Phytoplankton and thermal structure in the tropical ocean

Phytoplankton
Tropic
Solar heating

Phytoplancton
Tropique
Ensoleillement

Marlon R. LEWIS
Department of Oceanography, Dalhousie University, Halifax, Nova Scotia B3H 4J1, Canada.

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ABSTRACT

Variation in the absorption of solar irradiance in the upper ocean is strongly influenced by variation in the concentration of phytoplankton. In optically clear tropical regions, where the mixed layer depth is shallow relative to the depth of significant penetration of irradiance ($\lambda < 700\ nm$) all year, such variation is particularly important for the thermal structure and dynamics of the upper ocean. The nonuniformities in vertical chlorophyll distribution associated with " typical tropical structure " can give rise to local heating rates that increase with depth ; the vertical motions that may result are discussed. The penetration of irradiance through the mixed layer is a downward energy flux which is shown to be of the same order as that thought to be transported vertically by turbulent fluid motions. Variability in sea-surface chlorophyll in time and in the horizontal results in variation in the vertical radiative energy flux out of the base of the upper mixed layer ; it is suggested that this variability may be important for tropical and global heat budgets.

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RÉSUMÉ

Phytoplancton et structure thermique dans l'océan tropical

Les variations de concentration du phytoplancton influencent fortement l'absorption de l'énergie solaire dans la couche supérieure de l'océan. La variation d'absorption est particulièrement importante pour la structure thermique et la dynamique de la couche supérieure de l'océan dans les régions tropicales optiquement claires, où la couche mélangée est peu profonde pendant toute l'année par rapport à la profondeur de pénétration significative de la lumière ($\lambda < 700\ nm$). Les différences dans la distribution verticale de la chlorophylle associées avec la " structure tropicale typique " peuvent entraîner un échauffement qui augmente avec la profondeur. Les mouvements verticaux pouvant en résulter sont discutés. La pénétration de la lumière dans la couche mélangée correspond à un flux d'énergie dirigé vers le bas, que nous estimons du même ordre de grandeur que celui transmis verticalement par les mouvements fluides turbulents. La variabilité temporelle et horizontale de la chlorophylle de surface provoque des variations du flux vertical d'énergie radiative qui traverse la base de la couche supérieure mélangée ; il est suggéré que cette variabilité peut être importante pour les bilans thermiques tropical et global.

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INTRODUCTION

The dominant source term in tropical, oceanic heat budgets is the absorption of solar radiation (*e.g.* Reed, 1983). Approximately half of the incoming radiation is in spectral bands with wavelengths longer than 700 *nm* and can properly be treated as a surface flux ; the balance is associated with shorter, more penetrating wavelengths. The parameterization of the vertical distribution of radiation absorption, as well as its absolute magnitude, is a strong source of variability in mixed layer models which have included this dependence (Kraus, Rooth, 1961 ; Denman, 1973 ; Simpson, Dickey, 1981 ; Dickey, Simpson, 1983 ; Woods, 1980).

For most tropical waters removed from terrestrial influences, variations in scattering and absorption coefficients, the optical properties which determine the vertical distribution of irradiance, are due to variations in biological constituents, primarily phytoplankton and their breakdown products. Variability in the vertical distribution of irradiance absorption is particularly important in tropical waters because of the potential deep penetration of visible wavelengths relative to a shallow mixed layer (Woods *et al.*, 1984). Here we would like to consider how variability in phytoplankton concentration, and hence the depth dependence of irradiance absorption, might have a significant effect on thermal structure and physical dynamics in the upper, tropical ocean.

PHYTOPLANKTON AND ABSORPTION OF IRRADIANCE

Conservation of energy and mass for incompressible flow requires that for an annual average,

$$c_p\rho\,(\nabla \cdot \boldsymbol{U}T + \nabla \cdot \boldsymbol{U}'T') = \nabla \cdot \boldsymbol{E} \qquad (1)$$

and

$$\nabla \cdot \boldsymbol{U} = 0 \qquad (2)$$

where T is the mean temperature, $c_p\rho$ is the product of specific heat and density, **U** is the mean velocity, and **E** is the vector irradiance. The primed quantities refer to deviations from the mean. For now we concentrate only on the term on the right-hand side ; we will return to the full equation below.

To a good approximation for clear ocean waters (Jerlov, 1976), the radiative flux divergence can be simplified such that,

$$\nabla \cdot \boldsymbol{E} = \frac{\partial E_D}{\partial Z} = -K_T E_D\,, \qquad (3)$$

where E_D is the downward irradiance. The attenuation coefficient for downward irradiance, K_T, has a strong spectral dependence ; for the most accurate work, one could express this dependence explicitly and convolute the attenuance spectrum with the downward irradiance spectrum to calculate the local heating rate (*e.g.* Woods, 1980). The error associated with spectrally averaging the coefficient and irradiance is small for waters deeper than a shallow surface layer however, since the ratio of total quanta to total energy in the visible band is relatively constant below this depth (*see* Morel, Smith, 1974). In addition, all spectral models collapse to produce the same temperature profile as wind velocities approach 20 ms^{-1} (Simpson, Dickey, 1981 *b*).

The attenuation coefficient to first order can be partitioned into a part due to attenuation by water alone, and a part due to attenuation by phytoplankton pigments, specified here as chlorophyll (*e.g.* Lorenzen, 1972).

$$K_T = K_W + K_C B \qquad (4)$$

where K_W is the clear water attenuation coefficient (Morel, Prieur, 1977 ; Smith, Baker, 1981), and K_C is the chlorophyll specific optical cross-section. The pigment concentration, B, in practice is associated with numerous other auxiliary pigments which contribute to variations in K_C. It is clear from this parameterization how variations in pigment concentration might lead to variation in the vertical distribution of local heating ; to fix the order of magnitude of the influence of phytoplankton absorption, note that the e-folding optical depth scale in the rich upwelling area off Peru can be ca. 5 *m* or less compared to ca. 30 *m* in the relatively impoverished offshore waters.

VERTICAL STRUCTURE OF ABSORBING PHYTOPLANKTON PIGMENTS

The most outstanding feature of the numerous profiles of chlorophyll taken in the tropical regions is the presence of a sub-surface chlorophyll maximum layer. This feature is a portion of what has been called the " typical tropical structure " (*e.g.* Herbland, 1983) where a strong thermocline separates a mixed layer low in chlorophyll and nutrient, from relatively nutrient rich waters below (*see* Fig. 1). The profile is

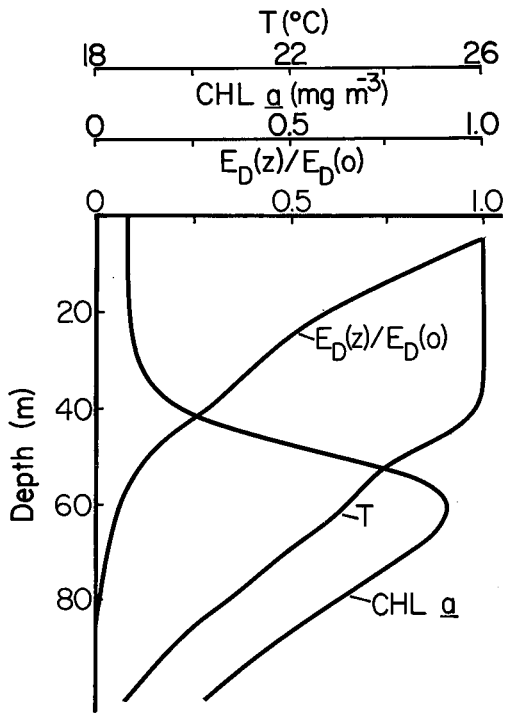

Figure 1

Vertical variations in temperature, chlorophyll a *and irradiance which characterize the " typical tropical structure ". Data from Lewis* et al.*(1983).*

from the Canary basin (Lewis *et al.*, 1983). Given such pronounced variation in chlorophyll concentration, one might expect, based on equations (3) and (4), corresponding variation in light absorption and consequently heating rates. A series of stations off the coast from Cap-Blanc, where both vertical irradiance profiles and chlorophyll profiles were measured, showed clearly this effect (Spitzer, Wernand, 1981). The vertical gradient of scalar irradiance exhibits a sharp increase at depths which decrease with distance offshore (Fig. 2 *a*). The factor responsible for this decrease is immediately identified with the sub-surface chlorophyll maximum when the depth corresponding to increase in attenuation is plotted against depth of the upper boundary of the chlorophyll maximum layer (Fig. 2 *b*). A similar result is found in the data of Cullen and Eppley (1981), as discussed in Lewis *et al.* (1983).

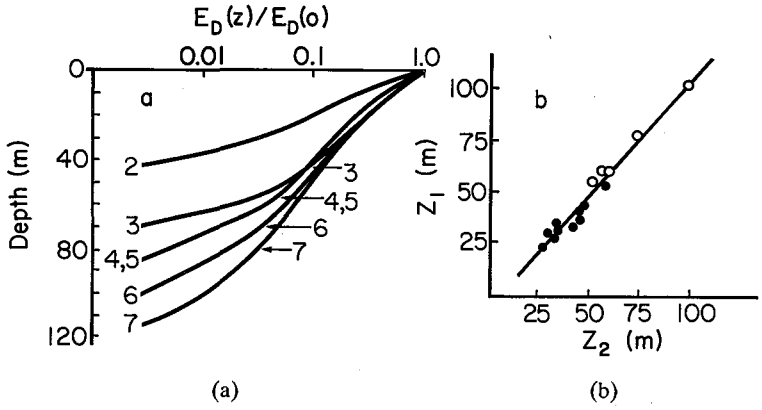

Figure 2

a) Irradiance profiles from the tropical Atlantic with increasing numbers corresponding to increasing distance offshore. Depth of the leading edge of the chlorophyll maximum layer is indicated ; b) relationship between the depth of the chlorophyll maximum (Z_2) and the depth associated with the break in the slope of the irradiance profile (Z_1). • indicate stations in addition to those in a). Redrawn from Spitzer and Wernand (1981). All stations were occupied on a transect at 20°N. Distance from shore was ca. 100 km (2), 475 km (3), 1 000 km (4), 1 425 km (5), 2 000 km (6), 2 400 km (7).

Lewis *et al.* (1983) considered the possibility that radiation absorption associated with these subsurface layers might be intense enough to induce a thermal inversion. Non-dimensional criteria for potential instability were developed based on empirical parameterization of the chlorophyll distribution as a Gaussian function of depth. One station was identified from the tropical Atlantic (Fig. 3 *a*) which showed potential for reversal of the thermal heating gradient (Fig. 3 *b*) ; however, it was concluded that turbulent processes were probably sufficiently strong enough to diffuse heat away before convection could occur on the leading edge of the chlorophyll maximum layer.

The heat flux associated with absorption by the chlorophyll maximum layer may not be trivial (*see* Fig. 3 *b*). The chlorophyll maximum layer exists at a depth where irradiance levels are between 1 and 15 % of the visible radiation flux at the sea surface (Cullen, 1982) ; this represents a downward energy flux through the base of the mixed layer of order 1-15 Wm^{-2}. The water thus heated is not in immediate contact with the atmosphere and this energy is ultimately transported to higher latitudes (Woods *et al.*, 1984).

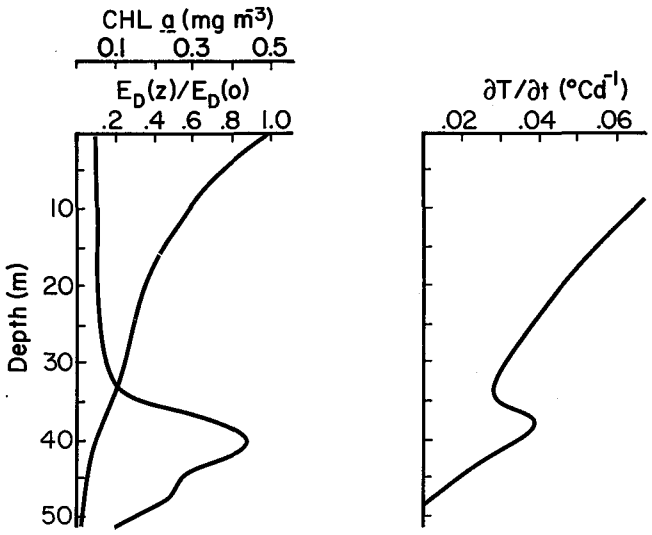

Figure 3

a) Vertical distribution of chlorophyll and irradiance from a station in the tropical Atlantic ; b) computed local heating rated based on a). Data from Lewis et al. *(1983).*

CONSEQUENCES FOR EQUATORIAL HEAT BUDGETS

Niiler and Stevenson (1982) recently integrated equation (1) over a tropical ocean basin, enclosed at the top by the sea surface, on two sides by insular continental margins, and at the northern, southern and basal boundaries by a surface of constant temperature. By virtue of equation (2), the net advective terms vanish and the solution is a balance between the net surface heat flux and a heat flux associated with turbulent fluid motions,

$$c_p\rho \int\int_{A_T} \overline{U_n'T'}\, ds = \int\int_{A_s} Q_0 dA \,, \qquad (5)$$

where ds and dA are the three-dimensional subsurface and two-dimensional sea-surface elements respectively, U'_n is the velocity fluctuation normal to ds and Q_0 is the net sea-surface heat flux. If, as suggested by Niiler and Stevenson (1982), heat flux associated with fluctuating Ekman transport is a small fraction of the left-hand side in equation (5), then the dominant balance is between net surface flux and turbulent, downward heat flux associated with small scales of fluid motion. A lower constraint on the magnitude of vertical diffusion can then be calculated.

Such calculations may be too high however, if a portion of the required heat flux passes the lower boundary of the volume as light rather than as warm

water. The vertical integration implied in Q_0 implicitly assumes an absorption rate sufficiently high that the downward radiation term is treated as a surface flux only ; simple calculations show that this may be a poor assumption, and furthermore, that the downward irradiation at the base of the volume is high enough to account for much of the required downward heat flux in the equatorial ocean.

The boundary of the " warm-water pool " can be defined by the 25 °C isotherm in the Atlantic (Niiler, Stevenson, 1982). If the depth of the layer is, say, 50 m, the chlorophyll concentration in the upper mixed layer is ca. 0.1 mg chl m^{-3} (*e.g.* from Herbland 1983 in the tropical Atlantic) and K_C and K_W take typical values [0.02 m^2 (mg chl)$^{-1}$, 0.036 m^{-1} respectively], then the visible radiation flux at the base of the layer is ca. 15 % of that at the sea-surface. For a sea-surface visible radiation flux of order 100 Wm^{-2}, this represents a vertical radiation flux at the base of the 50 m mixed layer of order 15 Wm^{-2}. For purpose of comparison, Niiler and Stevenson require a vertical turbulent heat flux of 7-8 Wm^{-2} for the Atlantic and 18-22 Wm^{-2} for the Pacific to balance the net surface heating. Of course, heat produced by absorption of irradiance below the mixed layer must be dissipated as well, by turbulent and advective processes.

The assumption of a 50 m mixed layer depth does not hold over much of the western basin. However, the areas of highest net surface heating are in the eastern basin (Niiler, Stevenson, 1982), where the mixed layer depth is 25-50 m (Lamb, 1984) and concentration of chlorophyll is the highest and most variable (Feldman *et al.*, 1984).

The relatively high values of energy and temperature variance dissipation rates necessary for the balance in equation (5) are not required if there is significant penetration of irradiance. One independent constraint on the magnitude of the vertical diffusivity arises from the existence of a maximum in chlorophyll under dimly lit conditions. There are numerous ways such maxima can form (Cullen, 1982) ; for the tropical Atlantic however, Herbland (1983) has concluded that growth in situ is the dominant term in the chlorophyll budget. A typical net growth rate at 10 Wm^{-2} for phytoplankton populations in the chlorophyll maximum in the tropical Atlantic is ca. 10^{-6} s^{-1} (*e.g.* Platt *et al.*, 1983 ; Fasham *et al.*, 1985). If the time scale associated with turbulent diffusion is l^2/K_v, then the eddy coefficient, K_v, must be smaller that ca. 3.10^{-5} m^2s^{-1} for a chlorophyll maximum with a length scale of 5 m to exist. Such small values are consistent with those derived by Elliott and Oakey (1979) from temperature microstructure measurements in the tropical Atlantic. The higher rates of turbulent kinetic energy dissipation measured by Crawford and Osborne (1981) appear to be restricted to a narrow band near the Equator. However, recent experiments in the equatorial Pacific found no evidence for enhanced mixing (Gregg *et al.*, 1985 ; Moum, Caldwell, 1985).

CONCLUDING REMARKS

The focus here has been primarily on variation of chlorophyll and light absorption in the vertical. Variation in concentration of sea-surface chlorophyll with time (Dandonneau, Donguy, 1983) and in the horizontal as seen in the recent imagery near the Galapagos Islands (Feldman *et al.*, 1984), will influence the vertical distribution of radiation absorption in tropical regions. The critical non-dimensional number which will determine the radiation absorption in the upper mixed layer is the quantity $\int_0^l K_T dz$, where l is the depth of the mixed layer. Better measurement of this quantity is required in the future ; it is particularly important that the optical and mixed layer depth be measured simultaneously because of expected negative covariation in K_T and l. To the extent that such variations in surface chlorophyll (and consequently K_T) increase or diminish the flux of heat out of the upper mixed layer, they will be important considerations for tropical, and perhaps global, heat budgets.

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