS and the vertical nutrient transport is limited to the layer above the SCM, resulting in the reduced seasonality of surface chlorophyll and the relatively stable SCM all year round in the central SCS.

Plain Language Summary: Both satellite and Bio-Argo floats show a significant increase of surface chlorophyll concentration in winter in the northern South China Sea (SCS) but a very weak seasonal change in the central SCS. We used a coupled physical-biological model to systematically study the mechanism driving these spatial and seasonal differences. The model can reasonably simulate the different chlorophyll distribution patterns identified by observations. We found that the buoyancy flux-induced mixing plays a dominant role in controlling the seasonal change of chlorophyll in the northern SCS. In the central SCS, both buoyancy flux- and wind-induced mixing control the winter mixing dynamics. However, the combination of these two mixing dynamics is not as strong as that in the northern SCS and the vertical nutrient transport is only limited to the layer above the SCM, resulting in the reduced seasonality of surface chlorophyll and the relatively stable SCM all year round in the central SCS.

1. Introduction

Marine phytoplankton plays an important role in the global carbon cycle and climate change (Fasham, 2003). In the euphotic zone, phytoplankton growth is modulated by both physical (e.g., light, temperature, and mixing) and biological (e.g., nutrient availability, natural mortality, and zooplankton grazing) factors (Mann & Lazier, 1996). Phytoplankton standing stock in temperate regions often shows a prominent seasonal cycle, characterized by a strong increase of biomass in spring, when the surface mixed layer shoals to a depth shallower than a critical depth (Sverdrup, 1953). The variability of the mixed layer depth (MLD) is thus important for phytoplankton dynamics in these regions.
The South China Sea (SCS) is one of the world's largest marginal seas. It is characterized by diverse physical and biological dynamics at different spatial and temporal scales, which play an important role in biogeochemical cycles (Liu et al., 2002). Marginal seas have been suggested as a significant sink of atmospheric CO₂ in temperate regions (Thomas et al., 2004) and a source of CO₂ at low latitudes (Cai et al., 2006). The northern SCS, located in the subtropics and tropics, serves as a sink of atmospheric CO₂ in winter and a source in other seasons, whereas the central SCS is generally a source throughout the year. This phenomenon may be related to mixed layer variability and phytoplankton dynamics (Chai et al., 2009; Dai et al., 2013; Zhai et al., 2013).

The SCS is influenced by the East Asian Monsoon, with northeasterly winds in winter and southwesterly winds in summer (Gan et al., 2006; Xie et al., 2003). The wind is much stronger in winter than in summer and becomes weak and variable during the transitional periods in spring and fall. Previous studies have indicated that the MLD is modulated by both wind- and buoyancy flux-induced mixing in the northern SCS, whereas it is more affected by wind-induced mixing in the central SCS (Qu et al., 2007). In addition to the seasonal wind and surface buoyancy flux, other factors such as the Kuroshio intrusion and mesoscale eddies also make important contributions to the variability of the MLD in the SCS (Chu et al., 1999; Liu et al., 2001).

The seasonal change of upper ocean biology in response to the MLD variation has been investigated in previous studies. Using a monsoon-forced model, Liu et al. (2002) reproduced the main features of the surface chlorophyll (SChl) distribution in the SCS. In general, intermittent measurements in the SCS demonstrated noticeably lower surface production in summer and higher production in winter (Ning et al., 2004). Focusing on both the seasonal and spatial variability, studies using time series of remote sensing data have shown that there is a positive correlation between SChl concentration and wind speed over the entire SCS that is related to wind-induced mixing. There is also a robust negative correlation between SChl and sea surface temperature (SST) in the northern SCS that is related to the buoyancy flux-induced mixing (Shen et al., 2008; Tang et al., 2014). However, the evaluation of these two mechanisms is challenging, as satellites can only observe chlorophyll concentration in the surface layer. Thus, the response of phytoplankton to these mechanisms within the full euphotic zone remains unclear, especially with the existence of the subsurface chlorophyll maximum (SCM), which is a common feature in the SCS basin contributing significantly to phytoplankton stocks and production (Gong et al., 2014; Ning et al., 2004).

Using vertical profiles measured during the South East Asian Time-series Study (SEATS), Tseng et al. (2005) suggested that the winter enhancement of depth-integrated chlorophyll was induced by the combined effect of wind- and buoyancy flux-induced mixing in the northern SCS. Recently, using two Bio-Argo floats, one deployed in the northern and one in the central SCS, Zhang et al. (2016) were able to show distinct vertical chlorophyll structures with a strong winter surface phytoplankton bloom in the northern SCS and a very weak bloom in the central SCS. They attributed this spatial difference in phytoplankton to the difference in strength of wind mixing and surface cooling. However, quantification of the relative roles of these two mechanisms in terms of phytoplankton dynamics has not been conducted.

The ecosystem in the SCS basin also experiences interannual variation, particularly during El Niño–Southern Oscillation (ENSO) events. Generally, on the basinwide scale, there is a significant negative correlation over the last 20 years between SChl and the ENSO index (Liu et al., 2013; Palacz et al., 2011; Tang et al., 2011). The weakened wind mixing and strengthened vertical stratification during the 1997–1998 El Niño event produced anomalously low SChl in both the northern and central SCS (Tseng et al., 2009; Zhao & Tang, 2007).

To elucidate both the spatial and temporal variability of phytoplankton dynamics in the SCS, a comparative study was conducted in this work by combining observations and numerical models. We constructed 2 one-dimensional (1D) physical-biological models for the northern and central SCS, near the general locations of the two Bio-Argo floats (Figure 1a). These coupled models were forced by realistic sea surface data and compared with satellite and in situ observations. The satellite data and model results were used to analyze the driving mechanism for the different chlorophyll distribution patterns in the northern and central SCS by investigating phytoplankton dynamics in response to wind- and buoyancy flux-induced mixing over the euphotic zone.
2. Model and Data

2.1. One-Dimensional Coupled Model

Two 1D physical-biological models were constructed with one located in the northern SCS (116°E, 18°N; SEATS) and the other in the central SCS (114°E, 13°N; SCS2; Figure 1a). The two sites selected for modeling are located in the SCS deep basin, far from the continental shelf, with a bottom depth of >3,800 m. In these regions, horizontal advection and diffusion are thought to be less important for chlorophyll dynamics than vertical mixing (Du et al., 2017; Li et al., 2015; Lu et al., 2015). The physical model is based on the Princeton Ocean Model (Blumberg & Mellor, 1987). The Mellor-Yamada Level 2.5 turbulence closure scheme is used in the model.

The biological model is based on the ecosystem model developed for the SCS by Liu et al. (2002, 2007). It has four state variables: phytoplankton ($P$), zooplankton ($Z$), detritus ($D$), and dissolved inorganic nitrogen ($N$). Chlorophyll is derived from phytoplankton biomass based on a variable chlorophyll-to-phytoplankton ratio. The growth of phytoplankton is constrained by nutrient and light availability. The loss terms for phytoplankton include mortality, aggregation, and zooplankton grazing. Blooms of phytoplankton in this type of model can originate from increased light and nutrient levels or reduced grazing. This simplification is appropriate in the SCS basin because phytoplankton in the basin is generally dominated by picoplankton and its growth is mainly limited by nitrate (Chen et al., 2004; Chen et al., 2009). Other factors that may be important but are not incorporated in the model include phytoplankton composition and multiple nutrient cycling dynamics. Detailed descriptions of the biological model and parameters are given in Appendices B and C.

Figure 1. Model stations, trajectories of the Bio-Argo floats, and time-series of mixed layer depth (MLD) and subsurface chlorophyll maximum (SCM) depth observed by the floats. (a) Bathymetry of the South China Sea (SCS) with the trajectories of the two Bio-Argo floats. The yellow (black) curve denotes the trajectory of float 0347 (0348), with the red (pink) square indicating the starting point of the float. The red (green) star indicates the model station at SEATS (SCS2). (b) Time series of MLD observed by float 0347 (blue line) and float 0348 (red line). (c) As in (b), but for SCM depth.

Figure 2. Climatological monthly mean fields at SEATS (green curves) and SCS2 (red curves): (a) remotely sensed surface chlorophyll concentration (SChl) from Ocean Color Climate Change Initiative data; (b) wind speed from Cross-Calibrated Multi-Platform data; and (c) sea surface temperature (SST) from advanced very high resolution data. The monthly mean fields are calculated using daily data from January 2010 to December 2015.

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The models were initialized using data from the World Ocean Atlas 2013. The models use six-hourly surface forcing fields, including surface wind components at 10 m, air temperature at 2 m, relative humidity at 2 m, sea level pressure, total cloud coverage, and net shortwave radiation obtained from National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis data (Kalnay et al., 1996). For the control runs, both models were integrated for 26 years from January 1992 to December 2017.

Using model outputs, we calculate the euphotic zone depth as the depth at which photosynthetically active radiation (PAR) is reduced to 1% of its surface value (Kirk, 1994). The MLD is calculated as the depth at which the density is equal to the sea surface density plus an increment in density equivalent to 0.8 °C (Kara et al., 2000). The nutricline depth is calculated as the depth at which the nitrate concentration is 0.1 mmol N/m^3. The modeled vertical kinematic viscosity ($K_M$) is used to indicate the strength of vertical mixing. The definition of $K_M$ is provided in Appendix A.

To evaluate the model performance, multiple observations conducted at the two modeling locations were collected. At SEATS, in situ chlorophyll profiles from four cruises covering four seasons were taken from Chen (2004; c673 in winter, c629 in spring, c644 in summer, and c585 in fall). During the South China Sea Monsoon Experiment, two moorings were set up near the SEATS and SCS2 locations (Liu et al., 2001). One-year continuous temperature measurements from these two moorings were obtained during April 1998 to April 1999. These data along with satellite-derived SST from the Advanced Very High Resolution Radiometer (AVHRR) Pathfinder Version 5.2 with a spatial resolution of 4 km were used for comparison with modeled temperature distributions. All the model data comparisons were performed for the observation periods.

2.2. Floats and Satellite Data

Continuous measurements from two Bio-Argo floats equipped with a SBE 41CP CTD and a WET Labs ECO-MCOMS fluorometer were obtained from Zhang et al. (2016). These two floats (0347 and 0348, Figure 1a)
were deployed in the northern and central SCS basin on 27 June and 11 July 2014, respectively. Vertical profiles of temperature, salinity, and chlorophyll were measured every 3 days for approximately 1 year and were used to determine the various spatiotemporal distribution patterns of chlorophyll and the mixed layer between the northern and central SCS.

A statistical analysis of the relationships among wind, buoyancy flux, and phytoplankton, was performed using remote sensing products at daily resolutions for SEATS and SCS2. The SChl was derived from the Ocean Color Climate Change Initiative (OC-CCI) data set with a horizontal resolution of 4 km. The daily Cross-Calibrated Multi-Platform Winds data set with a resolution of 0.25° was used to calculate surface wind speeds (Atlas et al., 2011). Daily SST derived from the AVHRR Pathfinder Version 5.2 was used to quantify the response of surface ocean to heating- and cooling-induced buoyancy changes, as the seasonal variation of SST is predominantly controlled by surface heat flux in the SCS (Qu, 2001). Previous studies have found significant positive correlations between the seasonal change of SST and that of air temperature, indicating effective air-sea heat exchange in the SCS (e.g., Pan et al., 2015; Tseng et al., 2005). Here we did not use global heat flux data sets for analysis, because most of the global heat flux data sets have spatial resolutions that are too coarse for comparison with high-resolution chlorophyll data. Furthermore, large biases in the SCS were found between in situ observations and these data sets (Wang et al., 2017).

3. Results

3.1. Observed Differences Between the Northern and Central SCS

The Bio-Argo float (0347) deployed in the northern SCS moved southwestward during the 1-year period (Figure 1a). The MLD along the float trajectory showed a clear seasonal pattern (Figure 1b). The
shallowest MLD around 10 m occurred in spring (MAM) and the deepest MLD reaching ~80 m was found in winter (DJF). The depth of the SCM was shallowest in winter (Figure 1c). The seasonal variability of the MLD measured by the other float (0348) in the central SCS was relatively weak (Figure 1b). Although the MLD in winter was deeper than that in other seasons, it was generally shallower than 60 m. Consequently, the depth of the SCM was relatively stable in this region, typically ranging between 60 and 80 m during the year (Figure 1c).

Differences in seasonal patterns between the northern and central SCS were also revealed by remotely sensed SChl (Figure 2a). In winter, SChl at SEATS is much higher than at SCS2, while their magnitudes are comparable in other seasons. The seasonal patterns of wind speed at these two locations are quite similar, especially from January to June (Figure 2b), which does not appear to explain the large difference of SChl in winter. We then examined the seasonal pattern of SST that is associated with buoyancy flux-induced mixing in the upper ocean. The SSTs at both SEATS and SCS2 are low in winter and high in summer, and the difference between these two locations is considerably larger in winter than in other seasons, which is consistent with the pattern of SChl (Figure 2c and Table 1).

Scatterplots of daily wind speed, SST, and SChl also show clear differences between SEATS and SCS2 (Figure 3). At SEATS, high SChl concentrations (>0.25 mg/m³) are present in winter during the period with low SST (<25 °C). Overall, the absolute value of the correlation coefficient between SChl and SST (R = −0.8) is larger than that between SChl and wind speed (R = 0.3) at SEATS. Compared with that at SEATS, the SST in winter is higher at SCS2. The absolute value of the correlation coefficient between SChl and SST is lower (−0.6) and that between SChl and wind speed is higher (0.4) at SCS2 than those at SEATS.

3.2. Modeled Distributions in the Northern and Central SCS

The comparison between modeled and remotely sensed SChl is similar to that shown in Geng et al. (2012), who used the same model configuration at SEATS. Here we also show comparisons of model results and in

Figure 5. Comparisons of modeled vertical profiles of chlorophyll (blue solid curves) and field measurements (red symbols) at the SEATS station in (a) winter, (b) spring, (c) summer, and (d) fall. The unit for chlorophyll is mg/m³.
situ observations. Modeled SSTs were quantitatively compared with observations from moorings and AVHRR. During 1998–1999, the model reproduced the annual pattern of SST reasonably well compared with both mooring and AVHRR observations at SEATS (Figure 4a). As shown in the Taylor diagram, the correlation coefficients between modeled SST and mooring and AVHRR observations are 0.94 and 0.88, and the root mean square differences (RMSDs) between them are 0.35 and 0.49, respectively (Figure 4c). At SCS2, modeled SST compares very well with mooring observations with a large correlation coefficient of 0.97 and a small RMSD of 0.26 (Figures 4b and 4d). The correlation coefficient and RMSD between model and AVHRR are 0.83 and 0.74, respectively. Overall, the model performs better against mooring observations than AVHRR data at both locations. The discrepancy between model results and AVHRR is probably due to the fact that the AVHRR data are not a regional product and have a relatively low spatial resolution (4 km) that cannot resolve small-scale processes.

Modeled vertical profiles of chlorophyll are compared with observations (Figure 5). The in situ measured vertical distributions of chlorophyll at SEATS generally agree with model results. The model reproduced the depth and magnitude of the SCM. However, for some measurements, the model did not simulate the distribution pattern very well in the euphotic zone, which is probably due to the simple structure of the

Figure 6. Modeled time series (January 1997 to December 2000) of vertical profiles at (a–c) SEATS and (d–f) SCS2. (a, d) Vertical kinematic viscosity ($K_M$), (b, e) temperature, and (c, f) chlorophyll. The black lines in panels (a), (b), (d), and (e) represent the mixed layer depth (m).
biological model. In addition, mesoscale eddies may also cross the study region sporadically and affect vertical nutrient transport; these eddies are not represented in the 1D model.

At SEATS, the modeled chlorophyll concentration in winter is high in the upper part of the water column mainly due to strong vertical mixing (Figures 6a and 6c and Table 1). At SCS2, the magnitude of winter chlorophyll concentration in the upper water column is greatly reduced associated with weakened vertical mixing (Figures 6d and 6f and Table 1). In other seasons, the SChl is clearly reduced (<0.1 mg/m^3) at both SEATS and SCS2, which is consistent with the satellite data. There is a significant SCM at SEATS near the bottom of the euphotic zone below the MLD during most of the year. The SCM is strong in spring and summer but very weak in winter. The maximum chlorophyll concentration in the SCM layer can reach 0.7 mg/m^3 in spring, which is even higher than the SChl in winter. Unlike the surface values, the magnitude of the SCM is reduced during winter. The winter MLD is shallower at SCS2 than that at SEATS (Table 1). The dominant feature of the vertical chlorophyll is the presence of a SCM all year round that stays relatively stable around 70- to 80-m depth (Figure 6f). These modeled chlorophyll distribution patterns in the northern and central SCS are consistent with observations from the Bio-Argo floats (Figure 1).

3.3. Factors Affecting Phytoplankton Dynamics in the Northern and Central SCS

The modeled euphotic zone depth and nutricline depth were compared with the MLD to illustrate the different driving mechanisms for the winter phytoplankton bloom in the northern and central SCS.
It is clear that the euphotic zone depth is quite stable, at about 90 m, which is deeper than the MLD and nutricline depth (Table 1), suggesting that light is not a limiting factor for phytoplankton growth at either SEATS or SCS2. The surface phytoplankton bloom at SEATS therefore starts when the MLD is deep enough to reach the nutricline and bring subsurface nutrients to the surface layer.

The responses of SChl and depth-integrated chlorophyll over the euphotic zone (IChl) to winter mixing are not the same in the northern and central SCS (Figures 7b and 7d). At SEATS, the winter peak of SChl can reach about 6 times the magnitude in fall, and it decreases rapidly to reach the minimum in summer. As a comparison, the IChl also peaks in winter, but unlike that of SChl, the winter peak of IChl is only ~1.7 times higher than that in other seasons. After the winter peak, the IChl appears to decrease gradually with time and reaches its minimum before the next winter. At SCS2, where the MLD is generally located above the nutricline, the SChl and IChl in winter are 2 and 1.6 times higher than those in fall, respectively.

Generally, the depth-integrated chlorophyll over the SCM layer (IChlSCM) accounts for the majority (~80.0%) of the IChl (Figure 8). Climatologically, during winter, the contribution of SCM (the ratio defined as IChlSCM/IChl) sharply decreases to about 55.0% at SEATS and about 68.0% at SCS2 (Table 1), but it remained high (70.0% at SEATS and 74% at SCS2) during the winter of 1997–1998. These results indicate that winter mixing not only increases SChl but also decreases the IChlSCM. Climatic forcing can change SChl significantly, whereas the existence of the SCM appears to buffer the biological response to climate variability over the entire euphotic zone.

Note that we only show 4-year modeling results here covering a typical ENSO cycle (1997–1999) and a normal year (2000). The analysis was performed for the whole modeling period. As seasonal variability is the dominant signal in the study region, the primary patterns of phytoplankton dynamics remain unchanged when analyzing the results from different years.

### 3.4. Budget Analysis in the Northern and Central SCS

To further illustrate the different contributions of physical and biological processes to the variation of phytoplankton dynamics in the northern and central SCS, we conducted a nitrogen budget analysis of the phytoplankton governing equation based on model outputs. The terms responsible for the change of nitrogen contained in the phytoplankton biomass are defined as follows:

\[
\frac{\partial P}{\partial t} = \frac{\partial}{\partial z} \left( K_v \frac{\partial P}{\partial z} \right) + \frac{\partial SMSP}{\partial z} \tag{1a}
\]

\[
SMSP = \mu_P P \left[ 1 - \exp \left( \frac{\alpha R_{1E}}{\mu_P} \right) \right] \frac{R}{R_{1H} (N + k_N)} - \eta P - \omega P^2 - \mu_z [1 - \exp(-\lambda P)] \tag{1b}
\]

where SMSP is the source-minus-sink term for phytoplankton. Definitions and values of the parameters used here are listed in Table C1. Budget terms in the equation are defined as follows: (1) the tendency term representing the rate of change of phytoplankton biomass; (2) the vertical diffusion term denoting the effects of vertical mixing; and (3) the source-minus-sink term (SMSP) including all biological processes related to phytoplankton variation. This last term is made up of (4) the phytoplankton growth representing phytoplankton primary production utilizing nutrients and light to photosynthesize; (5) the remineralization term; (6) the...
phytoplankton aggregation term; and (7) the zooplankton grazing term. All terms are given in units of mmol N·m⁻³·day⁻¹. Here the budget analysis assumes that horizontal diffusion and advection are small relative to other terms (e.g., Li et al., 2015; Lu et al., 2015).

At SEATS, model results suggest that the surface phytoplankton growth term approaches its maximum in winter (~100 × 10⁻³ mmol N·m⁻³·day⁻¹) but is low in other seasons due to nutrient limitation, with a minimum value of ~5.07 × 10⁻³ mmol N·m⁻³·day⁻¹ in summer (Figure 9a). When phytoplankton blooms in winter, all the sink terms reach their maximum (negative) values. The surface SMS_p term representing the net biological contribution reaches its maximum in winter with a mean value of ~43 × 10⁻³ mmol N·m⁻³·day⁻¹. In contrast, this term has a negative value of ~−1.3 × 10⁻³ mmol N·m⁻³·day⁻¹ in spring, indicating the decay of phytoplankton blooms. In other seasons, the SMS_p term is generally low and remains positive.

The maximum diffusion term of phytoplankton is found in winter, with a maximum value of ~−21.0 × 10⁻³ mmol N·m⁻³·day⁻¹ at the time of maximum phytoplankton biomass (Figure 9b). In other seasons, the vertical diffusion term is also negative but smaller in magnitude. The rate of phytoplankton change (tendency term) including the combined effects of physical and biological contributions is shown in Figure 9c. It reaches a maximum in winter, consistent with the phytoplankton bloom. However, this term is very low in the rest of the year with a minimum of ~−1.6 × 10⁻³ mmol N·m⁻³·day⁻¹ in spring. Comparison of

Figure 9. Modeled time series (January 1997 to December 2000) at SEATS of the terms in equation (1) for surface phytoplankton: (a) phytoplankton growth, phytoplankton remineralization, zooplankton grazing, and phytoplankton aggregation; (b) total phytoplankton source minus sink (SMS_p) and vertical diffusion; (c) phytoplankton tendency; (d) surface phytoplankton concentration (mmol N/m³). All budget terms are in units of mmol N·m⁻³·day⁻¹.
the variance of the $SMSP$ and vertical diffusion terms indicates that the direct dilution effect on phytoplankton cells appears to be small relative to other processes. There is a balance between phytoplankton growth and loss terms. The dominant loss terms are the remineralization and zooplankton grazing terms.

Vertical distributions of budget terms at SEATS are shown in Figure 10. During winter, the growth term is very low in the SCM layer, in contrast to that in the surface layer. Zooplankton grazing appears to be the dominant sink for phytoplankton, especially in the SCM layer, followed by aggregation and remineralization. High values of $SMSP$ occur in the surface during winter, consistent with the more intense phytoplankton bloom at the surface than below the mixed layer. Negative $SMSP$ values are present in the SCM layer in winter, probably related to the reduced nutrient concentration caused by mixing. The vertical diffusion term is generally negative in the SCM layer but with a relatively small magnitude (Figure 10f). Right above the SCM layer, the vertical diffusion term shows large positive values in winter corresponding to the negative values in the surface layer, suggesting accumulation of phytoplankton cells modulated by physical processes.

At the SCS2 station, the growth term in the SCM layer remains high in winter, unlike the case at SEATS (Figure 11). The $SMSP$ term does not show clear negative values above the SCM layer in winter. The magnitude of vertical diffusion is greatly reduced, and the winter mixing does not appear to affect the SCM layer as strongly as at SEATS (Figure 11f).

![Figure 10. Vertical distributions of budget terms in equation (1) at SEATS: (a) phytoplankton growth; (b) phytoplankton remineralization; (c) zooplankton grazing on phytoplankton; (d) phytoplankton aggregation; (e) phytoplankton source minus sink ($SMSP$); and (f) vertical diffusion. All the budget terms are in units of mmol N m$^{-3}$ day$^{-1}$.](image-url)
4. Discussion

4.1. Mechanisms Leading to Different Phytoplankton Responses in the Northern and Central SCS

In winter, the wind is generally stronger and the SST is lower at SEATS than at SCS2 (Table 1). These two factors are likely to drive a significant change in the MLD (~68.3 m at SEATS and ~60.8 m at SCS2). In seasons other than winter, the nutricline depth stays relatively stable, around ~58 m at both SEATS and SCS2. In winter, the MLD at SEATS becomes deep enough to reach the nutricline and bring nutrients to the surface layer. However, it is not the case at SCS2 due to the shallow MLD and weak mixing (Figures 7a and 7c and Table 1). The deeper MLD at SEATS in winter can consequently stimulate stronger phytoplankton production in the upper layer and suppress production in the SCM layer (Figure 10). In comparison, the intensity of winter mixing is reduced at SCS2 (Table 1). Although surface production increases, the mixing is unable to significantly affect production in the SCM layer, which results in a relatively stable SCM over the year (Figures 6f and 11).

To further evaluate the roles of wind- and buoyancy flux-induced mixing in triggering phytoplankton blooms in the northern and central SCS, sensitivity experiments were carried out. One experiment (Exp1) was forced by climatological annual mean wind and real-time (six-hourly) heat fluxes and another experiment (Exp2) was forced by real-time (six-hourly) wind and climatological annual mean heat flux. As Exp1 is the experiment without variability from wind-induced mixing and Exp2 is the one

| Table 2 |
| Seasonal Amplitudes (Maximum − Minimum) of MLD, SChl, and IChl for the Three Model Runs (Control, Exp1 and Exp2) at SEATS and SCS2 |
| Parameter | SEATS | Control | Exp1 | Exp2 | Control | Exp1 | Exp2 |
| MLD (m) | 47.2 | 46.3 | 5.8 | 34.5 | 32.9 | 5.3 |
| SChl (mg/m³) | 0.16 | 0.06 | 0.01 | 0.07 | 0.03 | 0.03 |
| IChl (mg/m²) | 6.00 | 5.99 | 3.19 | 4.20 | 3.79 | 3.18 |

Note. MLD = mixed layer depth; SChl = surface chlorophyll concentration; IChl = depth-integrated chlorophyll concentration over the euphotic zone.
without buoyancy flux-induced mixing, we can estimate the relative contributions of the two mechanisms from these runs. In Exp1, the model produced clear seasonal changes of water column stratification and MLD dynamics with a seasonal amplitude of MLD of 46.3 m, which is very close to that simulated in the control run (47.2 m) at SEATS (Table 2). However, the modeled MLD (with a deepest monthly mean of ~72.0 m) was still shallower than that of the control run (with a deepest monthly mean of ~81.4 m; Figures 7a and 12). Weakened vertical mixing leads to a warmer upper water column. Without forcing by seasonally varying heat flux in Exp2, the seasonal variability of water column stratification and MLD was greatly reduced with a seasonal amplitude of MLD of only ~5.8 m (Figure 13 and Table 2). This indicates that buoyancy flux-induced mixing plays a more important role in maintaining the seasonal evolution of water column stratification and MLD dynamics. Similar results were found at SCS2 (Figures 14 and 15). The modeled seasonal amplitudes of MLD are 32.9 and 5.3 m from the Exp1 and Exp2 runs, respectively, compared with the 34.5 m from the control run at SCS2 (Table 2).

Without both forcing factors, the winter phytoplankton bloom in the whole water column is significantly weakened, especially in Exp2 where winter nutrient flux to the surface is reduced (Figures 12–15). In

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**Figure 12.** Modeled time series of temperature (°C), nitrate (mmol N/m$^3$), and chlorophyll concentration (mg/m$^3$) vertical profiles at SEATS from the sensitivity experiment Exp1 that is forced with climatological annual mean wind and real-time heat flux. Black lines indicate mixed layer depth (m).
January 1999 at SEATS, mean SChl decreased from 0.24 to 0.17 mg/m³ in Exp1, and 0.10 mg/m³ in Exp2, equivalent to about 71% and 42% of the control run values, respectively (Figure 16a). The magnitudes of SCM are also different. Unlike SChl, weakened mixing in Exp1 and Exp2 can result in increased SCM in January and February. The SCM in Exp1 was slightly reduced in other months compared with the control run (Figure 16b). The increase in SCM in January and February is mainly caused by the reduced upward nutrient flux leaving more nutrients below the mixed layer. The IChl shows distinct differences in winter, when reduced vertical mixing in Exp1 and Exp2 leads to low phytoplankton concentration (Figure 16c). In January 1999, the mean IChl was about 23 mg/m² in the control run, reduced to about 22 mg/m² in Exp1, about 94% of the control run, and to about 18 mg/m² in Exp2, about 79% of the control run, indicating the greater importance of buoyancy flux-induced mixing, which is consistent with the MLD dynamics (Figures 12 and 13). In other months, the IChl values from the three runs were similar.

At SCS2, SChls were significantly reduced in both Exp1 and Exp2, especially in winter (Figure 17a). The difference of SChl between Exp1 and Exp2 was much smaller at SCS2 than at SEATS. For example, in January 1999, mean SChl was 0.08 mg/m³ in Exp1 and 0.07 mg/m³ in Exp2, about 67% and 58% of the value in the control run, respectively. The SCM in Exp1 was stronger than in Exp2 and very close to the control run.
throughout the year (Figure 17b). A similar pattern was found for the IChl distribution, except in January when the IChl in the control run was higher than in Exp1 and Exp2 (Figure 17c). In January 1999, the mean IChl was about 17 mg/m² in the control run, about 15 mg/m² in Exp1 and Exp2, suggesting comparable contributions from wind- and buoyancy flux-induced mixing.

The above results suggest different contributions of wind- and buoyancy flux-induced mixing to phytoplankton dynamics in the northern and central SCS. At SEATS, the seasonal amplitude of SChl in Exp2 was 6.3% of the control run value, which may suggest a 93.7% contribution from the buoyancy flux-induced mixing (Table 2). At SCS2, the contribution from buoyancy flux-induced mixing was lower and that from wind-induced mixing was higher, resulting in comparable values that were both around 57.0% for SChl. Note that the seasonal change of SChl does not linearly respond to changes in MLD. For example, at SCS2 the difference of MLD amplitude between the control run and Exp2 is much larger than between the control run and Exp1, whereas they have similar differences in SChl amplitude (Table 2). For the whole euphotic zone, buoyancy flux-induced mixing accounts for 46.8% of the seasonal cycle of IChl at SEATS, while the contribution from wind-induced mixing was extremely low. At SCS2, the contribution from buoyancy flux reduced to

Figure 14. Modeled time series of temperature (°C), nitrate (mmol N/m³), and chlorophyll concentration (mg/m³) vertical profiles at SCS2 from sensitivity experiment Exp1, which is forced with climatological annual mean wind and real-time heat flux. Black lines indicate the mixed layer depth (m).
24.3% and that from wind increased to 9.8%. It indicates that the buoyancy flux-induced mixing plays a more important role in the seasonal variation of surface phytoplankton in the northern SCS than in the central SCS. For the depth-integrated chlorophyll, the contributions from buoyancy flux-induced mixing were significantly reduced in both the northern and central SCS compared with those for SChl. This study reveals different responses to winter mixing of SChl and depth-integrated chlorophyll. SChl is more sensitive than depth-integrated chlorophyll to changes in winter mixing.

4.2. Uncertainties in the Model

In this study, we used two 1D models to simulate phytoplankton dynamics in the northern and central SCS. One advantage of employing a 1D model is the better representation of the upper mixed layer by using more vertical model layers than the coarse resolution normally used in a three-dimensional model to save computational cost (Gan et al., 2006; Liu et al., 2002). However, the 1D model is unable to resolve horizontal advection processes that are often important in transporting nutrients from shelf and coastal regions. This influence is relatively small at the two stations chosen in this study that are located in the deep basin of the SCS away from the major rivers, coastal upwelling, and currents (Hu et al., 2000; Jing et al., 2015; Shu et al., 2018; Wong et al., 2007). Biogeochemical processes in these locations are thought to be largely

Figure 15. As in Figure 14, but for sensitivity experiment Exp2 at SCS2, which is forced with real-time wind and climatological annual mean heat flux.
driven by local processes, especially on a seasonal time scale (e.g., Du et al., 2017; Li et al., 2015; Tseng et al., 2009). Previous 1D modeling studies conducted for the SCS basin have been able to simulate the vertical structures of biogeochemical variables reasonably well (e.g., Geng et al., 2012; Gong et al., 2014; Li et al., 2015).

Mesoscale eddies are ubiquitous in the SCS (Wang et al., 2003; Xiu et al., 2010), and they can affect phytoplankton growth through vertical nutrient injection and horizontal advection (Guo et al., 2017; Ning et al., 2004). While mesoscale eddies are important sources of vertical nutrient fluxes, they generally occur sporadically in the SCS and do not show a clear seasonal pattern, especially in the deep basin (Xiu et al., 2010). Results from Zhang et al. (2016) showed that the Argo float released in the northern SCS only encountered a weak cyclonic eddy once throughout the whole winter. Without considering these mesoscale and sporadic processes, the 1D model is simplified but is potentially suitable to investigate the representative seasonal cycle of phytoplankton dynamics in the SCS.

Climatological mean forcings were used to drive the model in the two sensitivity experiments (i.e., Exp1 and Exp2). Note that chlorophyll concentrations will be different when using mean winter or summer conditions due to the complex nonlinear influence on vertical nutrient flux. To reduce uncertainties, we used the annual amplitude as a measure for evaluating the relative contributions from different physical processes. We separated buoyancy flux-induced mixing and wind-induced mixing as independent factors in the two experiments, but realistically, they also covary to a certain degree.

The modeling results are consistent with previous studies. Tseng et al. (2005) performed a case study for the northern SCS using a numerical model and found a dominant role for cold convective overturn in determining winter mixing. As shown in Figure 13, the simulated MLD is located at a relatively stable depth (45–61 m) when the model is forced with real-time wind and climatological annual mean heat fluxes.
However, this does not necessarily mean that wind does not contribute to the vertical stratification of the water column, just that the seasonal variation of wind is not strong enough to generate the observed seasonal variation of MLD under these water column stratification conditions. This is consistent with a recent study by Tai et al. (2017), who found that the intra-annual variability of MLD was mainly affected by convective overturn induced by surface cooling in winter. Moreover, the effect of wind on mixed layer dynamics can vary with background conditions. To explore these effects, two additional sensitivity experiments—that is, Exp1_summer and Exp1_winter—forced by summer and winter mean wind,

Figure 17. As in Figure 16, but for the distributions at SCS2.

Figure 18. Modeled time series of mixed layer depth (MLD) from the control run (black curves), Exp1_winter (blue curves), and Exp1_summer (red curves) for January 1997 to December 2000 at the SEATS station.
respectively, and both with real-time surface heat fluxes were conducted. Results from these two experiments were compared with a control run that was forced by real-time wind and heat flux (Figure 18). This comparison demonstrates that the MLD of summer can increase by >10 m when surface wind is increased to its winter mean value. In comparison, the MLD of winter can decrease by up to 20 m when surface wind is reduced to its summer mean value.

5. Conclusions

The contrasting seasonal patterns of chlorophyll distribution between the northern and central SCS were observed by two Bio-Argo floats. The underlying mechanism causing this difference was investigated by analysis of phytoplankton dynamics in the euphotic zone. Results show that both wind- and buoyancy flux-induced mixing are responsible for the enhancement of vertical mixing and the surface phytoplankton bloom in winter. In the northern SCS, buoyancy flux-induced mixing plays a dominant role in controlling the seasonal variability of vertical nutrient transport and phytoplankton production. Model sensitivity experiments suggest that >90% of the seasonal variability in SChl is caused by the variation of surface buoyancy flux. In the central SCS, the intensity of buoyancy flux-induced mixing is lower, and both wind and buoyancy flux control the winter mixing dynamics with comparable contributions (~57%) to the seasonal cycle of SChl. However, the combination of these two mixing processes is not as strong as in the northern SCS (~8-m difference in MLD; 0.03-m²/s difference in $K_M$), and the vertical nutrient transport is limited to the layer above the SCM. These results help explain the significant shoaling of the winter SCM in the northern SCS and the relatively stable SCM throughout the year in the central SCS. The findings presented here contribute to the quantitative understanding of phytoplankton dynamics in the SCS.

Appendix A: Definitions of Vertical Kinematic Viscosity ($K_M$) and Vertical Diffusivity ($K_V$)

In the model, the vertical kinematic viscosity ($K_M$) and vertical diffusivity ($K_V$) are calculated using the Mellor-Yamada 2.5 turbulent model (Mellor and Yamada, 1992) and are defined as

$$K_M = q l S_M \text{ and}$$

$$K_V = q l S_V,$$

where the coefficients $S_M$ and $S_V$ are functions of the Richardson number ($G_H$) given by

$$S_V[1-(3A_2 B_2 + 18A_1 A_2)G_H] = A_2[1-6A_1/B_1] \text{ and}$$

$$S_M[1-9A_1 A_2 G_H]-S_V[(18A_1^2 + 9A_1 A_2)G_H] = A_1[1-3C_1 - 6A_1/B_1],$$

where

$$G_H = \frac{q^2 g}{l^2 \rho_0} \left[ \frac{\partial \rho}{\partial z} - \frac{1}{c_s^2} \frac{\partial^2 \rho}{\partial z^2} \right].$$

Here $q$ is the turbulence kinetic energy, $l$ is the turbulence length scale, $(A_1, A_2, B_1, B_2, C_1) = (0.92, 16.6, 0.74, 10.1, 0.08)$ are constants, $g$ is the acceleration due to gravity, $\rho_0$ is the reference density, $\frac{\partial \rho}{\partial z}$ is the vertical density gradient, and $c_s$ is the speed of sound.

Appendix B: Description of the Biological Model

The biological model that was developed by Liu et al. (2002, 2007) is used in this study. The state variables are expressed in terms of nitrogen concentration (mmol N/m³).

The general equation for the biological state variable ($C$) in the 1D model is
\[ \frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left( K_v \frac{\partial C}{\partial z} \right) + SMS_C, \]  

where the state variable \( C \) represents the four state variables \( P, Z, D, \) and \( N; K_v \) is the vertical diffusion coefficient; and \( SMS_C \) is the biological process-related source and sink terms calculated in the biological model. The \( SMS_C \) for each state variable is calculated as follows:

\[ SMS_P = \mu_P \left[ 1 - \exp \left( - \frac{\alpha R_{\text{fl}} E}{\mu_P} \right) \right] \frac{R}{R_{\text{fl}}} \left( \frac{N}{N + \kappa N} \right) - \eta P - \omega P^2 \]  

\[ SMS_Z = \left( 1 - \gamma \right) \mu_Z Z \left[ 1 - \exp(\lambda P) \right] - n_1 Z - n_2 Z^2, \]  

\[ SMS_D = \omega P^2 + \gamma \mu_Z Z \left[ 1 - \exp(\lambda P) \right] + \varepsilon_1 n_1 Z + \varepsilon_2 n_2 Z^2 \]  

\[ -\delta D - W_{\text{sink}} \frac{\partial D}{\partial z}, \]  

and

\[ SMS_N = \eta P + (1 - \varepsilon_1) n_1 Z + (1 - \varepsilon_2) n_2 Z^2 + \delta D \]  

\[ -\mu_P \left[ 1 - \exp \left( - \frac{\alpha R_{\text{fl}} E}{\mu_P} \right) \right] \frac{R}{R_{\text{fl}}} \left( \frac{N}{N + \kappa N} \right). \]  

The four terms on the right-hand side of equation (B2) represent phytoplankton growth, mortality, aggregation, and zooplankton grazing. The \( SMS_C \) term for zooplankton includes an assimilation term and two sink terms for zooplankton mortality (equation (B3)). The \( SMS_D \) term for detritus contains source terms for phytoplankton aggregation, zooplankton grazing loss, and zooplankton mortality, and two sink terms for detritus remineralization and vertical sinking (equation (B4)). The vertical sinking velocity for detritus \( (W_{\text{sink}}) \) of 10 m/day is used in the model. As the detritus reaches the seafloor, it is assumed that ~86% is immediately remineralized into dissolved inorganic nitrogen and released to the water column (Liu et al., 2007). The \( SMS_N \) term for dissolved inorganic nitrogen includes all remineralization terms and removal by phytoplankton uptake (equation (B5)).

The maximum specific growth rates of phytoplankton \( (\mu_P) \) and zooplankton \( (\mu_Z) \) depend on temperature \( (T \) in °C) as follows:

\[ \mu_P = \mu_{P_{20}} \left( \frac{T - 20}{10} \right)^{10} \]  

\[ \mu_Z = \mu_{Z_{20}} \left( \frac{T - 20}{10} \right)^{10}, \]  

where \( \mu_{P_{20}} \) and \( \mu_{Z_{20}} \) are the maximum specific growth rates of phytoplankton and zooplankton, respectively, at 20 °C, and \( q_{10} \) is the temperature-dependent growth rate.

In the phytoplankton growth term, the time rate of the chlorophyll-to-phytoplankton ratio, \( R = \text{Chl}/P, \) depends on light \( (E: \text{PAR}, \) in units of W/m²) and \( N \) availability:

\[ \frac{\partial R}{\partial t} = \mu_P \left[ 1 - \exp \left( - \frac{\alpha R_{\text{fl}} E}{\mu_P} \right) \right] \frac{R}{R_{\text{fl}}} \left[ \frac{N}{N + \kappa N} \right] R_{\text{l}} - (\Delta R) \min \left( \frac{E}{E_k}, 1 \right) - R \]  

where \( \Delta R = R_{\text{l}} - R_{\text{fl}} \). Extreme values of \( R \) for high-light \( (R_{\text{fl}}) \) and low-light \( (R_{\text{l}}) \) conditions are assumed to be 1.0 and 2.5 mg Chl/(mmol N), respectively.

PAR \( (\text{W/m²}) \) in the water column is calculated as

\[ E(z) = E_S \exp[-\kappa W z + \kappa C I_{\text{ch}}(z)], \]

where \( I_{\text{ch}}(z) \) is the depth-integrated chlorophyll \( (\text{mg/m²}) \) at depth \( z \) (m), and \( E_S \) is the surface PAR.

Descriptions and values of the biological parameters are listed in Table C1 in Appendix C.
Appendix C: Biological Parameters in Appendix B

Descriptions and values of biological parameters are provided in Table C1.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Unit</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$x_a$</td>
<td>Light attenuation due to chlorophyll-$a$</td>
<td>$(mg\text{ Chl/m}^2\text{s}^{-1})$</td>
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<tr>
<td>$x_w$</td>
<td>Light attenuation due to sea water</td>
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<td>$q_{10}$</td>
<td>Temperature-dependent growth rate</td>
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<tr>
<td>$\mu_{20}$</td>
<td>Maximum specific growth rate for phytoplankton at 20 °C</td>
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<tr>
<td>$\alpha$</td>
<td>Initial slope of P-E curve</td>
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<tr>
<td>$x_N$</td>
<td>Half saturation for nitrogen</td>
<td>(mmol N/m$^3$)</td>
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<tr>
<td>$R_l$</td>
<td>Minimum chlorophyll/phytoplankton ratio (for high-light conditions)</td>
<td>$mg\text{ Chl/(mmol N)}$</td>
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<tr>
<td>$E_K$</td>
<td>Light level at which photoadaptation starts</td>
<td>$W/m^2$</td>
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<td>$\eta$</td>
<td>Phytoplankton mortality rate constant</td>
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<td>$\lambda$</td>
<td>Ivlev grazing constant</td>
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<tr>
<td>$\gamma$</td>
<td>Fraction of grazing loss</td>
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<td>$n_1$</td>
<td>Linear zooplankton mortality rate constant</td>
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<td>$n_2$</td>
<td>Quadratic zooplankton mortality rate constant</td>
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<td>$e_1$</td>
<td>Fraction of linear rate of zooplankton loss that turns into detritus</td>
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<tr>
<td>$e_2$</td>
<td>Fraction of quadratic rate of zooplankton loss that turns into detritus</td>
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<td>$\delta$</td>
<td>Remineralization rate constant of detritus</td>
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<td>$W_{sink}$</td>
<td>Vertical sinking velocity relative to water for detritus</td>
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<td>$\rho_{PAR}$</td>
<td>Fraction of PAR in net shortwave radiation</td>
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Note. Most parameter values are taken from Liu et al. (2002). PAR = photosynthetically active radiation.

References


Chen, Y. J. (2004). Estimation of primary production in the South China Sea and observation of particulate organic matter at SEATS station: Application in the validation and improvement of the coupled physical-biogeochemical model of South China Sea, (Master's thesis, 124 pp.). Institute of Oceanography, National Taiwan University, Taipei, Taiwan.


