Numerical study of the effects of ocean color on the sea surface temperature in the southeast tropical Indian Ocean: the role of the barrier layer

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Abstract
The effects of biological heating on upper ocean temperatures in the southern tropical Indian Ocean region during the boreal summer were investigated by comparing the results of two modeling experiments using a solar radiation penetration scheme with and without chlorophyll effects. During the southeastern monsoon season, an increase in the chlorophyll concentration leads to cold anomalies off Java but warm anomalies off Sumatra. This contradictory effect is primarily caused by the difference in the barrier layer (BL) thickness in the two regions. Although the increasing phytoplankton tends to warm the surface and cool the subsurface in both regions, the existence of a thick BL in the region off Sumatra prevents cold anomalies from reaching the surface mixed layer (ML), whereas the thin BL off Java is favorable for the upwelling of cold subsurface anomalies into the surface ML, nullifying the warming effect of the increasing chlorophyll concentration.

Keywords: chlorophyll concentration, barrier layer, the southeast tropical Indian Ocean

1. Introduction
The absorption of shortwave radiation by the upper ocean mixed layer (ML) can be significantly affected by phytoplankton, the presence of which is now commonly measured using the chlorophyll concentration: phytoplankton trap more solar radiation in the topmost layer (e.g., Siegel et al 1995). However, a number of previous studies have found surface cooling in response to the reduction in shortwave penetration length-scale in numerical models (Nakamoto et al 2001, Murtugudde et al 2002, Sweeney et al 2005, Anderson et al 2007, Lin et al 2008). The situations under which such cooling is found and the exact mechanisms responsible for the cooling remain poorly covered in the literature so far, but regions where it has been found include the relatively stagnant shadow zones (Gnanadesikan and Anderson 2009) as well as the equatorial upwelling region (Sweeney et al 2005, Anderson et al 2007). The present study therefore proposes that it is regions in which there is both a shallow mixed layer and significant upwelling/entrainment from below this ML that will see the strongest responses. In these regions, the presence of both the shallow ML and upwelling/entrainment...
allow transformation of thermocline waters, and eventually affect the sea surface temperature (SST). The southeastern tropical Indian Ocean (STIO) represents a useful location to test this hypothesis.

Because the STIO is one pole of the Indian Ocean Dipole mode (e.g. Saji et al, Webster et al 1999), it is now considered to be one of the key areas of air–sea interaction. Upwelling only occurs in the STIO in boreal summer when the prevailing wind along the Sumatra and Java coasts is from the southeast (figure 1(b)). Figure 1 also shows the surface chlorophyll concentrations (shaded) in February and August from the STIO from the climatology of Sea-viewing Wide Field-of-view Sensor (SeaWiFS) ocean color data during 1997–2007. It is clear that high chlorophyll concentrations with maximum values exceeding 0.4 mg m$^{-3}$ occur along the southwestern coast of Sumatra and Java during August; an isopleth of 0.2 mg m$^{-3}$ extends south to 15$^\circ$S and west to 100$^\circ$E. This growth is mainly due to the abundance of nutrients brought into the mixed layer by the upwelling. In addition, the upwelling generally cools the SST in the STIO, especially off the coast of Java (data not shown).

Vertical heat transport can be strongly affected by the barrier layer (BL), which is a thin and stable layer between the bottom of the ML and the top of the thermocline. The BL is a unique feature that modulates the upper layer heat budget in the STIO. This effect has already been systematically investigated in previous studies (Qu et al 1994, Du et al 2005). Since the BL affects/modulates the connection between subsurface waters and the mixed layer, we hypothesize that it may also modulate the response of the SST to a change in the chlorophyll concentration. The purpose of this study is to test this hypothesis in the STIO region through numerical experiments.

**2. Model and experiments**

In the present study, the LASG/IAP (State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics/Institute of Atmospheric Physics) climate system ocean model (LICOM) (Liu et al 2004a, 2004b) is employed. LICOM is the oceanic component of the flexible global ocean–atmosphere–land system model (FGOALS) (Yu et al 2004), which is the model utilized in the Coupled Model Intercomparison Project phase 4 (CMIP4). The horizontal resolution of LICOM is 1$^\circ$ in the zonal direction, and the resolution varies from (1/2)$^\circ$ at the equator to 1$^\circ$ outside the tropics (10$^\circ$S–10$^\circ$N) in the meridional direction. The model has 30 vertical layers in total with 15 layers of 10 m in the top 150 m.

The two-exponential formula of Paulson and Simpson (1977) was used to parameterize solar radiation penetration in LICOM:

\[
Q_{\text{pen}} = Q_{\text{sw}} (A_1 e^{-\frac{z}{A_2}} + A_2 e^{-\frac{z}{B_2}}),
\]

where $Q_{\text{sw}}$ represents the net surface solar radiation and $Q_{\text{pen}}$ is the portion of solar radiation penetrating below a certain depth, $z$. The two terms represent the infrared and the visible–ultraviolet bands of solar radiation, respectively. The parameter $A$ is the fraction of radiation for a certain band; and $B$ is the e-folding penetration depth. The first term is usually absorbed in the upper 1 m, which is much smaller than the vertical grid distance (10 m); the latter can penetrate below 50 m in clear water. Thus, the variation in solar radiation penetration in the model is primarily caused by variations in the e-folding attenuation depth of the visible–ultraviolet band (B2).

Two 18 yr integrations were conducted starting from a 100 yr spin-up experiment and forced by the daily mean atmospheric variables from the Common Ocean-ice Reference Experiments (CORE) data set (Large and Yeager 2004) during 1990–2007. To reproduce the BL, the daily mean surface freshwater flux and an annual mean climatology of runoff from CORE were also used to drive the model. The only difference between the two experiments lies in the solar radiation penetration schemes. In the control run, hereafter referred to as NOCHL, the parameters of the two terms are assigned as constants following the method of Jerlov (1968). A1, A2, B1 and B2 are set to 0.58, 0.42, 1 and 23 respectively. To represent the influence of chlorophyll concentration on solar radiation penetration, an Ohlmann
Figure 2. The temperature averaged over the J region (105°E–120°E, 7°S–12°S) for (a) World Ocean Atlas 2005 (WOA05), (b) the NOCHL run and (c) the CLIMCHL run. (d)–(f) The same as (a)–(c), but for the S region (95°E–105°E, 5°S–2°N). The dotted and dashed lines are the mixed layer depth (MLD) and isothermal layer depth (ILD), respectively.

(2003) scheme was employed in the sensitive run, hereafter referred to as CLIMCHL. The four parameters in this scheme are all functions of chlorophyll concentration. The chlorophyll data from the climatology of SeaWiFS ocean color data collected during 1997–2007 were used. Figure 1 also shows the e-folding depth of the visible–ultraviolet band (contour) computed by the Ohlmann (2003) scheme. In the STIO region, the e-folding depth is shallower than 23 m at every location, i.e., more solar radiation will be absorbed by the ML in the CLIMCHL run than in the NOCHL run.

3. Results

The upper layer heat budgets in the regions off Sumatra and off Java, hereafter referred to as the S region (95°E–105°E, 5°S–2°N) and the J region (105°E–120°E, 7°S–12°S), respectively, are significantly different (Du et al 2005). The balance is mainly between the surface heat flux and the sub-grid processes in the S region, whereas upwelling is primarily balanced by horizontal heat transport of the Indonesian Throughflow (ITF) in the J region. The study domains are shown in figure 1.

To evaluate the performance of LICOM, we first compared the simulated temperatures in these two regions with observations from the World Ocean Atlas 2005 (WOA05) (Locarnini et al 2006). Figures 2(a)–(c) show the temperatures averaged over the J region for WOA05, the NOCHL run and CLIMCHL, respectively. The dotted and dashed lines are the mixed layer depth (MLD) and isothermal layer depth (ILD), respectively. The ILD is defined as an upper layer with a temperature higher than SST $-\Delta T$. The MLD is defined as the depth with a potential density lower than a specific density that is computed using the sea surface salinity and SST $-\Delta T$. Here, we selected $\Delta T = 0.8^\circ$C following the methods of Kara et al (2000) and Du et al (2005). The layer between the bottom of the ML and the isothermal layer is the so-called BL.

In general, the averaged temperature over the J region for the CLIMCHL run is slightly warmer than that for the NOCHL run in the topmost layer, and slightly colder below the topmost layer. However, because the effects of chlorophyll are not the primary determinants influencing the upper layer heat budget, the differences between the two experiments are small (figures 2(b) and (c)). The small cooling effects also can be found in the S region (figures 2(e) and (f)).

In order to properly simulate the heat budget it is necessary to capture the annual cycle of the thermocline and ML. The rise and fall of the thermocline reflect the upwelling and downwelling at the bottom of the ML, respectively, and the BL affects the connection between the thermocline and the ML. In the J region, the thermocline from the WOA05 has an annual cycle with one peak appearing in August, which is primarily caused by the upwelling driven by the prevailing southeast surface wind. LICOM reproduces the thermocline very well, but the peak of the thermocline occurs approximately one month later. Generally, LICOM also tends
to simulate a thinner BL in the J region. The annual mean BL thickness is approximately 4 m based on LICOM, but the observed thickness is approximately 7 m. During the boreal summer, the BLs derived from LICOM and observation almost disappear. Thus, the water in the thermocline can easily enter the ML.

Figures 2(d)–(f) show the temperatures averaged over the S region for WOA05, the NOCHL run and the CLIMCHL run, respectively. In this region, the observed thermocline has semiannual signals, with two peaks in March–April and August (figure 2(d)). The depression in the thermocline in June and November is associated with the equatorial downwelling Kelvin wave. Upwelling-favorable wind prevails in the boreal summer and raises the thermocline. LICOM can reproduce this semiannual variation of the thermocline although the peak in the boreal summer occurs in September, which is one month later than the observed peak. The S region is characterized by a thick BL; the annual mean BL thickness is approximately 20 m, which is almost three times as thick as that found in the J region. LICOM can simulate a thick BL in the S region, but it is thinner than the observed BL thickness. The thickness of the BL derived from LICOM is approximately 10 m in boreal summer, which is much thicker than that in the J region. This thick BL tends to isolate the ML from the effects of the thermocline water.

The above analyses show that LICOM has the ability to reproduce two key factors in the upper layer stratification found in the tropical eastern Indian Ocean region: the upwelling that occurs during the boreal summer and the thick BL found in the S region. These results indicate that LICOM is suitable for studying the effect of chlorophyll on the SST in this area. The discrepancies between WOA05 and the LICOM simulations are also significant. The horizontal and vertical resolutions of the model, the mixing parameterization and the biases in the forcing data are all possibly responsible for the discrepancies. The differences observed in SSTs, currents and heat budgets for the two regions based on the two experiments (CLIMCHL run and NOCHL run) are presented in the following paragraphs.

Figures 3(a) and (b) show the differences in SSTs (shaded) and surface currents (vectors) between the CLIMCHL and NOCHL runs for February and August, respectively. Differences exceeding 0.1°C primarily occur along the western coast of Sumatra and Java. In the boreal winter, the positive SST anomalies extend from the western point of Sumatra all the way to the islands east of Java (figure 3(a)), whereas in the boreal summer, negative anomalies occur off Java despite large heating effects due to the phytoplankton bloom (figure 1(b)). Thus, there must be other processes involved that nullify the extra biological heating.

The differences in temperature (shaded) at a 45 m depth are shown in figures 3(c) and (d) for February and August, respectively. This depth is approximately at the bottom of the ML along the eastern coast of the Indian Ocean. Because more solar radiation is trapped in the ML at shallower penetration depths due to a larger concentration of chlorophyll, negative temperature anomalies are observed in nearly all areas in both seasons. In the boreal summer, the large negative anomalies mainly appear along the coast of the J region where the chlorophyll concentration is large. This location is consistent with negative anomalies at the surface, and this result implies that there may be a connection between subsurface and surface temperature anomalies.

To better understand this dynamic ocean process, the currents observed in these two experiments were investigated. Figures 4(a) and (b) show the vertical velocities (vectors)
and temperatures (contours) along 9°S and 1°S in August for the NOCHL run. The blue curves also show the MLD (with circles) and ILD (without circles) for the NOCHL run. It is clear that the upwelling is much stronger and the ML and BL are both much shallower along the coast of Java than those along the coast of Sumatra. Moreover, the source of upwelled water in the J region is below the BL whereas the water in the S region is partly upwelled from the BL, where the temperature is almost as same as in the ML.

The contours in figures 4(c) and (d) show the temperature differences between the CLIMCHL and NOCHL runs along 9°S and 1°S with the MLD and ILD modeled by these two experiments. The largest cold anomalies in the J region lie just below the ILD with a magnitude exceeding 0.8°C. These cold anomalies are also the source of water for the ML. However, the small cold anomalies in the S region are located in the ML. Therefore, in the J region, the cold temperature anomalies resulting from the increase in chlorophyll concentration below the BL can be transported into the ML and cool the SST, whereas, in the S region, the warm water in the BL prevents the upper layer from being cooled by the upwelling.

As shown in figure 3, the current patterns have also been significantly altered. In the boreal summer, there are off-coast currents at the surface and towards-coast currents in the subsurface (figures 3(b) and (d)). This result implies that there is an anomalous upwelling in both regions; however, this kind of pattern is not observed in the boreal winter. The anomalous upwellings are clearly seen in figures 4(c) and (d), which show the differences in vertical velocities and temperatures between the two experiments along 9°S and 1°S, respectively. Lin et al. (2007) investigated current changes along the equator and found that the enhancements of the upwellings are primarily due to the chlorophyll concentration gradient through enhancing the meridional temperature gradient. The enhanced upwelling may also increase the cold vertical advection and cool the surface. However, it is interesting that the upwelling anomalies in the S region primarily occur above the ML; as a consequence, the anomalous upwelling cannot cool the ML to a significant degree due to the weak vertical temperature gradient.

To further support our observations of these processes, we computed the heat budget in the surface ML based on the method of Du et al. (2005). The ML temperature was derived as follows:

$$\frac{\partial T}{\partial t} = \frac{Q_0}{\rho C_p h} - \frac{Q_{pen}}{\rho C_p h} - \vec{V} \cdot \nabla T - \frac{w(T - T_d)}{h} + R,$$

where $T$ is the ML temperature, $h$ is the ML depth, $C_p$ is the specific heat capacity of sea water which is assigned as 3996 J (kg K)$^{-1}$, $Q_0$ is the net surface heat flux, $Q_{pen}$ is the downward radiative flux across the MLD, $\vec{V}$ is the vector of the horizontal velocity, $w$ is the entrainment rate, $T_d$ is the
temperature just below the ML and $R$ is the residual term. The third and fourth terms on the right-hand side in the above equation are the horizontal advection and entrainment terms, respectively, and $w$ is defined as follows:

$$w = \frac{\partial h}{\partial t} + w_b + V \cdot \nabla h,$$

if $\frac{\partial h}{\partial t} + w_b + V \cdot \nabla h > 0$,

$$w = 0, \quad \text{otherwise,}$$

where $w_b$ is the velocity at the base of the ML.

Here, we only examine the climatology of the heat budget between the two experiments. Table 1 shows the difference in the ML heat budget between the CLIMCHL and NOCHL runs for the J and S regions in August. The large warming effects are mainly due to the solar radiation penetration terms, $0.21 \, ^\circ C/\text{month}$ for the J region and $0.14 \, ^\circ C/\text{month}$ for the S region. The former term is primarily balanced by the entrainment, $-0.27 \, ^\circ C/\text{month}$, and the latter term is balanced by the residual term, $-0.11 \, ^\circ C/\text{month}$, which primarily includes the sub-grid processes, such as turbulent mixing. This observation further confirms the inference above: the entrainment of cold water into the ML leads to the cold anomaly in the J region, and the thick BL prevents the cold water from affecting the surface layer in the S region.

4. Summary

This letter investigated the effects of biological heating on upper ocean temperature in the STIO region by comparing two experiments using a solar radiation penetration scheme with and without chlorophyll effects. In the boreal summer, an increase in the chlorophyll concentration leads to cold anomalies off Java and warm anomalies off Sumatra. The upwelling driven by the prevailing wind transports nutrients from the thermocline and results in a phytoplankton bloom during this season. This tends to warm the surface and cool the subsurface due to a 1D thermodynamic process. However, because of the thin BL off Java, the upwelling can bring the cold subsurface anomalies directly into the ML and nullify the warming provided by the phytoplankton bloom. In the region off Sumatra, the existence of a thick BL prevents the cold anomalies from entering the surface ML and causes warm anomalies to occur off Sumatra, although upwelling also occurs. The process observed in the Java region is similar to that in the eastern equatorial Pacific, where upwelling also dominates the heat budget in the absence of a BL.

As noted by previous studies (e.g., Du et al. 2005), the annual mean SST in the STIO is much warmer than that in the eastern equatorial Pacific due to the existence of the BL and ITF. The present study shows that the BL can also reduce the impact of changes in shortwave penetration on SST by limiting the impact of cold upwelled water.

The magnitude of the SST anomaly in this study is small at approximately $0.1 \, ^\circ C$. This result is mainly due to the lack of atmosphere–ocean coupled processes in the stand-alone ocean model. Gnanadesikan and Anderson (2009) suggest that the impact on SST might quadruple when atmospheric temperatures are allowed to adjust. Identical experiments conducted using the coupled model are necessary for further investigation.

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Table 1. The differences of heat budget between the CLIMCHL and NOCHL runs for the J and S regions in August.

| Term | J region (°C/month) | S region (°C/month) |
|------|---------------------|---------------------|
| $\partial T / \partial t$ | 0.08 | −0.02 |
| $Q_h / \rho C_p h$ | 0.14 | −0.04 |
| $- Q_{pen} / \rho C_p h$ | 0.21 | 0.14 |
| $- V \cdot \nabla T$ | −0.06 (U: 0, V: −0.06) | 0.03 (U: 0, V: 0.03) |
| $- w (T - T_d) / h$ | −0.27 | −0.04 |
| $R$ | 0.06 | −0.11 |

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