

## Phytoplankton and Thermal Structure in the Upper Ocean: Consequences of Nonuniformity in Chlorophyll Profile

MARLON R. LEWIS,<sup>1,2</sup> JOHN J. CULLEN,<sup>2,3</sup> AND TREVOR PLATT<sup>2</sup>

Solar irradiance absorption in the upper layers of the ocean is influenced by the concentration of phytoplankton. Nonuniformities in vertical chlorophyll (phytoplankton) distributions lead to nonuniformities in absorption and consequently to variations in local heating. Local heating rate tends to decrease monotonically with increasing depth, but certain vertical distributions of chlorophyll lead to a change of sign of the gradient. Data from an oligotrophic ocean station showed a subsurface maximum of chlorophyll intense enough to allow increased heating on the upper slope of the chlorophyll maximum, with a maximum gradient of  $0.003^{\circ}\text{C d}^{-1} \text{ m}^{-1}$ . It is suggested that the importance of such subsurface heating for augmenting vertical mixing and influencing mixed layer deepening may be greatest in the optically clear, open ocean regions.

### INTRODUCTION

The vertical gradient of downwelling radiation is an important determinant of upper ocean structure. Models of thermal structure in the upper ocean are very sensitive to the way the gradient is represented in terms of the components of optical attenuation [e.g., *Simpson and Dickey*, 1981a, b] and to the magnitude of the attenuation coefficients themselves [e.g., *Kraus and Rooth*, 1961; *Denman*, 1973].

In the open ocean, where concentrations of suspended sediments and absorptive dissolved materials are small, an important variable affecting the penetration of radiant energy is the concentration of phytoplankton pigments [e.g., *Jerlov*, 1976]. Concentrations of phytoplankton are variable over many scales, in both the horizontal [*Platt*, 1975] and in the vertical planes [*Cullen*, 1982]. It follows that the attenuation of irradiance and therefore the local heating may vary on these same scales.

Vertical fluctuations in heating rate caused by vertical structure in the phytoplankton pigments are a potential source of thermal instability in the upper ocean. A ubiquitous feature of highly resolved vertical profiles of chlorophyll fluorescence [*Lorenzen*, 1966] taken in many parts of the world is the presence of a strong sub-surface maximum [*Cullen*, 1982]. Although this maximum of fluorescence may not necessarily represent a maximum of phytoplankton cells, it generally does represent a maximum of chlorophyll or chlorophyll-like pigments. In this paper, we address the question 'Under what circumstances can this sub-surface chlorophyll maximum induce an inversion in heating rate leading to a density gradient that is inherently unstable?'

### FORMULATION OF THE PROBLEM

The conservation of radiant energy is generally stated as [*Gershun*, 1939; *Preisendorfer*, 1976],

$$\nabla \cdot \mathbf{E} = -a\dot{E} \quad (1)$$

where  $\mathbf{E}$  is the vector irradiance,  $a$  is the absorption coefficient, and  $\dot{E}$  is the scalar irradiance. We can assume the horizontal divergence to be negligibly small, and the upward contribution to be a small fraction of the vertical component of the irradiance vector [*Jerlov*, 1976]. Therefore the absorbed energy is closely approximated by the derivative of the downwelling irradiance with respect to depth. The radiant energy absorbed at a particular depth  $z$  (positive downward) gives rise to a local increase in heat content, hence temperature, according to the following [*McEwen*, 1929; *Ivanoff*, 1977; *Simpson and Dickey*, 1981a],

$$\frac{\partial \theta(z, t)}{\partial t} = \frac{-1}{c_p \rho} \frac{\partial E_D(z)}{\partial z} = \frac{-E_D(0)}{c_p \rho} \frac{\partial F(z)}{\partial z} \exp [F(z)] \quad (2)$$

Here,  $E_D$  is the magnitude of the downwelling irradiance,  $c_p$  is the thermal capacity of water,  $\rho$  is the density, and  $F(z)$  is the integral from the surface to  $z$  of the attenuation coefficient for downward irradiance. The depth  $z = 0$  refers to just below the sea-surface.

In the vicinity of the chlorophyll maximum, most of the energy will be in the spectral bandwidth  $\approx 400$  to  $700$  nm (photosynthetically available radiation (PAR)). Higher and lower bandwidths are absorbed strongly in the upper meter of the ocean [*Jerlov*, 1976].

We now ask under what conditions the following inequality holds:

$$\frac{\partial^2 \theta(z, t)}{\partial z \partial t} > 0 \quad (3)$$

This is the necessary (but not sufficient) condition for the development of instability.

Differentiating (2) with respect to  $z$ ,

$$\frac{\partial^2 \theta(z, t)}{\partial z \partial t} = \frac{-E_D(0)}{c_p \rho} \left[ \frac{\partial^2 F(z)}{\partial z^2} + \left( \frac{\partial F(z)}{\partial z} \right)^2 \right] \exp [F(z)] \quad (4)$$

Therefore, in order to satisfy (3) the terms enclosed by the bracket in (4) must be less than zero. Let us denote the value of this bracket by the symbol  $G(z)$ .

*Functional form of  $F(z)$ .* An expression for  $F(z)$  based on diffuse attenuation for the downward irradiance which includes the influence of the chlorophyll concentration [ $B(z)$ ;  $\text{mgChl a m}^{-3}$ ] is [*Platt*, 1969; *Lorenzen*, 1972; *Smith and Baker*, 1978],

$$F(z) = - \left( K_w z + K_c \int_0^z B(z) dz \right) \quad (5)$$

<sup>1</sup> Dalhousie University, Department of Biology, Halifax, Nova Scotia.

<sup>2</sup> Marine Ecology Laboratory, Bedford Institute of Oceanography, Dartmouth, Nova Scotia.

<sup>3</sup> Now at University of Texas, Port Aransas Marine Laboratory, Port Aransas, Texas 78373.

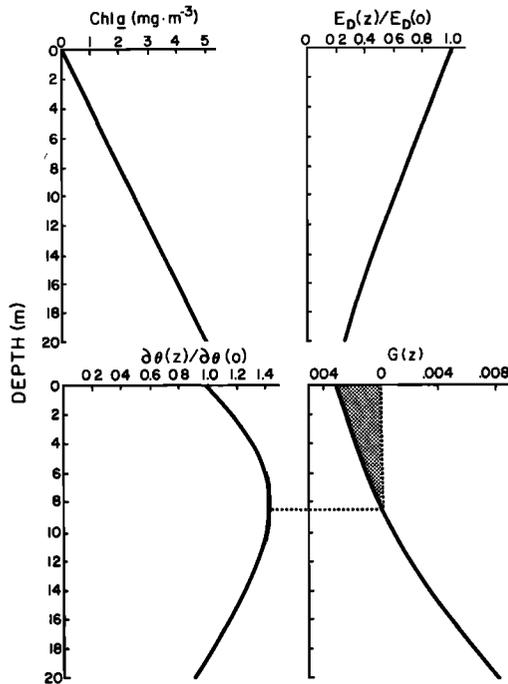


Fig. 1. Hypothetical depth distributions of chlorophyll, irradiance, heating rate, and the function  $G(z)$  (equation (6)) for the case when chlorophyll exhibits a linear increase with depth with slope  $m$ . Parameter values here are  $m = 0.25$  ( $\text{mg chl } a \text{ m}^{-3}$ )  $\text{m}^{-1}$ ,  $K_w = 0.03$   $\text{m}^{-1}$ ,  $K_c = 0.016$  ( $\text{mg chl } a$ )  $\text{m}^{-2}$ . The shaded region indicates where  $G(z) < 0$  and  $\partial^2\theta/\partial z\partial t > 0$ .

$K_w$  is the (diffuse) attenuation coefficient for pure seawater plus the contributions from any other absorbing materials distributed uniformly with depth. The chlorophyll specific attenuation coefficient  $K_c$  in practice includes other materials (i.e., 'yellow substance' *Bricaud et al.* [1981]) whose abundance is correlated positively with chlorophyll concentration.

Substituting (5) into the equation for  $G(z)$  we find (subject to  $K_c \int_0^z B(z) dz$  being continuous and differentiable),

$$G(z) = -K_c \frac{\partial B(z)}{\partial z} + [-K_w - K_c B(z)]^2 \quad (6)$$

For a given  $B(z)$  then, we look for conditions under which  $G(z)$  is less than or equal to zero, conditions which provide presumptive evidence for the production of respectively unstable or neutrally stable density distributions.

**Chlorophyll distributions and their consequences for heating.** We can dismiss immediately the trival condition of  $B(z) = \text{constant}$ . In this case  $G(z) = [-K_w - K_c B(z)]^2$  and a stable density field will always result.

As a first approximation to actual chlorophyll distributions, we might take  $B(z)$  to be 0 at  $z = 0$  and to increase linearly with depth. In this case,  $B(z) = mz$ , and

$$G(z) = -K_c m + (-K_w - K_c m z)^2 \quad (7)$$

The function  $G(z)$  crosses zero at depths corresponding to the roots of (7) and is negative between the roots, since we have taken  $m$ ,  $K_c$ , and  $K_w$  to be positive. The roots are

$$z = \frac{-K_w}{K_c m} \pm \frac{1}{(K_c m)^{1/2}} \quad (8)$$

Since the sea surface is at  $z = 0$ , only one of the roots

potentially has physical meaning. If the root is to be positive, then  $(K_c m)^{1/2}/K_w > 1$ . This non-dimensional number can be interpreted as a ratio of the optical length scale of pure seawater to the optical length scale associated with the chlorophyll distribution. If the ratio is above 1, there is the potential for the evolution of instability from the surface to a depth  $z = (-K_w/K_c m) + (1/(K_c m)^{1/2})$ . Typical values for  $K_c$  and  $K_w$  are  $0.016$  ( $\text{mg chl } a \text{ m}^{-3}$ )  $\text{m}^{-2}$  [*Smith and Baker, 1978*] and  $0.03$   $\text{m}^{-1}$  [*Smith and Baker, 1981*], respectively, and therefore the gradient in chlorophyll  $m$  under this approximation must be greater than  $\approx 0.056$   $\text{mg chl } a \text{ m}^{-4}$  for instability to be plausible. A hypothetical example with  $m = 0.25$   $\text{mg chl } a \text{ m}^{-4}$  is given in Figure 1.

Although this simple example is useful to clarify the problem, a linear model of the chlorophyll distribution is not too realistic (see Figure 4). A Gaussian distribution captures more accurately the major features of observed vertical profiles.

$$B(z) = \frac{\bar{B}}{(2\pi)^{1/2}\sigma} \exp\left(-\frac{1}{2}\left[\frac{z-l}{\sigma}\right]^2\right) \quad (9)$$

Here  $\bar{B}$  is the integrated chlorophyll ( $\text{mg chl } a \text{ m}^{-2}$ ), over the entire water column,  $l$  is the depth of maximum chlorophyll, and  $\sigma$  is a measure of the 'width' of the distribution (Figure 2). With this approximation then,

$$G(z) = \frac{K_c \bar{B} (l-z)}{(2\pi)^{1/2} \sigma^3} \exp\left(-\frac{1}{2}\left[\frac{z-l}{\sigma}\right]^2\right) + \left(-K_w - \frac{K_c \bar{B}}{(2\pi)^{1/2} \sigma} \exp\left(-\frac{1}{2}\left[\frac{z-l}{\sigma}\right]^2\right)\right)^2 \quad (10)$$

Substituting  $z^* = (z-l)/\sigma$ , expanding the squared term, and rearranging,

$$G(z) = (2K_w \sigma + z^*) \frac{(2\pi)^{1/2}}{K_c \bar{B}} \exp\left(\frac{-z^{*2}}{2}\right) + \frac{2\pi(K_w \sigma)^2}{(K_c \bar{B})^2} + \exp(-z^{*2}) \quad (11)$$

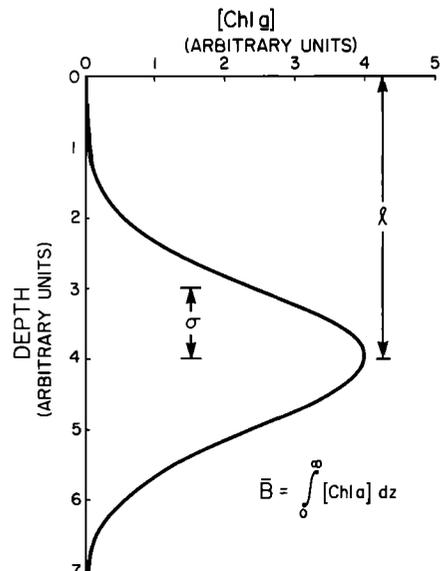


Fig. 2. Schematic diagram of the distribution of chlorophyll with depth, using the Gaussian distribution as the model.

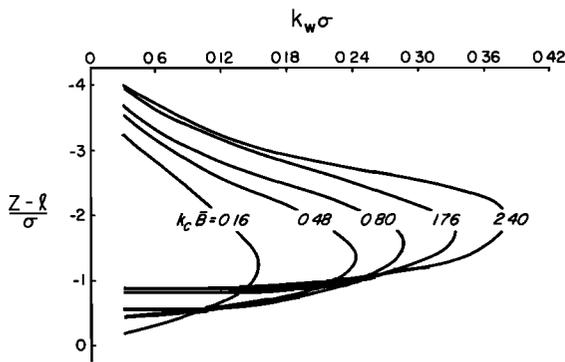


Fig. 3. Numerical solutions to the roots of equation (11) as a function of the nondimensional groups  $(z-l)/\sigma$ ,  $\sigma K_w$ , and  $B\bar{K}_c$ .

where  $G(z)$  is now dependent on the three nondimensional numbers  $z^*$ ,  $K_w\sigma$ , and  $K_c\bar{B}$ . As before, we seek the two roots of  $G(z)$  which, if they are real, define a depth interval in which instability could occur. The roots of (11) were obtained numerically and plotted as a function of the three dimensionless groups (Figure 3). Envelopes enclosing the critical region for various values of  $K_c\bar{B}$  are defined by isopleths. A vertical line from a given value of  $K_w\sigma$  intersects the isopleths at the nondimensional depth where neutral stability results, enclosing regions where an unstable density distribution could evolve. All of the parameters are accessible to field observation.

#### DATA

Three examples of vertical sections which exhibited maxima of chlorophyll at depth were chosen for further analysis. The first was taken inshore off Southern California (station 3A, Cullen [1980]), the second on the continental shelf of Nova Scotia and the third from an open ocean station near the Azores (Figure 4). The fluorescence data in the Southern California study were collected by pumping water from a submersible pump through a shipboard fluorometer. More details can be found elsewhere [Cullen, 1980]. The profile from the Scotian Shelf was defined from an in situ fluorometer on the Batfish towed vehicle [Dessureault, 1976; Herman *et al.*, 1981], and that from the Azores was determined also

from an in situ fluorometer [Fasham *et al.*, 1982]. In all cases, fluorometer output was calibrated by regression analysis on extracted chlorophyll. Temperature, irradiance (PAR), and salinity distributions with depth were also available. For each profile, values of the chlorophyll specific attenuation coefficient  $K_c$  and the uniform attenuation coefficient  $K_w$  were found by regression analysis [e.g., Smith and Baker, 1978] using equation (5) as the model. For the simulations,  $E_D(0)$  was taken as  $120 \text{ J m}^{-2} \text{ s}^{-1}$ , the thermal capacity  $c_p$  as  $4167 \text{ J Kg}^{-1} \text{ C}^\circ^{-1}$  and  $\rho$  as  $1024 \text{ kg m}^{-3}$ .

#### RESULTS AND DISCUSSION

All three profiles show a clear maximum of chlorophyll at depth (Figure 4), although depth  $l$  and the intensity or width of the maxima  $\sigma$  vary. The peak concentrations fall at depths where the irradiance is between 10% and 1% of the surface irradiance as found in a more exhaustive survey of deep chlorophyll maxima [Cullen and Eppley, 1981].

Regression analysis gave values of the chlorophyll specific attenuation coefficient  $K_c$  of 0.05, 0.12, and 0.11 ( $\text{mg chl m}^{-3}$ ) $^{-1} \text{ m}^{-1}$  and values for the uniform attenuation coefficient  $K_w$  of 0.1, 0.07 and  $0.036 \text{ m}^{-1}$  for the Southern California, Scotian Shelf, and Azores stations, respectively (Table 1). The parameter values for the Gaussian function, chosen by inspection of the chlorophyll profiles, are given in Table 1.

Of the three stations studied only the oligotrophic open-ocean station met the requirement of  $G(z) < 0$ . The combination of optically clear water and an intense chlorophyll maximum layer relatively near the surface were sufficient to allow an increased heating rate at depth. It is perhaps surprising that such a combination is found at all; by definition it is an intrinsically unstable distribution and it may not be a typical example (M. Fasham, personal communication, 1982). The region of maximum heating does not correspond to the maximum chlorophyll, but is on its leading edge (shallow side). This is also seen in the nondimensional representation (Figure 3) where the critical depths fall shallower than the center of chlorophyll mass, indicating that conditions for instability develop where optical attenuation increases with depth at a greater rate than light decreases. A similar result holds for the depth of maximum photosynthe-

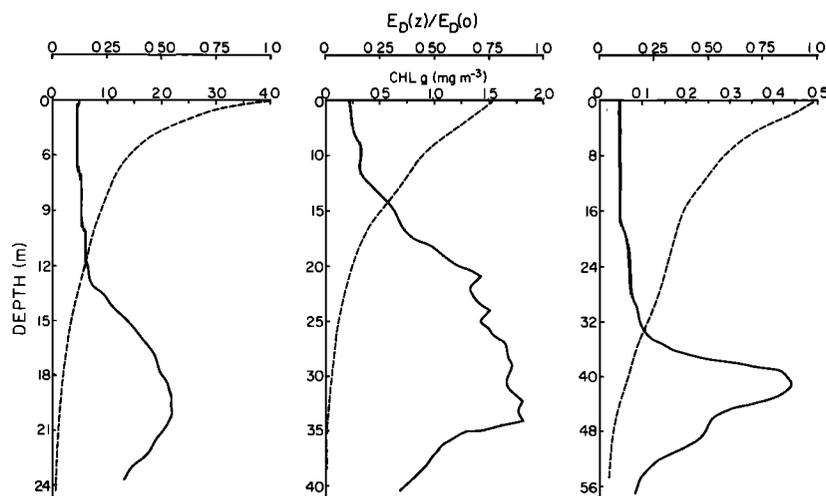


Fig. 4. Vertical profiles of chlorophyll and irradiance. Profile A is from coastal Southern California, B is from the Scotian Shelf, and C is from the tropical Azores.

TABLE 1. Data and Parameter Values Used in Computations of Heating Rate Profiles (Figure 5)

	Station		
	Southern California	Scotian Shelf	Azores
Latitude	33°10'N	42°N	33°34'N
Longitude	117°20'W	63°25'W	33°28'W
$K_c$ ,* (mg Chl m <sup>-3</sup> ) <sup>-1</sup> m <sup>-1</sup>	0.053 ± 0.008	0.12	0.114 ± 0.023
$K_w$ ,* m <sup>-1</sup>	0.105 ± 0.011	0.072	0.036 ± 0.005
$\bar{B}$ , mg Chl m <sup>-2</sup>	28.07	50.0	4.5
$l$ , m	19.75	32	41
$\sigma$ , m	5	11	4
$E_D(0)$ , J m <sup>-2</sup> s <sup>-1</sup>	120	120	120
$c_p$ , J Kg <sup>-1</sup> C <sup>o</sup> -1	4167	4167	4167
$\rho$ , kg m <sup>-3</sup>	1024	1024	1024

\*Regression coefficient ± Std error of the estimate.

sis associated with chlorophyll maximum layers [Platt and Herman, 1983; Herman and Platt, 1983]. Here the maximum storage of radiant energy falls in the region of increasing biomass, for the same reasons.

The simulated (and measured) irradiance profiles all show increased attenuation in the upper region of the chlorophyll maximum, even though the increased heating rate is only observable as a flattening of the heating rate profile for the Southern California and Scotian Shelf stations. The ubiquity of such features in irradiance profiles is evident in the data of Spitzer and Wernand [1981] where the depth of the sharp changes in attenuation correspond closely to the depth of maximum chlorophyll for many stations in the equatorial Atlantic.

#### Sources of Error

Several sources of error in the preceding calculations should be mentioned. To the extent that the upwelling component represents a significant fraction of the net downward irradiance, our estimates of the rate of storage (and consequently, the increase in heating rate) will be overestimated [see Morel and Bricaud, 1981, Appendix A]. This fraction is of the order 1–2% [Morel and Prieur, 1977]. In addition, as pointed out by Morel and Bricaud [1981], the coefficient of downward irradiance attenuation, a parameter which is an apparent property in the sense of Preisendorfer [1961], does not rigorously lend itself to additive partition as we have done in equation (5). Nevertheless, all regressions of attenuation on chlorophyll were highly significant ( $p < 0.001$ ), even though the chlorophyll specific attenuation coefficient for both the Scotian Shelf and Azores stations are higher than the 'average' value of 0.016 (mg chl)<sup>-1</sup> m<sup>-2</sup> [Bannister, 1974]. Concentrations of absorbing (and scattering) particulate and 'dissolved' matter other than chlorophyll (i.e., detritus) can be positively correlated with chlorophyll concentrations leading to the higher values observed [Spitzer et al., 1982] and such associations may be particularly prevalent in the open, optically clear portions of the oceans (A. Morel, personal communication, 1982). Much of this may, in fact, be due to chlorophyll associated with cells smaller than recovered on the filters typically employed for chlorophyll analysis. The evidence for the existence of such a small size fraction is increasing [Waterbury et al., 1979; Sieburth, 1979; Platt et al., 1983; Li et al., 1983] and may be responsible for reports of 'soluble' fluorescence [Herbland and LeBoutieller, 1981]. The effective optical cross sections

of cells increases as the size of the particle decreases if intracellular pigment concentrations remain constant [Morel and Bricaud, 1981] leading to higher values of the optical attenuation coefficient.

The coefficients  $K_c$  and  $K_w$  are strongly wavelength dependent, a feature we have suppressed, again for lack of information. In this analysis, only data below 5–10 m were used in regressions. Below this depth, the ratio of total energy to total quanta is relatively constant [Morel and Smith, 1974] and the error introduced is relatively small.

Our estimates of heating rates are also overestimated due to the neglect of the small 'background' chlorophyll concentrations near the surface. This can be dealt with as a linear increase in  $K_w$ : the main conclusion will be unaffected.

Some of the absorbed irradiance, rather than being directly converted into heat as we have assumed, is of course, stored as photosynthate by the algae. The storage is a relatively small fraction of the total absorbed energy due to the inevitable thermodynamic inefficiencies associated with the transformation [Morel, 1978; Platt et al., 1983]. Of the photosynthate stored, most is respired locally with consequent recovery as heat. The fraction exported as fecal pellets, etc., represents such a small fraction of the total storage in the open ocean [Platt et al., 1983], that it is negligible.

#### Nonuniform Heating and Vertical Mixing

Having established the conditions under which non-uniform distributions of chlorophyll can give rise to positive gradients in the heating rate, as in the Azores station, it remains to evaluate the importance of the process for augmenting vertical mixing. To do this, it is necessary to compare the rate of local heating (which may lead to a statically unstable density structure) to the rate at which temperature deviations would be dissipated by water motion. A first-order approximation to the Eulerian change in temperature with time is

$$\frac{\partial \theta}{\partial t} = \frac{E_D(0)}{c_p \rho} \frac{\partial F(z)}{\partial z} \exp[F(z)] + K_v \frac{\partial^2 \theta}{\partial z^2} \quad (12)$$

where the dissipative process is represented by a Fickian diffusion term with a constant coefficient of vertical eddy diffusivity for temperature,  $K_v$ . The parameterization of the turbulence is physically unrealistic, but it does allow us to compare the relative magnitudes of the two processes to determine which process will dominate.

The vertical temperature distribution that results from conditions that admit  $G(z) < 0$  is, if  $K_v$  is set to zero, a layer of warm water sandwiched between colder layers. A criterion which has proven useful in related studies [e.g., Turner, 1965], is the non-dimensional ratio of the contributions to the density gradient from temperature and salinity gradients,

$$\frac{-\alpha(\partial \theta / \partial z)}{\beta(\partial s / \partial z)} \quad (13)$$

where  $\alpha$  is the partial derivative of density with respect to temperature,  $\beta$  is the same derivative but with respect to salinity, and  $\partial \theta / \partial z$  and  $\partial s / \partial z$  indicate the temperature and salinity gradients, respectively. When the ratio exceeds values of about 0.5, enhanced (over molecular) vertical exchange of fluid can occur, initially through double-diffusive processes and then as the ratio exceeds one, through

convective overturn [Turner, 1973]. For given initial values of  $\partial\theta/\partial z$  and  $\partial s/\partial z$ , and a given rate of change of the temperature gradient (equation (3)), we can determine the time required for the ratio of the density contributions to attain the critical value of 0.5. This time scale can then be compared to the dissipative time scale. For this, we take a length scale equal to the standard deviation of the chlorophyll distribution  $\sigma$  and the time scale for variations in heating to be dissipated by eddy diffusion is then  $\approx \sigma^2/K_v$ .

The derivative of the computed heating rate with respect to depth in the Azores profile is, at its maximum,  $\approx 0.003^\circ\text{C m}^{-1} \text{d}^{-1}$  and it would take  $\approx 18$  days to attain the critical value of  $-\alpha(\partial\theta/\partial z)/\beta(\partial s/\partial z) = 0.5$ . Since the length scale  $\sigma$  is  $\approx 4$  m, it would require an eddy diffusivity of  $\approx 10^{-5} \text{m}^2 \text{s}^{-1}$  to dissipate the temperature deviations. A typical value of  $K_v$  for this layer is  $\approx 10^{-5} - 10^{-4} \text{m}^2 \text{s}^{-1}$ ; it thus seems unlikely that the absorbance of irradiance by the surface maximum of chlorophyll would lead to convection at this station.

However, much more pronounced subsurface maxima have been observed at similar or greater irradiance levels [Heaney and Talling, 1980; Cullen, 1982] increasing the possibility of algal-mediated convective processes. Stavn [1982] has discussed the trapping of energy in 'nepheloid layers' due to multiple scattering and absorption and cites Schindler *et al.*'s [1981] reported differential heating rates of a chlorophyll maximum layer in a lake of  $\approx 0.5^\circ\text{C h}^{-1}$ . Diurnal heating of the chlorophyll maximum resulted in convection, which dissipated structure in chlorophyll (which reformed by morning) and led to a deepening of the mixed layer. Even if the differential heating rate is considerably milder, the input of heat at depth may enhance the rate of deepening of the seasonal thermocline, particularly in the early spring when the upper layers are relatively isothermal and sunlight becomes sufficient to establish algal populations. The rate of deepening of the mixed layer is inversely proportional to Väisälä frequency  $N^2$  [Pollard *et al.*, 1973] at the thermocline. Decreases in  $N^2$  due to increased heating could potentially lead to increases in the rate of deepening of the mixed layer. Based on the magnitudes of the heating rates computed here, such an effect would be effective at the seasonal time scale.

In laboratory experiments where a solution stably stratified with respect to salinity is heated from below [Turner and Stommel, 1964], the qualitative result is first the formation of a layer of well-mixed water of small vertical extent just over the heat source. Rather than increasing the vertical scale of this layer, with time more layers formed successively on top of this resulting in a step-wise temperature (and salinity) structure, quite similar to that observed with micro-scale temperature profiles in the ocean [e.g., Neal *et al.*, 1969]. Such a stepped distribution of chlorophyll has also been reported [Derenbach *et al.*, 1979] and is somewhat evident in the Scotian Shelf data presented here. Similar structures have often been attributed to horizontal intrusions or cooling at the surface resulting in double-diffusion; the same result would occur with an increased heating rate associated with nonuniform vertical distributions of chlorophyll.

The stabilizing influence on the water column of uniformly distributed algal populations has been discussed [Denman, 1973; Zaneveld *et al.*, 1981], and in this regard it is interesting to look at depths below that of maximum differential heating. The removal of such a large portion of the available energy, coupled with the decreasing (but still high) chloro-

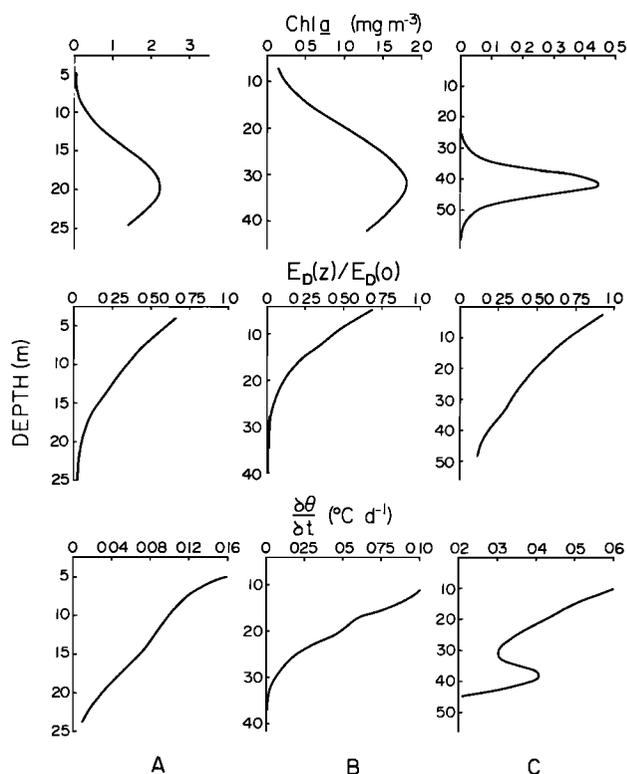


Fig. 5. Computed values of the depth distribution of chlorophyll, irradiance, and heating rates based on the data in Figure 4. The letters refer to the same profiles as in Figure 4.

phyll concentrations, gives rise to a steep decline in the rate of heating with depth in this region (Figure 5). The depth strata containing the deeper portion of the chlorophyll maximum would be increasingly more statically stable with time. The increased stability may permit the formation and maintenance of the chlorophyll maximum, which would gradually decrease in depth as the upper surfaces were eroded. It may be that the commonly observed association of chlorophyll maxima with high density gradients may be more the influence of algal absorption on the density structure than decreased sinking rates of algae in these depth strata, as usually implied [Cullen, 1982].

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