

Chlorophyll modulation of mixed layer thermodynamics in a mixed-layer isopycnal general circulation model – An example from Arabian Sea and Equatorial Pacific

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Abstract

Western tropical Indian Ocean, Arabian Sea and equatorial Pacific are known to be regions of intense bio-chemical-physical interactions with Arabian sea being largest in seasonal phytoplankton bloom while the equatorial Pacific bloom being perennial and supported by quasi-steady upwelling. Remotely sensed chlorophyll pigment concentrations from the Coastal Zone Color Scanner (CZCS) are used to investigate the chlorophyll modulation of ocean mixed layer thermodynamics in a bulk mixed-layer model, embedded in the ocean isopycnal general circulation model (OPYC).

A higher abundance of chlorophyll increases absorption of solar irradiance and heating rate in the upper ocean, resulting in decreasing the mixed layer thickness than they would be under clear water condition. These changes in the model mixed layer were consistent with Joint Global Ocean Flux Study (JGOFS) observations during the 1994-1995 Arabian Sea experiment and epi-fluorescence microscopy (EFM) on samples collected during Equatorial Pacific Ocean Climate Study (EPOCS) in November, 1988. In the Arabian sea, as the chlorophyll concentrations peak in August ($3\text{mg}/\text{m}^3$) due to summer plankton bloom induced by coastal upwelling, the chlorophyll induced biological heating enhanced the sea surface temperature (SST) by as much as 0.6°C while sub-layer temperature decreases and sub-layer thickness increases. In the equatorial Pacific, modest concentrations of chlorophyll less than $0.3\text{mg}/\text{m}^3$ is enough to introduce a meridional differential heating that results in reducing the equatorial mixed layer thickness by more than 20m. The anomalous meridional tilting of the mixed layer bottom enhances off equatorial westward geostrophic currents. Consequently, the equatorial undercurrent transport more water from west to east.

1. Introduction

Changes in climate are related to different modes of operation of the ocean-atmosphere system as well as land-ocean system involving bio-chemical processes, which are driven by insolation fluctuation spatially and temporally on different scales. The key physical features are water transport in the atmosphere and salt transport in the ocean, which together determine ocean density and circulation patterns. Changes in ocean circulation, thus, affect the exchange of trace gases including carbon dioxide between the ocean and atmosphere as well as chemical components and biological components in the ocean. The potential for feedback effects on the earth system exists through both physical and bio-chemical reactions in the ocean-atmosphere-land system.

Marine phytoplanktons influence the air-sea exchanges of carbon dioxide, sulfur compounds, and ocean ra-

diation transfer as well as albedo of ocean surface water, which affect earth climate system through thermodynamical processes within the closed system of the planet earth.

It has been recognized that absorption of solar radiation and the resulting local heating within the upper ocean are strongly influenced by chlorophyll concentration (e.g. Ivanoff, 1977; Simpson and Dickey, 1981a,b; Simonot et al., 1988; Lewis et al., 1990; Large et al., 1994; Morel and Antoine, 1994). Sathyendranath et al. (1991) demonstrated a significant biological impact on the Arabian Sea's sea surface temperature (SST) : assuming that SST and vertical temperature gradient at the mixed layer base are known at the start of each month, they integrated the model over one month using monthly mean air-sea fluxes for momentum and heat to obtain new fields. Martin (1985) compared different mixed layer models for some simple,

idealized forcing cases and tested the models using mixed-layer observational data. He found that the model simulation was particularly sensitive to seawater turbidity when the penetration of solar radiation below the mixed layer is capable of influencing the temperature development at the mixed layer base. The changes in ocean turbidity affected calculated SST by 3°C or more, and the change due to turbidity was as large as 1°C on daily time scales (Martin, 1985). Here, we investigate the role of chlorophyll pigment concentration on influencing mixed layer thermodynamics in the Arabian sea and in the eastern equatorial Pacific using an ocean isopycnal general circulation model.

2. Model and Forcing

Since the mixed layer depth is influenced not only by local vertical mixing but also by horizontal convergence of mass and heat, a mixed layer model must consider both full dynamics due to the use of primitive equations and a parameterization for the vertical mass transfer and related quantities across the mixed layer base. The primitive equation ocean model used in this study, hereafter referenced to as OPYC, is an updated version of the model developed by Oberhuber (1993). It is an isopycnal model with an actively coupled mixed layer and detailed flux parameterization. The model covers global ocean from North pole to Antarctica, and the model resolution is 1 degree zonally and 1 degree meridionally with thirteen vertical layers consisting of the top mixed layer and twelve underlying isopycnal layers. A realistic topography is used from NOAA topography, which is based on 5 minute resolution. In the OPYC model it is assumed that a fraction of the solar radiation penetrates through the ocean and has a penetration depth of 23 m (Paulson and Simpson, 1977), hereafter referred to as clear-water parameterization (A).

In order to study the effect of chlorophyll pigment on mixed layer dynamics, the clear-water parameterization (A) is replaced by an alternative parameterization accounting for a heating rate by chlorophyll pigments within the upper ocean (Morel and Antoine, 1994), hereafter referred to as chlorophyll-pigment parameterization (B). This parameterization allows one to calculate the vertical profiles of heating rate from remotely-sensed chlorophyll pigment concentration data, considering partitioning between "visible" and "near IR" radiation and the vertical profile of heat deposition (Morel and Antoine, 1994).

Note that the clear water parameterization implicitly includes chlorophyll pigment effects, but corresponding to only a few hundredths of mg/m^3 . The surface

forcing of the OPYC model, when used with either (A) or (B), consists of a climatological monthly mean seasonal cycle of wind stress, total surface heat flux, and turbulent kinetic energy input. Surface mixed layer salinity is relaxed towards the Levitus annual mean salinity fields, while the salinity in the interior isopycnal layers is dynamically determined without such a constraint. Instantaneous atmospheric fields are interpolated from the monthly means. Monthly mean climatology of chlorophyll pigment concentrations were obtained from the Coastal Zone Color Scanner (CZCS) from November 1978 to June 1986 (Feldman et al., 1989). The control run with clear-water parameterization (A) of OPYC was carried out for 50 years, which is sufficient to obtain a cyclo-stationary state for the upper ocean, i.e. the thermocline. Because no interannual variability can develop when Newtonian

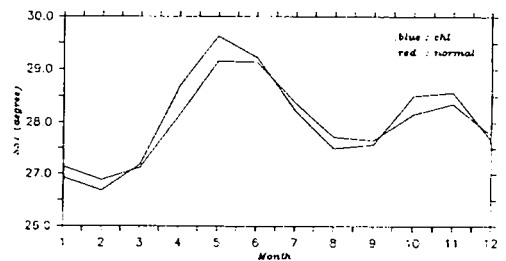


Fig. 1. Time evolution of zonal averaged (along 20°N) sea surface temperature (SST) in the Arabian Sea.

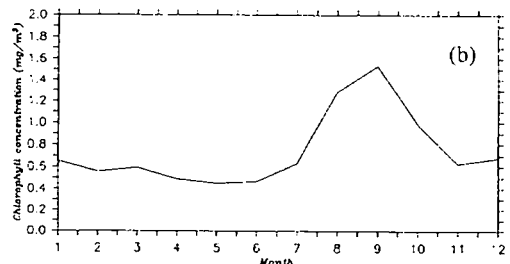
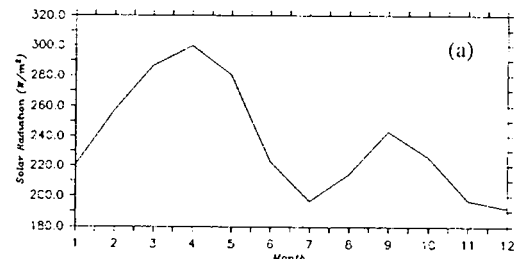


Fig. 2. Time evolution of zonal averaged (along 20°N) (a) incoming solar radiation (w/m^2) and (b) chlorophyll pigment concentration (mg/m^3) in the Arabian Sea. Solar radiation is based on Reed (1977) while chlorophyll is from CZCS.

feedbacks towards climatological data are present, it was sufficient to take the restart conditions from the last year of this 50 year control-integration (A). The experiments using chlorophyll pigment parameterization (B) were conducted for additional 10 years, until another cyclo-stationary state for global ocean SST was obtained.

3. Arabian sea surface temperature

Model results of upper ocean biological heating on a zonal strip along 20°N in the Arabian Sea are discussed with JGOFS time-series measurements during summer monsoon (Dickey et al., 1998) for this region. The annual evolution of SST (Fig.1) shows higher values

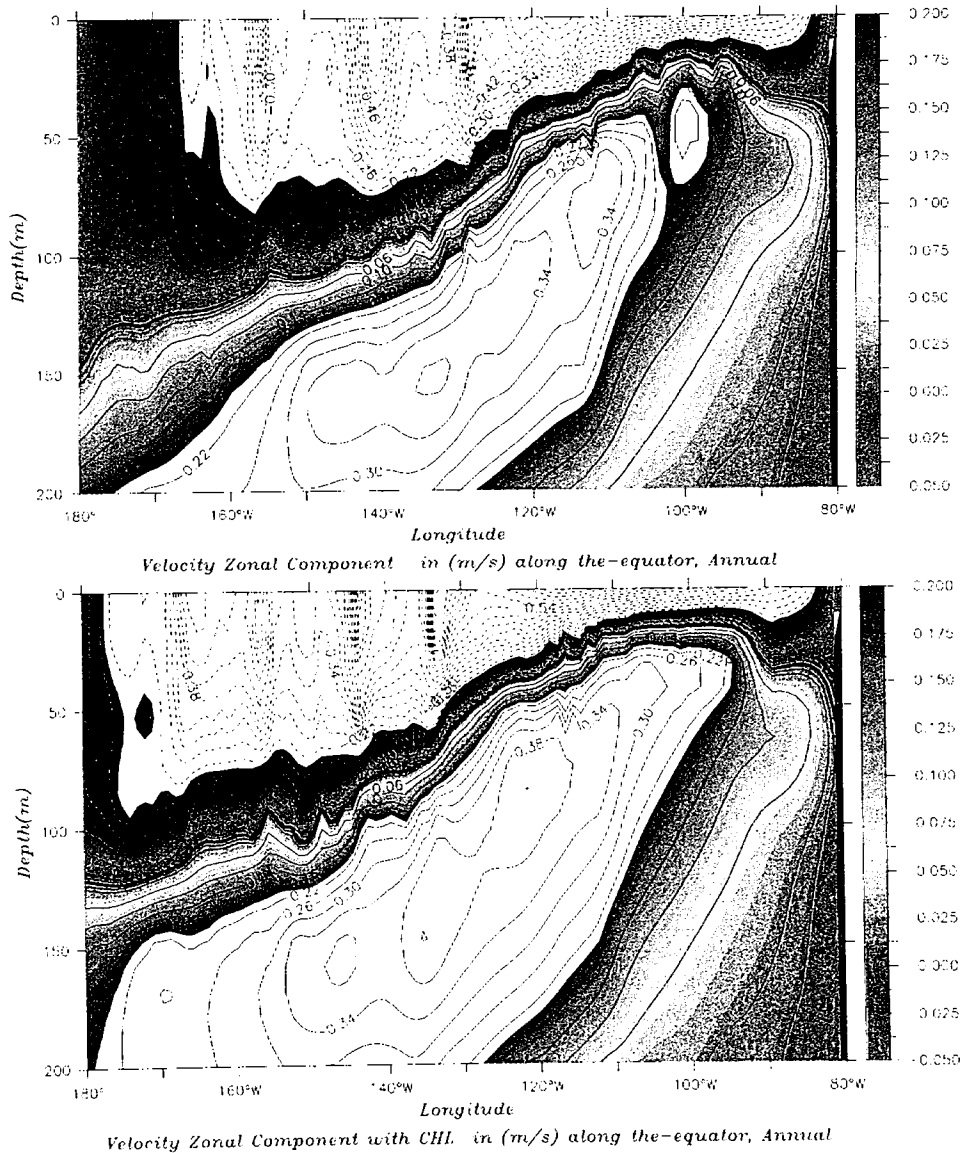


Fig.3. Vertical distribution of annual mean zonal velocity (m/s) (a) without and (b) with chlorophyll pigment concentration along equatorial Pacific.

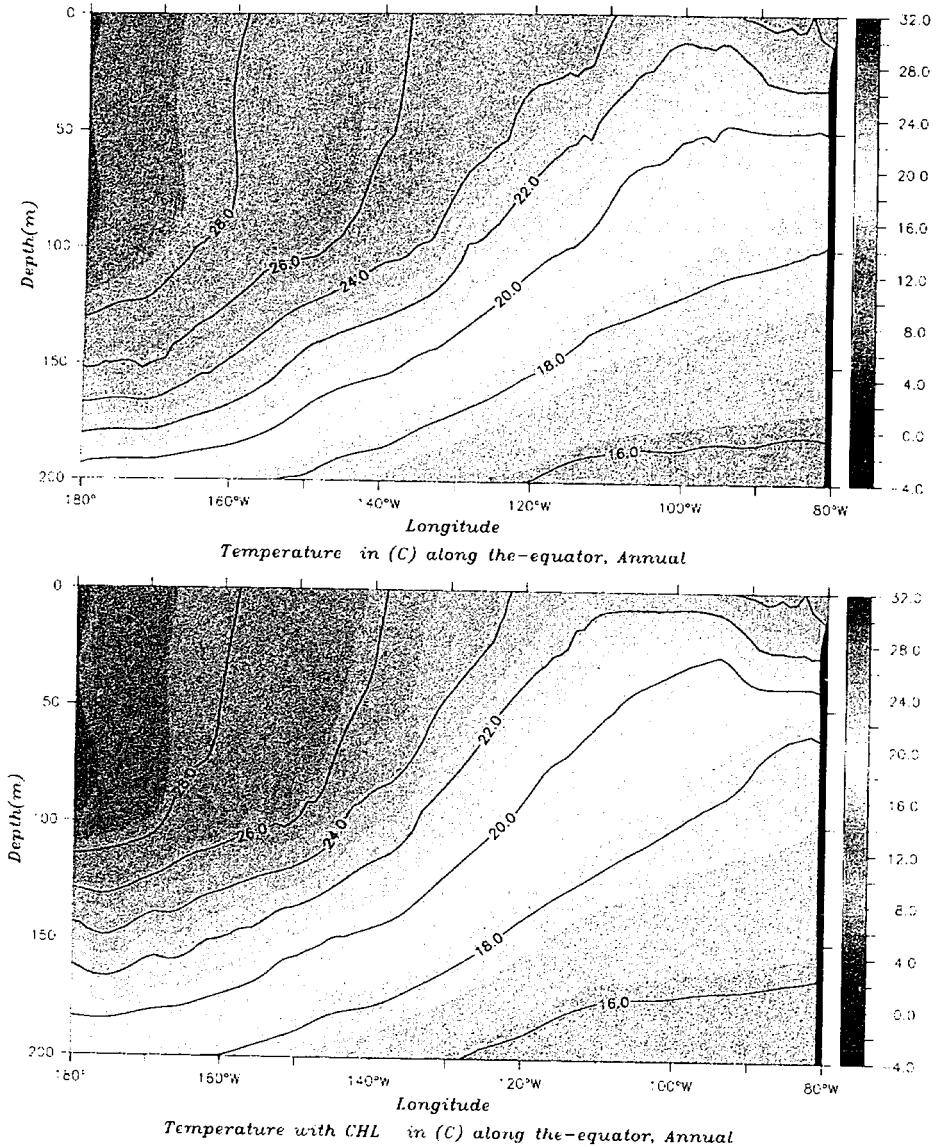


Fig. 4. Vertical distribution of annual mean temperature (a) without and (b) with chlorophyll pigment concentration along equatorial Pacific.

twice a year, from March to June and from September to November, when forced with chlorophyll pigment concentrations (B) as compared to the normal run (A). This is the effect of radiative heating by chlorophyll's absorption of incoming solar radiation that we refer to as biological heating of the upper ocean. It is interesting to note that the maximum anomalous SST, defined as the difference between the SST obtained from experiment B and that from A, during March-June period is 0.2°C higher than during the September-November period. This could be understood by analyz-

ing the time evolution of zonally averaged solar radiation (Fig.2a) and chlorophyll pigment concentration (Fig.2b). The incoming solar radiation shows a biannual cycle, with primary heating during March to June (Spring Intermonsoon) and the secondary heating during August to October (Fall Intermonsoon). The high chlorophyll pigment concentrations during August-September is associated with Arabian upwelling. When solar radiation peaks during Spring Intermonsoon, the low chlorophyll concentrations ($\sim 0.4 \text{ mg/m}^3$) significantly heats the upper ocean. If the solar radiation in

fall Intermonsoon is kept at the same magnitude as that of spring Intermonsoon, then the observed four-fold increase of chlorophyll pigment concentration would lead to four-fold positive values for the anomalous SST. However, in the real ocean, the decrease of solar radiation from Spring to fall Intermonsoon counteracts the net biological heating in the upper ocean to make the anomalous SST smaller in the fall Intermonsoon season than the spring Intermonsoon season.

Our numerical model experiment suggested that when chlorophyll pigment concentration peaks in October, the upper ocean warms and the subsurface cools through the effect of vertical distribution of absorbed solar energy, with more energy absorbed in the surface layers and less energy penetrating towards the lower layers than the case without chlorophyll pigments. This is in conformation with the U.S. JGOFS moored buoy measurements near the Arabian coast (15.5°N, 61.5°E). The time-series data showed a similar surface warming and subsurface cooling with increased chlorophyll pigment concentration, biological heating rate, and decreased mixed layer thickness during the southwest monsoon period (August-September) (Dickey et al., 1998). These authors speculate that the changes may be associated with cool meso-scale dynamical, i.e. cold core ring, features. The observed synchronization between upper water warming and lower water cooling, when analysed in the light of the model result could clearly be interpreted as the result of biological absorption of solar radiation in the water column.

4. Anomalous sub-surface currents in the equatorial Pacific

In the previous section, our model results indicate that by absorbing solar radiation chlorophyll pigments influence the mixed layer heat budget significantly: SST increases by up to 0.6°C and mixed layer thickness decreases due to the biological heating, whereas sub-layer temperature decreases due to reduced penetration of solar radiation.

Overall assessment of the influence of chlorophyll distribution on global scale must be needed. In the equatorial Pacific, the mixed layer bottom must rise where chlorophyll shifts the solar radiation towards surface, leading to geostrophically adjusted westward current, with equatorial wave-guide length scale of 500 km, in both the hemispheres in the tropical Pacific. From model result the maximum velocity of 3 cm/sec is generated through chlorophyll forcing in the 4th layer, in which Equatorial Undercurrent (EUC) is situated. These westward anomalous currents arrive at the west-

ern boundary in the tropical Pacific to join the EUC, thus strengthening the EUC. This is clearly discernible from the vertical distribution of model zonal velocity along the equator (Fig.3). This shows increased speed in the core of EUC, apart from its broadening in presence of chlorophyll (Fig.3b) as compared to the normal run (Fig.3a). In the eastern equatorial Pacific, the anomalous water masses transported by strengthened EUC must upwell towards the surface. This is evident from the vertical distribution of temperature along the equator (Fig.4). Note that the 23°C isotherm, which is seen just below the surface at 100°W in the normal run (Fig. 4a) surfaces and occupies the region between 115°W to 95°W when forced with chlorophyll (Fig.4b). This enhanced EUC would transport iron-rich, cold water from the deeper (~ 250 m) western equatorial Pacific to the upper layers (~ 50 m) of the eastern equatorial region. Model results show that this upwelling is accompanied by mixed layer lifting with subsurface-layer thickening (not presented), while the equatorial upwelling induce by trade wind do not accompany such subsurface layer thickening. Since direct measurement to verify the model result do not exist presently, we must seek for an alternative or indirect method.

During La Nina of 1988 (Kirk, 1988), the EPOC observation was conducted in the eastern equatorial Pacific (Chaves et al., 1990) which showed uplift of the isotherm and nitrate isopleths towards the central and eastern Pacific (Figure 5 of Chaves et al., 1990). They interpreted high chlorophyll concentration as a secondary effect of upwelling that strengthened during La Nina period. However, the subsurface layer between 14°C and 13°C isotherm thickened, and the layer of nitrate concentration of 20-24 micro-mol also thickened. It is to be noted that conventional wind driven equatorial upwelling does accompany such subsurface layer thickening. Neither could one explain the increase of nitrate concentration in the subsurface layer (~150m) between 140°W and 110°W without invoking horizontal transport. In the light of above, we feel that phytoplankton's heating effect in the eastern equatorial Pacific during La Nina period in 1988 might have caused the thickening of subsurface layers manifested by temperature and nitrate isopleths.

5. Discussion

Observed spatial and temporal anomalies derived from the JGOFS Arabian Sea process study and EPOCS biophysical observations during La Nina period of 1988 are in agreement with results obtained from the model simulation. The spatial and temporal variation of both SST and chlorophyll pigments leads to regional-scale variations in the atmosphere on similar scales. It should

be noted that the use of OPYC's standard heat flux parameterization implicitly contains some Newtonian feedback for the SST, thus modeled SST anomalies might be more pronounced if a coupled ocean-atmosphere model with more realistic feedback is used.

The sub-layer thickening and cooling associated with the mixed-layer thinning and warming reduces vertical mixing in the upper thermocline, and redistributes nutrients, besides heat and salt. The secondary effect of biological heating due to chlorophyll pigment may be very important as in the case of equatorial Pacific. Here, the intensification of EUC results in excess transport of deep waters rich in iron from the western equatorial Pacific to the euphotic zone in the eastern region. This has the potential to increase the carbon fixation in the eastern equatorial HNLC (high-nitrate low-chlorophyll) region where biological production is limited by macro-nutrient like iron. This calls for more studies to understand the implication of biological heating, by using more accurate chlorophyll pigment concentrations and coupling the ocean general circulation model with a carbon cycle model.

Acknowledgments

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