

New production and mixed-layer physics

SHUBHA SATHYENDRANATH^{1,2} and TREVOR PLATT²

¹Department of Oceanography, Dalhousie University, Halifax, Nova Scotia, Canada B3H 4J1.

²Biological Oceanography Division, Bedford Institute of Oceanography, Box 1006, Dartmouth, Nova Scotia, Canada B2Y 4A2.

Abstract. We treat the ocean carbon cycle as a coupled physical-biogeochemical process. The interactions between mixed-layer dynamics and growth of phytoplankton in the layer are discussed, and the formal relationship between phytoplankton accumulation and new production is examined. A coupled biological-physical model is presented for studying the classical spring bloom in the N. Atlantic, and possible differences in the mechanisms that drive the seasonal phytoplankton blooms in the N. Atlantic and the Arabian Sea are discussed. Finally, recommendations are made for observational programs to improve our understanding of the biologically-mediated carbon cycle in the Arabian Sea.

Keywords. Arabian Sea; new production; phytoplankton; primary production; mixed-layer dynamics; ocean carbon cycle; marine optics.

1. Introduction

In the ocean, physical, chemical and biological processes are coupled in an intimate manner. That is to say, the biological and chemical fluxes are, to a considerable extent, under physical control. But it also implies the possibility, at least, that the physical dynamics may unfold in a way that depends upon the local fields of biological and chemical properties. The ocean carbon cycle is the epitome of a coupled, physical-biogeochemical process. Although some workers would insist on isolating parts of the coupled system to emphasise their perceived importance, a growing majority would hold to the view that the best chance to deepen our understanding of the ocean carbon cycle is to maintain its interactive quality continuously in the foreground.

Evidence for the influence of biological properties on physical dynamics in the upper ocean has been published (Lewis *et al* 1983; Simonot *et al* 1988; Sathyendranath *et al* 1991): the presence of phytoplankton modifies the way in which the ocean absorbs solar radiation and thus affects the thickness and temperature of the surface mixed layer. To the extent that the biological contribution is seen only as a static addition to the heating rate, it may be treated in a purely physical manner. However, if the phytoplankton is allowed to grow, at a rate that depends on the depth and optical transparency of the mixed layer, the problem becomes a coupled, physical-biological one.

Simonot *et al* (1988) have studied the coupled problem over a two-year cycle at Ocean Weather Station R in the N. E. Atlantic, and shown that phytoplankton have a significant effect on sea surface temperature (SST) prediction. The coupled problem has also been treated, in a preliminary way (Platt *et al* 1994), in the context of the development of the spring bloom for temperate to high latitudes in the Atlantic Ocean.

This study showed that the presence and growth of phytoplankton in the Spring has a considerable influence on the evolution of the mixed-layer dynamics. In the Arabian Sea, however, the treatment has so far been only physical (Sathyendranath *et al* 1991). But we know that the Arabian Sea is subject to extreme variations, with respect to season, in both physical forcing and in concentration of phytoplankton.

In this paper we examine how the coupling between physical and biological processes can be demonstrated, theoretically, in the Atlantic Ocean. We show also how this coupling can be described in the language of the oceanic carbon-nitrogen cycles associated with the biological pump. Finally, we explore whether, given what is known about the oceanography of the Arabian Sea, the same theory could be applied there.

2. Physical models for the mixed layer

The depth of the mixed layer is set by the balance of processes that tend to stabilise, and destabilise, the density structure of the upper ocean. Solar heating tends to stabilise the density profile; wind-generated turbulence tends to erode it. An important parameter in the models that describe this balance is the optical attenuation coefficient, K [L^{-1}]. This quantity has been variously regarded as infinitely large, such that all the solar input is absorbed at the surface (McCreary and Kundu 1989); to have a value that is equal to the large-scale climatic average for the ocean (Niiler and Kraus 1977); or to have a value assigned according to the Jerlov water type for the area of the ocean considered (Gaspar 1988).

None of these is really satisfactory, especially for areas that experience large changes in standing stock of phytoplankton with season. This is because the photosynthetic pigments contained in phytoplankton exert a strong influence on the optical attenuation coefficient. Hence in areas subject to significant changes in phytoplankton abundance, no single optical attenuation coefficient will suffice for the entire year.

Thus, in the Arabian Sea, it was shown (Sathyendranath *et al* 1991) that the phytoplanktonic component of optical attenuation was variable, and could, on occasion, account for a contribution to the mixed-layer heating rate of more than three degrees Celsius per month. This conclusion was reached using synoptic, serial information on the phytoplankton abundance obtained by remote sensing of ocean colour.

The incentive exists, then, to improve the parameterisation of optical attenuation coefficient in mixed layer models by making explicit the contribution from phytoplankton. An analysis of the effect of phytoplankton on mixed-layer depth and temperature in a standard mixed-layer model has been made (Ravindran 1994). The effect is present at all times, but is strongest during periods of low winds or when the surface heat balance is dominated by the solar input.

This analysis is as far as one can go with a purely physical treatment. The next more complex step is to allow the phytoplankton concentration to wax and wane, in accordance with observation. That is to say, the analysis should be a coupled, physical-biological one, and this will be the subject of a later section.

3. Phytoplankton growth in the mixed layer

The mixed layer depth is not a static quantity: at any time it is either shallowing or deepening. Similarly, the phytoplankton biomass is not a static quantity. It is set by

a dynamic balance between growth and loss terms that, in general, have non-zero resultant. The prerequisite for a coupled, physical-biological model of the mixed layer is to describe how the biomass of the mixed layer changes under these competing growth and loss processes.

The classical approach to phytoplankton growth in the mixed layer is the theory of bloom initiation (Sverdrup 1953). Since primary production is a function of available light, the Sverdrup model treats phytoplankton growth as a function that decreases with depth. The loss terms, on the other hand, are treated as depth-independent, since it is assumed that losses are a function of biomass, which is constantly re-distributed uniformly within the upper layer due to active mixing. The Sverdrup theory treats the integrals of the growth term and the loss term from the surface to the mixed-layer depth. At the heart of the theory is the definition of the critical mixing depth: it is that mixed-layer depth for which the two integrals equal, such that net growth in the layer is zero. The condition for growth in the mixed layer can then be stated in terms of the critical depth: if the actual mixed-layer depth is less than the critical mixed-layer depth, then the net phytoplankton growth is positive, and conditions are suitable for the initiation of a bloom.

It has been shown how the magnitude of the growth rate (rather than just its sign) can be calculated for the mixed layer (Platt *et al* 1991a). The growth rate r per day for phytoplankton in the mixed layer (Platt *et al* 1991a) is given by

$$r = \log_e \left(1 + \frac{2\eta}{\pi K Z_m} (1 - M) - \frac{24L^B}{\chi} \right). \quad (1)$$

In this equation, η is the dimensionless group $\eta = P_B^m D I_*^m / \chi$, where P_B^m is the assimilation number, D is the day length in hours, I_*^m is the noon irradiance normalised to the photoadaptation parameter I_k , and χ is the carbon-to-chlorophyll ratio. The notation L^B represents the generalized, biomass-specific, loss term and $M = e^{-KZ_m}$.

Observe that the growth rate is a function of the dimensionless product KZ_m , where Z_m is the mixed-layer depth. This is the key to the coupling between the physical and biological processes affecting Z_m . With its aid we can now proceed to the development of a coupled, physical-biological model for the evolution of mixed-layer depth.

Equation (1) is based on the simplest possible photosynthesis-light model, which assumes a linear photosynthetic response to total available light in the photosynthetically-active region (400–700 nm). It is not difficult to extend the model to include non-linear effects and spectral dependence, but the fundamental principles would remain the same. However, in the interest of clarity, we prefer to use the less complex model for our preliminary analysis.

4. The biological pump

We now turn to the implications of coupled models for the understanding of the biological pump. First, we emphasise that the generalised loss term is to be interpreted in the broadest biogeochemical sense to embrace any process by which the products of photosynthesis are remineralised or removed from the mixed layer (Platt *et al* 1991a). It includes the catabolism of the entire pelagic community, and other losses such as sinking. Note that, in the short term, the difference between the daily production

in the mixed layer and the daily loss there is an index of the new production in the mixed layer for that period, provided that the sinking rate, and the zooplankton growth rate, are both zero (recall that the sinking term is part of the new production, as is the zooplankton growth term). The generalised loss term itself (actually $24BL^B Z_m/\chi$), would then be a measure of the regenerated production in the layer, that is of the production required to sustain the community metabolism in the layer.

Consider the case where the biomass in the mixed layer is increasing. If the net growth is set as a proportion of the gross daily production, this proportion can be interpreted as a lower bound on the local f -ratio. We say lower bound because any sinking of phytoplankton out of the mixed layer or any growth of zooplankton would increase the f -ratio. We use the term 'local' f -ratio to distinguish it from the f -ratio that would be measured by a bulk-property technique with a characteristic time constant at the seasonal or annual scale (Platt *et al* 1989). The local f -ratio is the one that would be measured by an *in vitro* method such as the N^{15} -method. The seasonal-, or annual-, scale ratio may be estimated by averaging the products of the local f and the corresponding daily production over the period required.

One possible application of the coupled, physical-biological model described above is as a vehicle to estimate the seasonal-scale f -ratio. That is, we could vary the percentage of daily production that is assigned to the generalised loss term and examine the corresponding evolution of the biomass field. We could choose the value of the local f -ratio for which the simulated biomass field most closely resembles the observations. The seasonal-scale f -ratio would then be found by extrapolation as already described.

In making the calculations on the seasonal development of the biomass field, we have to ensure that the increase in phytoplankton is not unreasonably high relative to the initial nutrient content of the mixed layer. This is a straightforward computation given information on the nutrient fields.

During the declining phase of the bloom (losses exceed production), the same conceptual approach will apply, except that care should be taken to include in the new production the increase in biomass at higher trophic levels: new production is net community production in the broadest sense. Note that the local f -ratio cannot be negative. If, on a particular day the community respiration is greater than the primary production, it means that the f -ratio is zero. All of the production that day is being respired, in addition to some accumulated biomass that constitutes new production from an earlier time. The time scales have to be kept in mind.

5. A coupled model for the classical spring bloom

We illustrate the coupled, physical-biological approach to the mixed layer (Platt *et al* 1994) with an application for the temperate-subarctic regime of the North Atlantic Ocean (67°N 10°W).

Monthly, climatological averages of the net flux at the sea surface (algebraic sum of latent heat, sensible heat, long-wave back radiation and short-wave radiation), shortwave radiation and frictional wind velocity at the sea surface used in this calculation are from the Max Planck Institute compilation (Oberhuber 1988) based on the COADS data set. These monthly climatological values were fitted with polynomial functions to interpolate daily changes in the forcing fields (see figure 1).

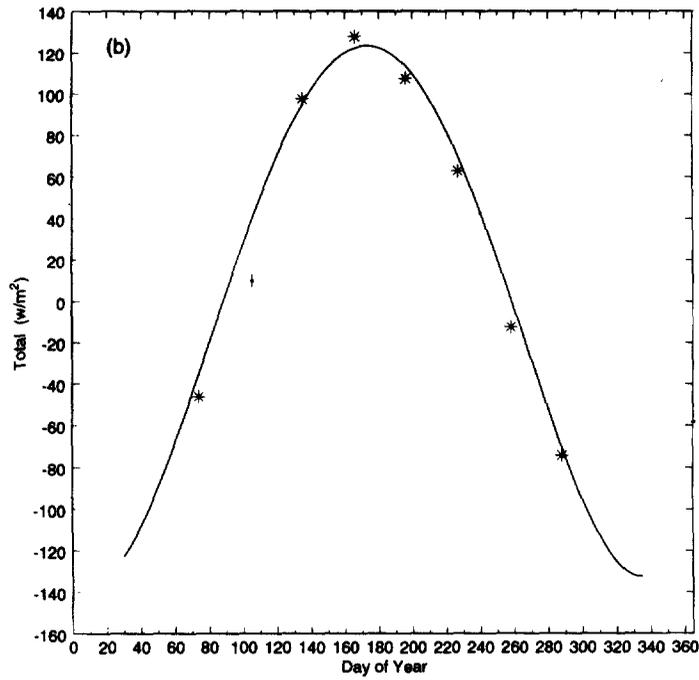
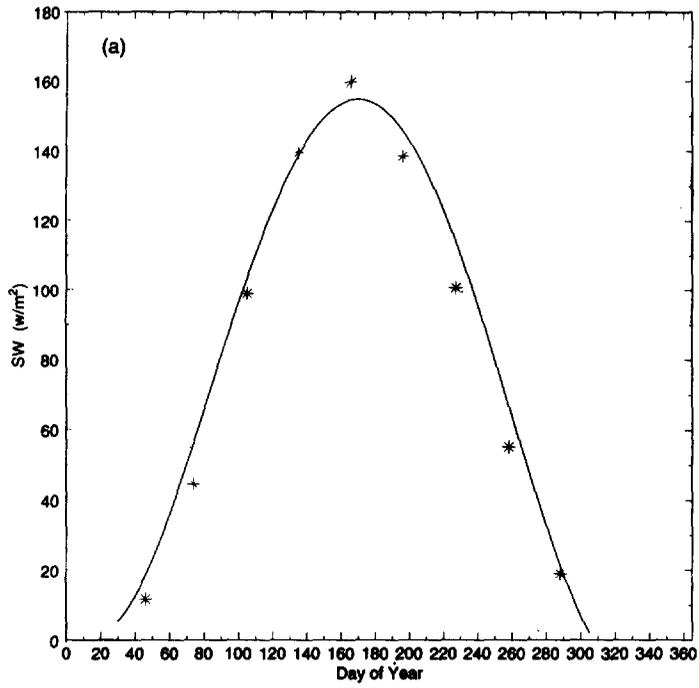


Figure 1.

Figure 1. (Continued).

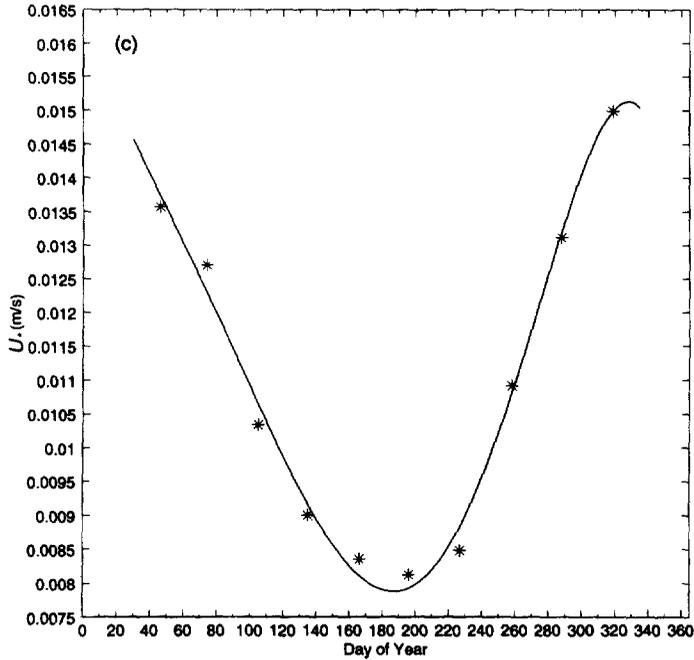


Figure 1. The climatological, monthly-averaged forcing fields from the Max Planck Institute compilation (Oberhuber 1988), for the location 67°N, 10°W, from February to November. The climatological values are indicated by asterisks, and the smooth curves are the fitted polynomial functions. (a) The short-wave radiation, (b) Total surface fluxes, and (c) Frictional velocity (u_*).

The phytoplankton field is allowed to evolve with time. The object of the calculation is to see how the mixed-layer depth at a given time will compare with that computed on the assumption that phytoplankton abundance does not change.

Rather than implement equation (1) itself, we used an equivalent expression for the fractional change Ξ in mixed-layer biomass during one day (refer the expression for the fractional growth ξ in equation (23) of Platt *et al* 1991a)

$$\Xi = 2P_m^B DI_*^m (1 - M) / (\pi K \chi Z_m) - 24L^B / \chi. \quad (2)$$

The motivation for using equation (2) rather than (1) was to allow the biomass and physical fields to change from day to day as the coupled system evolved.

We set the photosynthesis parameters, conservatively, at $P_m^B = 2 \text{ mg C (mg Chl)}^{-1} \text{ h}^{-1}$ and $\alpha^B = 0.05 \text{ mg C (mg Chl)}^{-1} (\text{W m}^{-2})^{-1} \text{ h}^{-1}$ (Platt *et al* 1991b). Recall that $I_k = P_m^B / \alpha^B$. The attenuation coefficient was estimated as a linear function of the phytoplankton biomass B as described in Platt *et al* (1991a). To obtain the noon irradiance we assumed that one half of the short-wave radiation represents photosynthetically-active radiation, and that it is distributed sinusoidally throughout the day (Platt and Sathyendranath 1993). The carbon-to-chlorophyll ratio was set at 30 (Platt *et al* 1991a). The generalised loss term was assigned as a percentage of the daily growth as described in the following section.

We used an initial biomass of $0.1 \text{ mg Chl m}^{-3}$ to represent the low winter biomass,

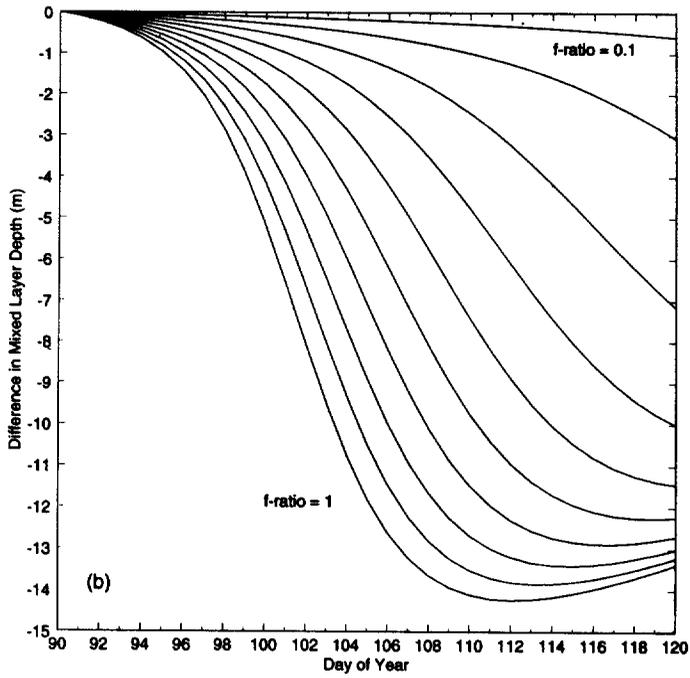
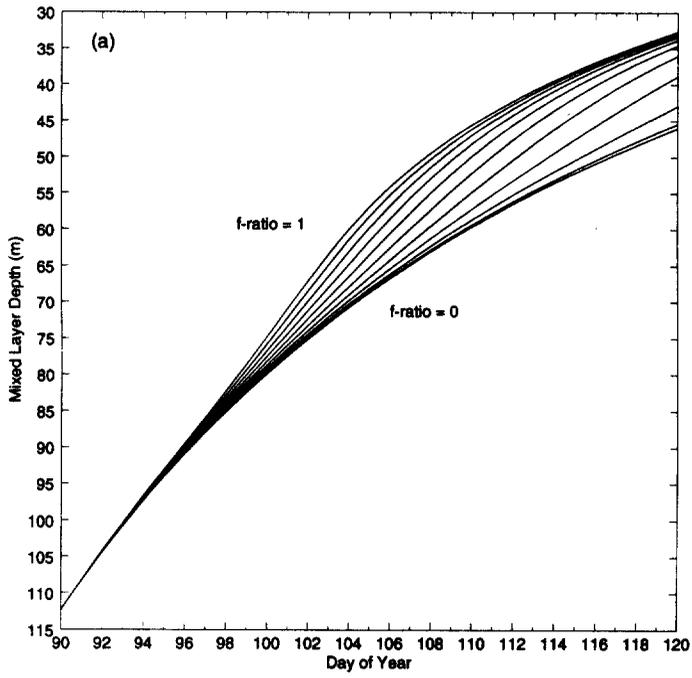


Figure 2.

Figure 2. (Continued).

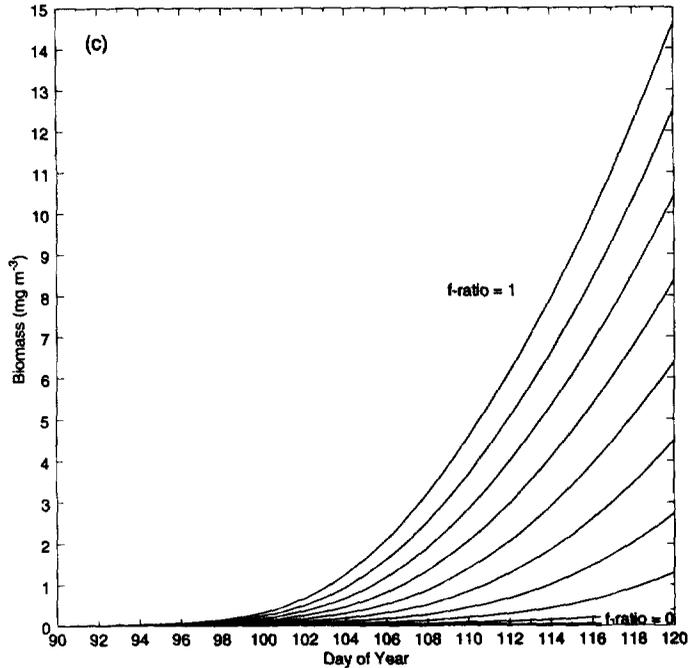


Figure 2. Model outputs, for one month of computation, for the station shown in figure 1. (a) Mixed-layer depth, for values of f -ratio going from 0 to 1, in increments of 0.1, as a function of day of year. (b) Difference in computed mixed-layer depth, with reference to the mixed-layer depth computed assuming no phytoplankton growth ($f = 0$), as a function of day of year. Different plots correspond to different values of f -ratio, from 0.1 to 1, in increments of 0.1. (c) Evolution of biomass with time, again for different values of f -ratio.

but we have verified that the choice could equally well have been ten times smaller without significant effect on the final results. We applied the Denman (1973) model for a shallowing mixed layer, as implemented by Shetye (1986), but with one notable exception: only the photosynthetically-active part of the radiation is considered to be penetrative. Given the high attenuation coefficient of water outside the visible range, the UV and near-infrared parts of the short-wave radiation are assumed to be absorbed at the surface. The model was applied day by day under two broad conditions: first that the biomass evolve based on equation (2), with the loss terms expressed as a fraction $(1 - f)$ of the growth term; and second that the biomass stay constant at its initial value. The computations were initiated when the total flux at the sea surface became positive, and were stopped after one month of simulated time.

The results we obtained are shown in figure 2, as a function of day of year for different values of the f -ratio. Recall that $f = 0$ means that all primary production is respired immediately, such that net growth is zero and the biomass of phytoplankton remains constant; $f = 1.0$ means that respiratory losses are zero and therefore all production is new production, leading to the maximum rate of biomass accumulation for the conditions.

The mixed-layer depth is affected in a significant way by the biomass accumulation (figure 2a). After some 10 days, it can differ by as much as 15 m (or about 25%) from

the mixed-layer depth expected if the biomass did not change (figure 2b). The biomass itself will have increased some 40 times during the same period with $f = 1$ (figure 2c).

6. The mixed layer of the Arabian Sea

The classical theory of Sverdrup seeks to explain the onset of spring blooms in high latitudes: winter mixing brings nutrients to the surface waters; in Spring, increased surface irradiance, stratification and shallowing of the mixed layer, all favour increased phytoplankton growth in the mixed layer. Thus, the spring bloom in the North Atlantic can be explained as a purely passive, biological manifestation of changing physical conditions. Although we recognise that the availability of nutrients may affect phytoplankton growth dynamics through changes in the parameters P_m^B and I_k , we could envisage the occurrence of a bloom even if such changes did not take place.

Before applying the Sverdrup model to the Arabian Sea, we should question whether the application is justified. The basic premise of the Sverdrup model, that the losses in the mixed layer should be less than the growth if there is to be an increase in

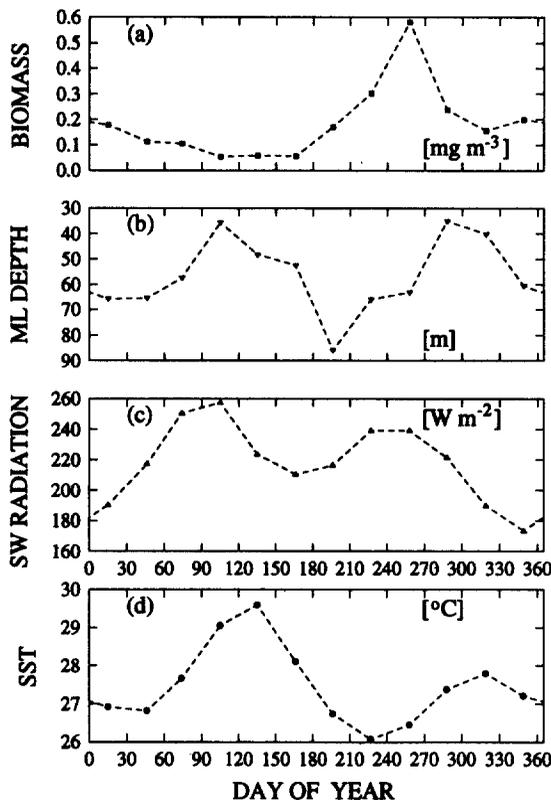


Figure 3. Monthly-averaged, climatological values at 9°N , 59°E in the Arabian Sea, of (a) Surface biomass, based on CZCS data (Feldman *et al* 1989), (b) Mixed-layer depth from Hastenrath and Greischar (1989), (c) Short-wave radiation from the Max Planck Institute compilation (Oberhuber 1988), and (d) Sea-surface temperature based on the COADS data set (McCreary and Kundu 1989).

biomass, is so fundamental that it always has to be true. But there are significant differences in the processes that promote the occurrence of a monsoonal bloom in the Arabian Sea, compared with the Spring bloom in the N. Atlantic.

By way of example, figure 3 shows the evolution with season (climatic-scale averages) in various oceanographic properties at 9°N 59°E in the Arabian Sea. Note that the shallowing and warming of the mixed layer from day of year 1 to 100 is accompanied not by a phytoplankton bloom, but by a progressive decline in the phytoplankton biomass.

Subsequently, the onset of the southwest monsoon leads to a steady deepening and cooling of the mixed layer. At about day 180, the surface shortwave radiation begins to increase. Around the same time, the biomass of phytoplankton also begins to increase. With a lag of roughly a month, the mixed layer begins to shallow, and continues to shallow and warm even after the phytoplankton biomass has started to decline.

Clearly, these two periods of minima in the mixed-layer depth follow different dynamics. Certainly, the earlier one does not follow the familiar pattern known for the temperate North Atlantic. The later one may follow this pattern, but more work needs to be done before alternative mechanisms can be ruled out. For example, given that advection is not important (see Brock *et al* 1991), the occurrence of a bloom during mixed-layer deepening could be explained if we assumed that, in response to the entrained nutrients in the mixed layer, the physiological parameters changed in such a way as to initiate a bloom. Note that this could be either an increase in P_m^B or a decrease in I_k , or both. It may be that the response of phytoplankton to nutrient availability (supplied as a consequence of mixing by the monsoon winds) is of more controlling importance in the Arabian Sea than in the N. Atlantic.

7. Discussion

In the Arabian Sea, the evolution of mixed-layer depth and temperature cannot be understood without taking into account the time variation in the phytoplankton biomass. With the new synoptic data on ocean colour expected from the SeaWiFs device (McClain *et al* 1992) it should be possible to build up archives on the seasonal variation in the biomass field, and to use instantaneous maps of the biomass field as an aid to better forecasting, in the short term, of sea surface temperature.

Not all the products of photosynthesis accumulate to modify the optical attenuation coefficient. A significant proportion is respired over a time period of hours to days. Hence the impact of primary production on the physical dynamics is modulated by the f -ratio, a macroscopic property of the pelagic ecosystem. In turn, the f -ratio is determined, in part, by the physical oceanography. If we consider that regenerated production is determined by processes internal to the mixed layer, whereas new production depends on the connection between the mixed layer and the layer below, the f -ratio may be interpreted as a measure of the coupling between the two layers, a function of physical factors such as wind stress.

For most oceanographic regions, we have reasonably good information on the evolution of the mixed-layer depth and of the surface biomass field. Of the factors that control the evolution of the biomass field, we know more about the parameters that control the growth, than about the rate constants that control the losses. This

is certainly true of the N. Atlantic. Coupled models of the type discussed here could be tuned to reproduce observed mixed-layer and biomass fields to obtain some understanding of the loss terms. But these computations would give only lower limits for the f -ratios. To improve our understanding of the f -ratio we require additional information on sinking rates and zooplankton growth rates.

These considerations point towards possible strategies for large-scale studies in the Arabian Sea designed for improving our understanding of the bio-geochemical cycles in this oceanic region: any such study would be incomplete if it did not pay due attention to the spatial distribution of parameters that determine phytoplankton growth and to their response to environmental factors, especially nutrient supply. These studies should also be designed to improve our understanding of the sinking rates of organic material leaving the mixed layer, and of the dynamics of zooplankton growth.

The approach we have followed here shows that, for a true insight into the upper ocean, we must approach it with the idea of a coupled system firmly in mind. This is no less true for a simple property like the surface temperature as it is for properties of the ocean carbon cycle.

Acknowledgements

The work presented in this paper was supported, in part, by the Office of Naval Research, the National Aeronautics and Space Administration, the European Space Agency, the Department of Fisheries and Oceans, Canada, and the Department of National Defence, Canada. Additional support was provided by the Natural Sciences and Engineering Research Council through Operating Grants to SS and TP. We thank Ernst Maier-Reimer and Ingo Jessel of the Max Planck Institute and Frank Muller-Karger and Frank Carprew of NASA for their help with getting access to some of the data used in this paper. We thank Osvaldo Ulloa and two anonymous referees for their encouraging comments on this manuscript, and Heidi Maass for technical assistance. This work was carried out as part of the Canadian contribution to the Joint Global Ocean Flux Study.

References

- Brock J C, McClain C R, Luther M E and Hay W W 1991 The phytoplankton bloom in the northwestern Arabian Sea during the southwest monsoon of 1979; *J. Geophys. Res.* **96** 20,623–20,642
- Denman K L 1973 A time-dependent model of the upper ocean; *J. Phys. Oceanogr.* **3** 173–184
- Feldman G, Kuring N, Ng C, Esaias W, McClain C R, Elrod J, Maynard N, Endres D, Evans R, Brown J, Walsh S, Carle M and Podesta G 1989 Ocean colour. Availability of the global data set; *EOS Trans. Am. Geophys. Union* **70** 634–635
- Gaspar P 1988 Modeling the seasonal cycle of the upper ocean; *J. Phys. Oceanogr.* **18** 161–180
- Hastenrath S and Greischar L L 1989 *Climatic Atlas of the Indian Ocean. Part III: Upper-Ocean Structure* (Wisconsin: University of Wisconsin Press) pp. 247
- Lewis M R, Cullen J J and Platt T 1983 Phytoplankton and thermal structure in the upper ocean: Consequence of nonuniformity in chlorophyll profile; *J. Geophys. Res.* **88** 2565–2570
- McClain C R, Esaias W E, Barnes W, Guenther B, Endres D, Hooker S, Mitchell G and Barnes R 1992 SeaWiFS Calibration and Validation Plan; *NASA Tech. Memo.* **3** 1–41

- McCreary J P and Kundu P K 1989 A numerical investigation of sea surface temperature variability in the Arabian Sea; *J. Geophys. Res.* **94** 16,097–16,114
- Niiler P P and Kraus E B 1977 One-dimensional models of the upper ocean; In: *Modelling and Prediction of the Upper Layers of the Ocean* (ed) E B Kraus (Pergamon Press) pp. 143–172
- Oberhuber J M 1988 *An atlas based on the 'COADS' data set: The budgets of heat, buoyancy and turbulent kinetic energy at the surface of the global ocean*, Report No. 15 (Max-Planck-Institut für Meteorologie Hamburg) pp. 20
- Platt T and Sathyendranath S 1993 Estimators of primary production for interpretation of remotely sensed data on ocean color; *J. Geophys. Res.* **98** 14,561–14,576
- Platt T, Harrison W G, Lewis M R, Li W K W, Sathyendranath S, Smith R E and Vézina A F 1989 Biological production of the oceans: The case for a consensus; *Mar. Ecol. Prog. Ser.* **52** 77–88
- Platt T, Bird D F and Sathyendranath S 1991a Critical depth and marine primary production; *Proc. R. Soc. London Ser. B.* **246** 205–217
- Platt T, Caverhill C and Sathyendranath S 1991b Basin-scale estimates of oceanic primary production by remote sensing: The North Atlantic; *J. Geophys. Res.* **96** 15,147–15,159
- Platt T, Woods J D, Sathyendranath S and Barkmann W 1994 Primary production, respiration and stratification in the ocean; *J. Geophys. Res.* In press
- Ravindran 1994 Optical variability in the upper ocean: Implications for mixed-layer modelling. In prep.
- Sathyendranath S, Gouveia A D, Shetye S R, Ravindran P and Platt T 1991 Biological control of surface temperature in the Arabian Sea; *Nature.* **349** 54–56
- Shetye S R 1986 A model study of the seasonal cycle of the Arabian Sea surface temperature; *J. Mar. Res.* **44** 521–542
- Simonot J -Y, Dollinger E and Le Treut H 1988 Thermodynamic-biological-optical coupling in the oceanic mixed layer; *J. Geophys. Res.* **93** 8193–8202
- Sverdrup H U 1953 On conditions for the vernal blooming of phytoplankton; *J. du Conseil* **18** 287–295