O_2 and CO_2 balances approach for estimating biological production in the mixed layer of the tropical Atlantic Ocean (Guinea Dome area)

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ABSTRACT

The rates at which concentrations of oxygen and carbon dioxide change under the influence of the exchange with the atmosphere and biological production are derived from *in situ* measurements of O_2 and CO_2 in the mixed layer over a 10-day period in the Guinea Dome area. The rates of O_2 production (10.3 mmol m^{-2} d^{-1}) and CO_2 consumption (13.0 mmol m^{-2} d^{-1}) by photosynthetic activity exceed the evasion of O_2 and CO_2 into the atmosphere (respectively 2.4 and 1.2 mmol m^{-2} d^{-1}). The net apparent production of O_2 from day to day is only 11% of the diurnal O_2 increase. The biological production of oxygen at a short distance from the center of the Guinea Dome appears to be higher than that in the exact center.

1. Introduction

The cycles of oxygen (O_2) and carbon dioxide (CO_2) in the mixed layer are influenced by the exchange of gases across the air-sea interface and biological activity (photosynthesis and respiration). The exchange rates of gases between ocean and atmosphere are of great interest, because of the role that the ocean is assumed to play in the earth's global cycle for several gaseous components. Studies to determine primary productivity by indirect methods have recently been stimulated by criticisms of the standard ^{14}C method which may seriously underestimate primary production in the open ocean.

Redfield (1948) was the first to propose a method for estimating the exchange rate of oxygen from the simultaneous changes with time in dissolved oxygen and phosphate concentrations in seawater. For that he used a relation between O_2 and PO_4. A similar method was followed by Pytkowicz (1964) off the Oregon coast and Tsunogai and Tanaka (1980) in the Funka Bay (Japan). Johnson *et al.* (1979) introduced a new method by measuring the simultaneous changes in oxygen, pH and titration alkalinity over a 15-day period in the Stuart Channel (British Columbia). Using Pytkowicz's procedure an attempt was made by the author (Oudot, 1984) to evaluate the rate of transfer of oxygen across the sea surface in the open ocean (eastern equatorial

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Atlantic). Other studies were undertaken, motivated by interest in determining the rate of CO_2 exchange because of concern about what happened to anthropogenic CO_2. These studies followed experimental approaches through measurements both in the laboratory (wind tunnel) and in the field (radon deficiency-sulphur hexafluoride evasion). Liss and Merlivat (1986) summarized the results from such studies and proposed relationships for the prediction of gas exchange rates across the sea surface from the wind speed in the marine environment. As an alternative to the methods of measurements of gas flux on the water side of the interface, Jones and Smith (1977) proposed the eddy correlation technique on the air side of the interface to determine the air-sea CO_2 flux. A large discrepancy (an order of magnitude) between results from eddy correlation and other techniques raises a serious controversy in estimates of CO_2 flux between geochemists and physicists of the air-sea interaction (Broecker *et al.,* 1986; Smith and Jones, 1986; Wesely, 1986). Recently Frankignoulle (1988) made air-sea CO_2 exchange measurements with a direct chamber method and reported exchange coefficients which agree well with those determined with radiotracers or in wind tunnels.

Gieskes *et al.* (1979) reviewed the current ^{14}C methods for measuring primary production and raised the question of the underestimates made by this method in large parts of oligotrophic oceanic waters. Tijssen (1979), Shulenberger and Reid (1981) and Jenkins (1982) came to the same conclusion by different methods. Although more recent research has reasserted the validity of the ^{14}C uptake method (Codispoti *et al.,* 1982; Williams *et al.,* 1983; Platt and Harrison, 1985), the controversy is not extinct (Jenkins and Goldman, 1985; Reid and Shulenberger, 1986; Platt and Harrison, 1986). Therefore the indirect methods using calculations based on *in situ* changes in oxygen and nutrients seems a realistic method to determine the magnitude of primary production in oligotrophic waters. Codispoti *et al.* (1982) calculated a net community organic carbon production from the interpretation of the changes in CO_2 (partial pressure and total inorganic carbon concentration). Minas *et al.* (1986) have recently shown that they were able to calculate a "net community production" in coastal upwelling areas derived from phytoplankton nutrient consumption or photosynthetic oxygen production and heat increase proportional to the time lapse. In this paper we chose to follow the approach of Johnson *et al.* (1979) and to apply their analysis of changes in O_2 and CO_2 to a tropical dome situation (Guinea Dome). The idea of using O_2 and CO_2 changes for the calculation of biological production is not new. Johnson *et al.* (1981) reviewed studies using O_2 and CO_2 changes to monitor ecosystem metabolism. Most often O_2 and ΣCO_2 (total inorganic carbon) variations have been studied in closed or semi-closed areas or in coastal marine waters but seldom in the open ocean. Moreover ΣCO_2 was not directly measured but calculated from pH and alkalinity. Most of these studies indicate that ΣCO_2-based estimates usually exceed those based on O_2 and that direct measurements consistently exceed ^{14}C determinations. Unlike ^{14}C measurements, which are thought to indicate some value between gross and net
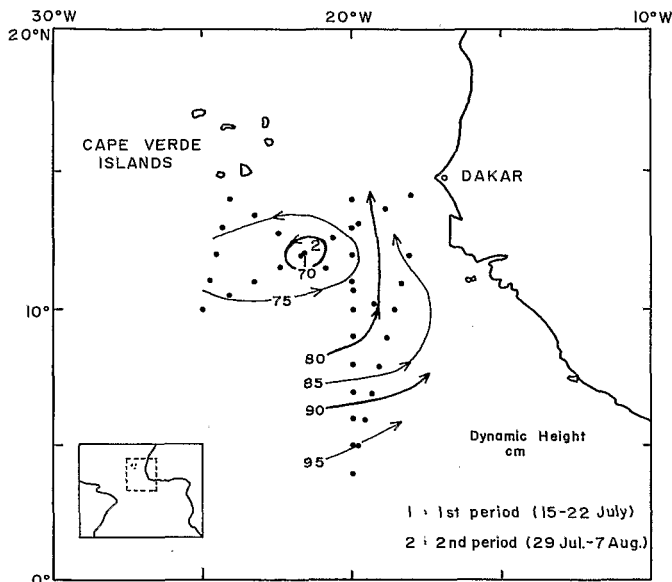
Figure 1. PIRAL cruise of the R.V. *Noroit* (July–August 1986). The dynamic height (cm) and geostrophic circulation (cyclonic) at the sea surface versus 500 dbar locate the Guinea Dome area. The numbers 1 and 2 place the two sites of continuous observations.

production of the phytoplankton, ΣCO_2-based calculations show net rates of production and respiration by the whole community and therefore represent the metabolism of the total ecosystem.

We present below the analysis of O_2 and ΣCO_2 changes over periods of 10 and 8 successive days. The observations of O_2 and ΣCO_2 using direct analyses were made at 07.30 h and 18.30 h (local time). The biological production is derived from the trend from day to day of the O_2 and ΣCO_2 variations after correcting for the physical exchanges of gases across the sea surface and for vertical eddy diffusion from the thermocline layers. The production rates deduced from O_2 and ΣCO_2 are compared with each other and with other results.

2. Source of data and methods

The hydrological and chemical data analyzed in this study come from the PIRAL cruise (June–August 1986) carried out aboard the French R.V. *Noroit* in the northern tropical Atlantic. After a preliminary phase (9–13 July) to locate and describe the Guinea Dome (Fig. 1), observations were carried out over 8 and 10-day periods in two locations in the central area: the first site (15–22 July) was at 11°50N, 22W and the second one (29 July–7August) was at 12N, 21W. We shall analyze the second set of data first because during this longer period the hydrological structure of the mixed

layer appears to have been more stable and the diurnal oxygen rhythm more regular and more pronounced.

The seawater samples were taken with a rosette (General Oceanics) of twelve Niskin PVC five-liter bottles which was associated with a Neil Brown CTDO probe. The observations of all chemical properties were made from depths of 0 to 500 m at 07.30 h and 18.30 h (local time). Additional observations of dissolved oxygen were made from 0 to 150 m at 13.30 h and 23.30 h (local time).

Temperature and salinity data from the CTDO probe were simultaneously collected when the bottles were closed.

Dissolved oxygen was titrated according to the Winkler method (Carritt and Carpenter, 1966) with a precision (σ) of 0.1% (Oudot *et al.*, 1988). The oxygen solubility was calculated by Weiss' (1970) polynomial relationship.

Dissolved molecular nitrogen and argon were measured by gas chromatography according to the methods described in Oudot (1982) and Oudot and Montel (1987). The precision (σ) of the results is 0.4% for nitrogen and 1.0% for argon. The solubility of nitrogen and argon was calculated by Weiss' (1970) polynomial relationships.

Measurements of CO_2 partial pressures in seawater (PCO_2) and in atmosphere (pCO_2) were made by infrared absorption according to the methods described in Oudot and Andrié (1986) and Andrié *et al.*, (1986).

Measurements of total inorganic carbon ΣCO_2 were made by the gas chromatography method described by Oudot and Wauthy (1978) and derived from that of Weiss and Craig (1973). The precision (σ) of ΣCO_2 measurements is 0.2%.

Total alkalinity data were calculated from ΣCO_2 and pH using Millero's (1979) polynomial relationships for the apparent ionization constants for the carbonate system and the borate system. The precision of the pH measurements, on the NBS scale, is evaluated at 0.001 pH unit (σ) (Oudot, 1987).

3. Hydrological conditions of the Guinea Dome

Thermal domes are the subject of a number of studies: Wyrtki (1964), Broenkow (1965), Minas *et al.* (1974) in the northern Pacific (Costa Rica Dome) and Mazeika (1967), Voituriez and Dandonneau (1974), Gallardo *et al.* (1974) in the northern and southern Atlantic (Guinea Dome and Angola Dome). The interest of these areas consists in the enhanced primary production which is observed there.

The dome, so-called owing to the anticlinal thermal structure, is caused by the northward deflection of the North Equatorial Contercurrent developing a cyclonic circulation with the North Equatorial Current (Fig. 1).

The horizontal mixing between the surface water at the center of the dome and the surrounding gyre is weak as shown by the horizontal distribution of salinity (Fig. 2). The high-salinity water ($S > 36.2$) in the Guinea Dome is well separated from the low-salinity water ($S < 35.4$) by the eastward flow deflecting northward as it comes up against the coast of Africa. In the central area of the dome, the thermocline is
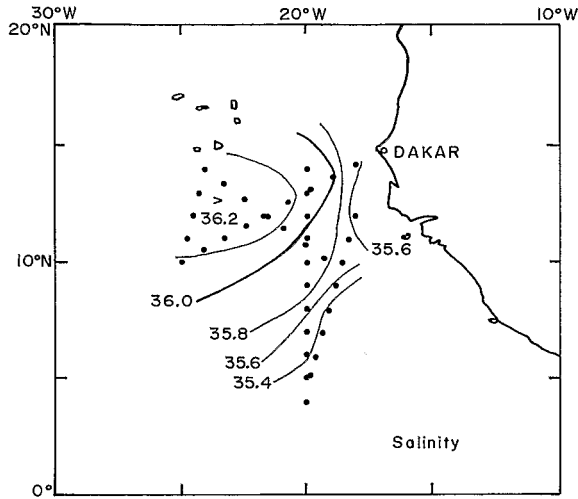
Figure 2. Horizontal distribution of salinity at the sea surface during PIRAL cruise.

shallower (Fig. 3) and the hydrological features are stable, as shown by the very small horizontal motions: depth, temperature and salinity do not change very much. Voituriez and Dandonneau (1974) described the Guinea Dome in detail. The center of the dome (12N, 22W) can be determined by the coldest temperature, the lowest salinity and oxygen and the highest phosphate content at 50 m depth (Fig. 3). The first period of our continuous observations (8 days) was chosen in the exact center of the dome whereas the second period (10 days) took place at a short distance from the center (about 110 km).

The most striking feature of the Guinea Dome is that it displays strong vertical gradients in all the physico-chemical properties of the water column quite near the sea surface without, however, quite reaching it (Fig. 3). Unlike the Costa Rica Dome, the mixed layer was never absent in the Guinea Dome. Figure 4 shows, for the two sites, the vertical extent of the mixed layer and how mixed it was. It is evident that the surface divergence was not strong enough so that the ascending motion did not reach the surface. Moreover the vertical distribution of dissolved oxygen shows a continuous subsurface maximum in the upper part of the thermocline (Oudot, 1989), that, however, appears to vanish in the center of the dome (Fig. 3).

4. Principle of data analysis

Johnson *et al.* (1979), inferring from the strong stratification (temperature and salinity) in the Stuart Channel that the mixing between the deep water and the surface layer was inhibited, expressed the change in the oxygen concentration of seawater as the result of the changes due to invasion or evasion of gas from or into the atmosphere and the changes due to biological processes. Nevertheless we cannot overlook the
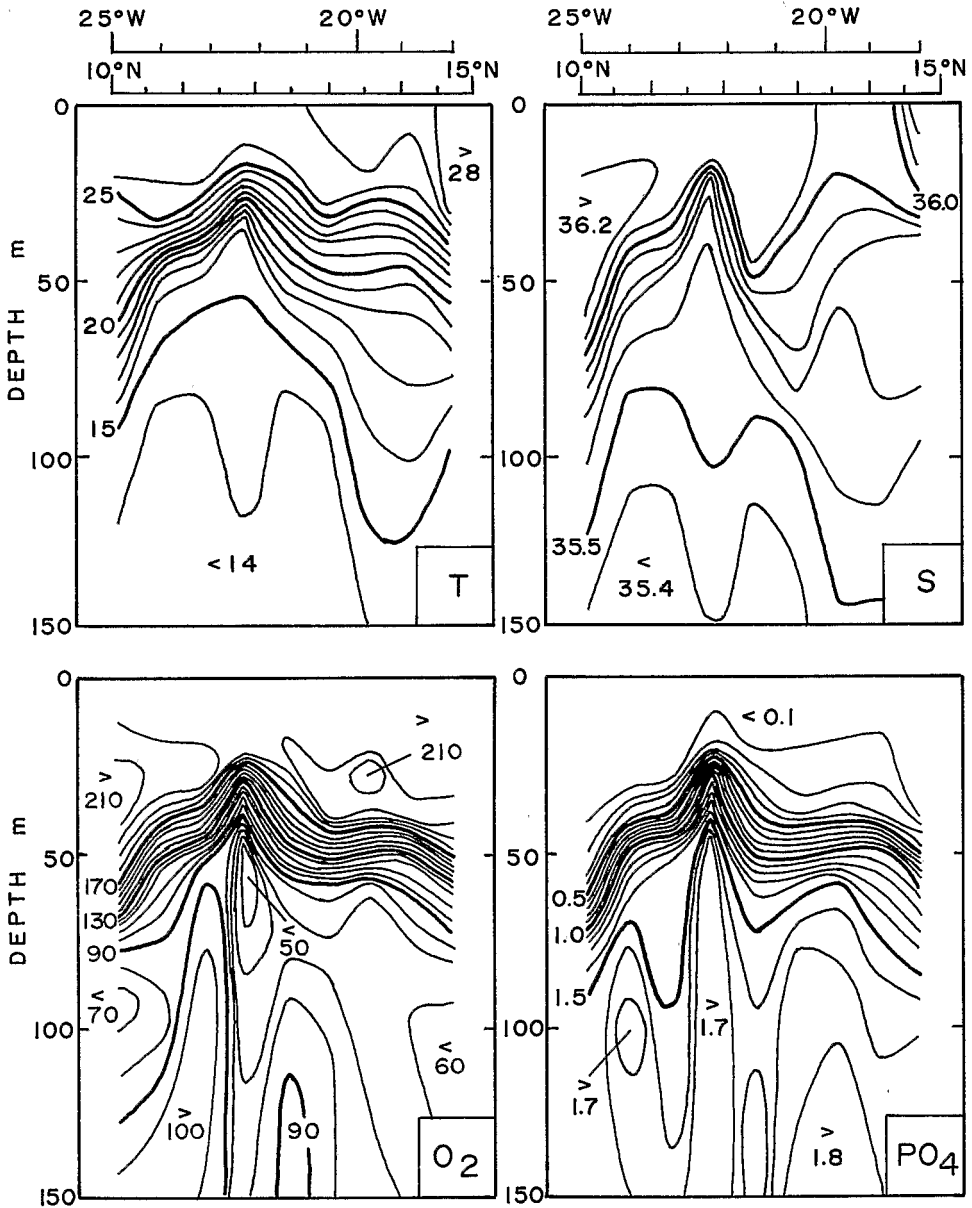
Figure 3. Vertical distributions of temperature T (°C), salinity, S, oxygen O_2 (μmol kg^{-1}) and phosphate PO_4 (μmol kg^{-1}) across the Guinea Dome. The center of the dome is determined by the coldest temperature, the lowest salinity and oxygen and the highest phosphate content at 50 m depth.
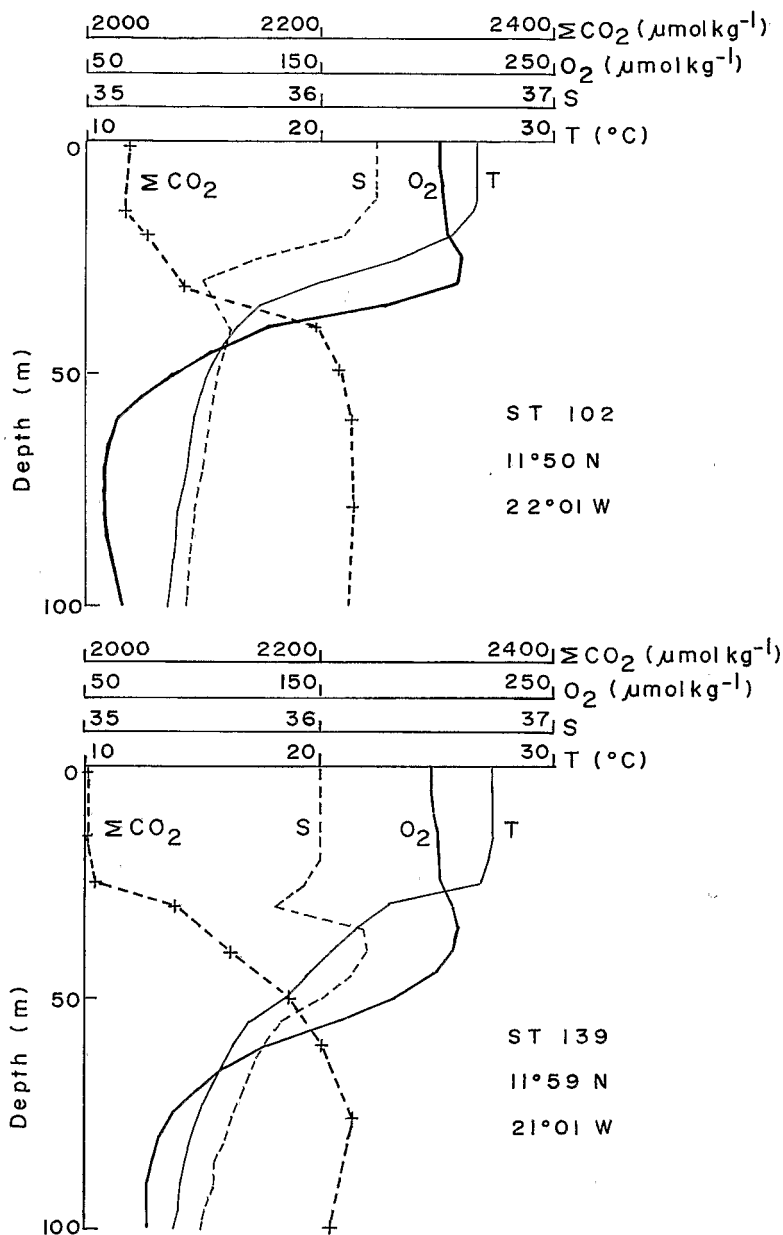
Figure 4. Vertical profiles of temperature (T), salinity (S), oxygen (O_2) and total inorganic carbon (ΣCO_2) at the exact center (St. 102) and at a short distance from the center (St. 139) of the Guinea Dome. Temperature, salinity and oxygen are continuous profiles given by the CTDO probe and total inorganic carbon is the result of the determinations (gas chromatography) made on the seawater samples taken with the rosette.

turbulent mixing between the surface layer and the upper layer of the thermocline which encloses the oxygen maximum and we shall state that the change in the oxygen content of the mixed layer over time ΔO_2 is:

$$\Delta O_2 = \Delta O_2^a + \Delta O_2^b + \Delta O_2^d \tag{1}$$

where superscript a indicates the exchange with the atmosphere, superscript b the exchange through biological processes and superscript d refers to changes due to vertical diffusional flux from the subsurface maximum. Likewise we shall express the change in the ΣCO_2 concentration over the same period of time $\Delta \Sigma CO_2$ as:

$$\Delta \Sigma CO_2 = \Delta \Sigma CO_2^a + \Delta \Sigma CO_2^c + \Delta \Sigma CO_2^b + \Delta \Sigma CO_2^d \tag{2}$$

where superscripts a and b have the same meaning as previously, superscript c refers to the change due to dissolution or precipitation of carbonate minerals (Johnson et al., 1979) and superscript d refers to changes due to vertical diffusional flux from the lower layers which are richer in dissolved inorganic carbon. Contrary to oxygen, the cycle of CO_2 in seawater is affected by the change of all the forms of dissolved inorganic carbon and not only by that of gaseous carbon dioxide, since dissolved CO_2 in seawater is in equilibrium with carbonate and bicarbonate ions:

$$\Sigma CO_2 = CO_2 + HCO_3^- + CO_3^{--}. \tag{3}$$
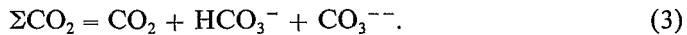
The quantities on the left-hand side of relations (1) and (2) are derived from the changes in O_2 and CO_2 concentrations as a function of time in the mixed layer.

The individual quantities on the right-hand side of relations (1) and (2) are evaluated separately.

a. The exchange with the atmosphere. ΔO_2^a and $\Delta \Sigma CO_2^a$ are calculated as the flux of O_2 and CO_2 across the air-sea interface. We chose to parameterize the gas flux between atmosphere and ocean by means of the boundary-layer model of gas-exchange (Deacon, 1977). When applied to air-sea exchange, the eddy correlation technique produces results much higher than all values obtained by other techniques (Liss and Merlivat, 1986). The net gas flux (F^a) across the interface is given by

$$F^a = k(C_{sw} - \alpha C_a) \tag{4}$$

where k is the gaseous exchange coefficient (or "piston velocity") (Broecker and Peng, 1974), C_{sw} and C_a are the gas concentrations in seawater and in air, respectively and α is the solubility of gas in seawater (Weiss, 1970; 1974).

The exchange coefficient k, changing from one type of gas to the other, is a function of wind speed and of temperature of seawater through the Schmidt number Sc dependency (Liss and Merlivat, 1986). The exchange coefficient $k(O_2)$ is derived from the one calculated for CO_2 taking into account the proportionality of k to $Sc^{-2/3}$ for wind speeds ≤ 3.6 m s^{-1} and to $Sc^{-1/2}$ for higher wind speeds. The relationships used

for the calculation of the CO_2 exchange coefficient are written as follows:

$$k_{20}(CO_2) = 0.17 \, V_{10} \qquad \text{for} \quad 0 \le V_{10} \le 3.6 \text{ ms}^{-1} \tag{5}$$

$$k_{20}(CO_2) = 2.85 \, V_{10} - 9.65 \qquad \text{for} \quad 3.6 \le V_{10} \le 13 \text{ ms}^{-1} \tag{6}$$

where $k_{20}(CO_2)$ is the CO_2 exchange coefficient at 20°C expressed in cm h^{-1} and V_{10} is the wind speed in ms^{-1} measured at 10 m height. The dependence of $k(CO_2)$ on the temperature is evaluated as follows:

$$k_t = 70.74 \, k_{20}(Sc_t)^{-2/3} \qquad \text{for} \quad 0 \le V_{10} \le 3.6 \text{ ms}^{-1} \tag{7}$$

$$k_t = 24.39 \, k_{20}(Sc_t)^{-1/2} \qquad \text{for} \quad 3.6 \le V_{10} \le 13 \text{ ms}^{-1} \tag{8}$$

where Sc_t, the Schmidt number for CO_2 at t°C, is linearly interpolated between 20 and 30°C with the following relationship:

$$Sc_t = 1065 - 23.5 \, t. \tag{9}$$

At 20°C, Sc is equal to 595 for CO_2 and to 470 for O_2 (Liss and Merlivat, 1986). Thus, taking into account the wind speed effect, $k_t(O_2)$ is related to $k_t(CO_2)$ as follows:

$$k_t(O_2) = 1.17 \, k_t(CO_2) \qquad \text{for} \quad 0 \le V_{10} \le 3.6 \text{ ms}^{-1} \tag{10}$$

$$k_t(O_2) = 1.13 \, k_t(CO_2) \qquad \text{for} \quad 3.6 \le V_{10} \le 13 \text{ ms}^{-1}. \tag{11}$$

Finally, the O_2 and CO_2 flux formulas that we will use for calculation are:

$$F^a(O_2) = 0.24 \, k(O_2 - O_2^*) \tag{12}$$

with $F^a(O_2)$, equivalent to ΔO_2^a, expressed in millimoles per square meter per day, k expressed in centimeters per hour, O_2 and O_2^*, respectively the dissolved oxygen concentration and oxygen solubility, expressed in micromoles per kilogram

$$F^a(CO_2) = 0.24 \, k\alpha(PCO_2 - pCO_2) \tag{13}$$

with $F^a(CO_2)$, identical with $\Delta\Sigma CO_2^a$ because only molecular CO_2 is exchanged with atmosphere, expressed in millimoles per square meter per day, k expressed in centimeters per hour, α expressed in moles per kilogram per atmosphere and PCO_2 and pCO_2, the CO_2 partial pressures in the water and the air, respectively, expressed in microatmospheres.

b. The vertical diffusional flux. ΔO_2^d and $\Delta\Sigma CO_2^d$ are evaluated according to the Fickian diffusion equation of which the flux F^d is given by

$$F^d = K_z dC/dz \tag{14}$$

where F^d is the vertical flux of oxygen or dissolved inorganic carbon, K_z the vertical

eddy diffusivity, C the concentration of oxygen or inorganic carbon and z the depth. We have chosen this approach ourselves (Oudot, 1984) following many other investigators (Dugdale, 1967; Eppley *et al.*, 1979; King and Devol, 1979; Denman and Gargett, 1983). Recently Garside (1985) used this model to describe the nitrate distribution in the mixed layer of the open ocean. K_z is determined as was done by Denman and Gargett (1983)

$$K_z = 0.25 \, \epsilon N^{-2} \tag{15}$$

where ϵ represents the turbulent energy dissipation and N the buoyancy frequency. N, expressed as the inverse of time (seconds), measures the local stratification and is calculated from the vertical density gradient $d\rho/dz$

$$N^2 = (g/\rho)(d\rho/dz) \tag{16}$$

where g is the acceleration due to gravity (9.8 ms^{-2}). For the upper part of the thermocline, with low winds (round about 5 ms^{-1}), the dissipation rate ϵ is taken as equal to $2 \times 10^{-8} \, \mathrm{m^2 \, s^{-3}}$ (Dillon and Caldwell, 1980). Then K_z is expressed in square meters per second.

c. The change of carbonate minerals. $\Delta\Sigma CO_2^c$, the change in ΣCO_2 due to precipitation or dissolution of carbonate minerals, is evaluated by Johnson *et al.* (1979) from the calculated change in the total alkalinity ΔTA and $\Delta\Sigma CO_2^b$ as:

$$\Delta\Sigma CO_2^c = 0.5(\Delta TA + 17/106 \, \Delta\Sigma CO_2^b). \tag{17}$$

The first component on the right-hand side of expression (17) arises because the change in TA due to the variations in carbonate ions is twice as large as $\Delta\Sigma CO_2$ (Skirrow, 1975). The second component arises from the change in TA with the release of acid (16 molecules of nitric acid and one of phosphoric acid for 106 molecules of CO_2) (Brewer *et al.*, 1975) during the degradation of biomass with Redfield's *et al.* (1963) average composition. Today the stoichiometric composition of biogenic material of Redfield's *et al.* model is seriously controverted and modified versions are proposed. For Jones *et al.* (1984) a better estimate of the composition of the biogenic material is: a $P/N/C/O_2$ ratio of 1/10/135/168 and for Takahashi *et al.* (1985) 1/16/122($\pm$18)/172. We chose the average composition of Takahashi *et al.* (1985) and we calculated the biological change in TA equal to (17/122) $\Delta\Sigma CO_2^b$

$$\Delta\Sigma CO_2^c = 0.5 \, (\Delta TA + 17/122 \, \Delta\Sigma CO_2^b). \tag{18}$$

Once the quantities ΔO_2^a, ΔO_2^d, $\Delta\Sigma CO_2^a$, $\Delta\Sigma CO_2^c$, and $\Delta\Sigma CO_2^d$ of relations (1) and (2) are evaluated, we have only to calculate by differentiation the quantities ΔO_2^b and $\Delta\Sigma CO_2^b$ representing the amounts exchanged through the biological processes.
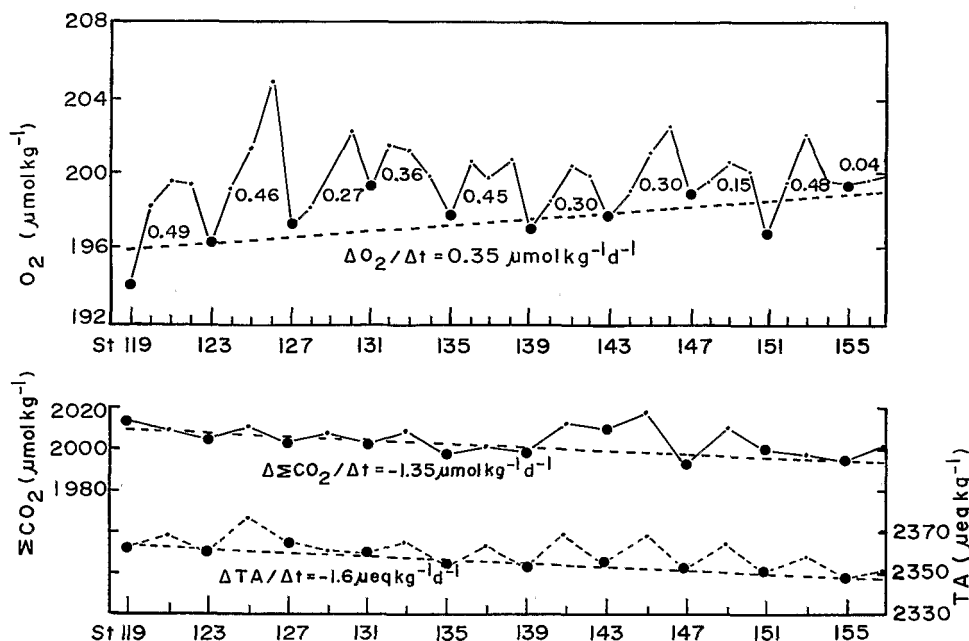
Figure 5. Variations of dissolved oxygen O_2, dissolved inorganic carbon ΣCO_2 and total alkalinity TA in the mixed layer (depth = 24 ± 3 m) over the 10-day period at 12N, 21W. The station numbers indicate the morning station (07.30 h). The slopes of the regression lines fitted to the morning station data (solid dots on the graphs) give the rate of change in each quantity over the 10-day period. For O_2 it is moreover showed for each day the mean diurnal increase of dissolved oxygen (μmol kg^{-1}h^{-1}) between the morning station (07.30 h), the midday station (13.30 h) and the evening station (18.30 h).

5. Results

The data for the two periods of observations are shown in Figure 5 and 6 and in Tables 1 and 2. All the observations are averaged over the depth of the mixed layer (24 ± 3 m for the first data set and 14 ± 6 m for the second one).

During the second period (Fig. 5), the time variation in dissolved oxygen concentration in the mixed layer is strongly modulated by the diurnal rhythm of photosynthetic activity. Moreover there appears to be a general trend toward an increase in dissolved oxygen between the beginning and the end of the 10-day period. At the morning station (07.30 h), the dissolved oxygen increased significantly (t = 2.42 for n = 10) from day to day. A linear regression is fitted to the data to quantify this trend from day to day and the slope of the regression line shown in Figure 5 gives the rate of O_2 change: 0.35 ± 0.14 μmol kg^{-1} d^{-1}. The data also allow us to calculate a mean diurnal increase in dissolved oxygen *in situ:* 0.33 ± 0.15 μmol kg^{-1} h^{-1} between the morning station (07.30 h), the midday station (13.30 h) and the evening station (18.30 h).

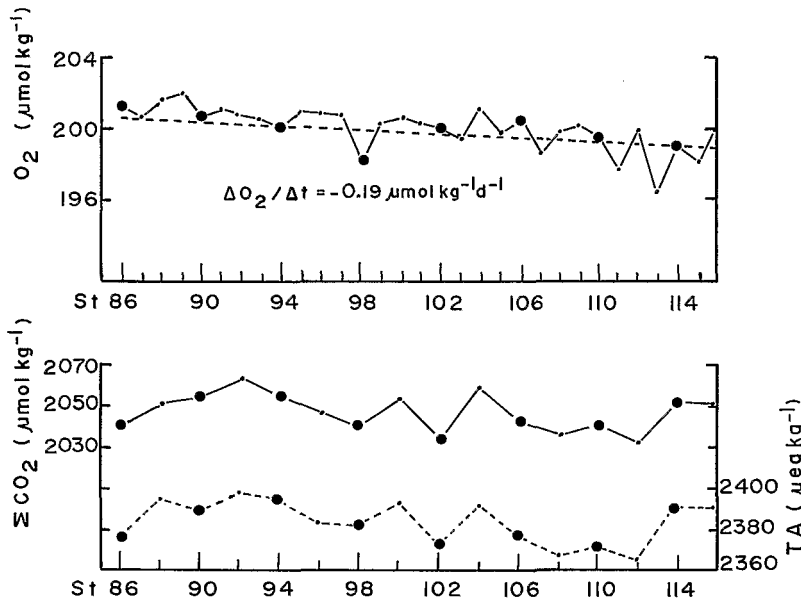Figure 6. Variations of dissolved oxygen O_2, dissolved inorganic carbon ΣCO_2 and total alkalinity TA in the mixed layer over the 8-day period at 11°50N, 22W. The station numbers indicate the morning station (07.30 h). The slope of the regression line fitted to the morning station data (solid dots on the graph) gives the rate of change in O_2 quantity over the 8-day period.

During the same time lapse we note a significant decrease ($t = 2.42$ for $n = 10$) in dissolved inorganic carbon between the first day and the last one. The same statistical treatment of ΣCO_2 data (morning station) leads us to quantify the rate of ΣCO_2 change as -1.35 ± 0.6 μmol kg^{-1} d^{-1}. The variation in ΣCO_2 seems strongly modulated by the variation of total alkalinity TA. The significant decrease ($t = 6.21$ for $n = 10$) in TA leads us to quantify the rate of TA change as -1.6 ± 0.3 μeq kg^{-1} d^{-1}.

During the first period (Fig. 6), the diurnal oxygen rhythm is much less apparent and the general trend from day to day seems to be a decrease (not significant: $t = 1.46$ for $n = 8$) between the first day and the last one. Nevertheless we calculated the slope of the regression line fitted to the data of the morning station (07.30 h) in order to attempt to estimate and interpret this oxygen decrease: the rate is -0.19 ± 0.13 μmol kg^{-1} d^{-1}. From the ΣCO_2 data no significant trend emerges either. During this shorter period, in which measurements were taken close to the center of the dome, the depth of the mixed layer was changed considerably: from 8 to 20 m with a mean depth of 14 m. Perhaps this lack of stability is the reason why no clear systematic trend in the plot of O_2 and ΣCO_2 over time becomes visible.

The results of calculation of the CO_2 and O_2 exchanges through the air-sea interface during the 10-day period, by means of the relations 13 and 12, are indicated in Table 3.

Table 1. Variations over the 10-day period (29 July–7 Aug. 1986) of O_2, ΣCO_2, TA and related parameters averaged in the mixed layer (depth Z) in the Guinea Dome (12N, 21W)

Station no.	Local time	Z (m)	T (°C)	S	O_2	ΣCO_2	TA
					(μmol kg^{-1})		(μeq kg^{-1})
119	0715	20	27.01	36.05	194.0	2016	2366
120	1300	24	27.11	36.03	198.1		
121	1830	25	27.13	36.02	199.4	2012	2372
122	2330	24	27.06	36.04	199.3		
123	0730	22	26.98	36.03	196.1	2008	2364
124	1330	22	27.14	36.03	199.0		
125	1845	24	27.34	36.02	201.3	2014	2381
126	2345	22	27.20	36.02	204.8		
127	0715	23	27.20	36.00	197.0	2007	2369
128	1320	24	27.29	36.00	198.1		
129	1830	22	27.42	36.00	200.1	2010	2365
130	2345	22	27.37	36.01	202.2		
131	0715	24	27.13	36.01	199.3	2006	2364
132	1330	29	27.22	36.02	201.5		
133	1830	23	27.30	36.04	201.1	2012	2369
134	2330	24	27.30	36.06	199.8		
135	0720	20	27.29	36.00	197.7	2001	2358
136	1335	20	27.44	36.01	200.6		
137	1840	22	27.41	36.01	199.7	2005	2367
138	2335	24	27.32	36.01	200.6		
139	0730	24	27.37	36.01	197.0	2002	2358
140	1335	28	27.41	36.01	198.4		
141	1830	24	27.41	36.02	200.4	2016	2374
142	2340	26	27.43	36.02	199.8		
143	0720	25	27.36	36.01	197.6	2014	2360
144	1320	22	27.31	35.98	199.0		
145	1830	30	27.33	35.99	201.0	2010	2373
146	2330	20	27.34	36.02	201.9		
147	0730	22	27.17	35.99	198.9	1997	2356
148	1330	24	27.35	36.00	199.6		
149	1840	36	27.33	35.99	200.5	2014	2370
150	2330	28	27.21	35.99	200.1		
151	0730	25	27.31	35.99	196.8	2003	2355
152	1330	28	27.57	35.98	199.6		
153	1822	23	27.45	36.00	202.1	2001	2363
154	2330	27	27.38	35.99	199.6		
155	0730	25	27.11	35.96	199.4	1998	2352
156	1330	21	27.30	35.96	199.6		
157	1720	28	27.36	35.96	199.8	2005	2356
Mean		24	27.29	36.01			
SD		± 3	± 0.13	± 0.02			

Table 2. Variations over the 8-day period (15–22 July 1986) of O_2, ΣCO_2, TA, and related parameters averaged in the mixed layer (depth Z) in the Guinea Dome (11°50N, 22W)

Station no.	Local time	Z (m)	T (°C)	S	O_2 (μmol kg^{-1})	ΣCO_2	TA (μeq kg^{-1})	N_2 (μmol kg^{-1})	Ar
86	0730	12	26.32	36.20	201.1	2041	2381	396	10.47
87	1310	8	26.51	36.21	200.5				
88	1835	7	26.74	36.22	201.5	2053	2399	394	—
89	2340	14	26.57	36.20	201.8				
90	0745	16	26.33	36.21	200.6	2056	2393	397	10.51
91	1315	18	26.51	36.22	200.9				
92	1825	16	26.67	36.22	200.7	2066	2402	396	10.38
93	2340	20	26.56	36.22	200.5				
94	0730	15	26.27	36.23	200.0	2057	2399	398	10.42
95	1330	14	26.57	36.23	200.8				
96	1830	10	26.66	36.23	200.7	2049	2387	397	10.30
97	2330	15	26.54	36.23	200.6				
98	0730	20	26.39	36.23	198.2	2042	2386	395	10.33
99	1330	23	26.59	36.23	200.2				
100	1840	21	26.69	36.23	200.5	2056	2396	395	10.12
101	2330	23	26.45	36.23	200.2				
102	0730	14	26.46	36.24	200.0	2035	2376	396	10.38
103	1330	12	26.58	36.24	199.4				
104	1845	10	26.66	36.24	201.2	2062	2395	394	10.42
105	2335	15	26.53	36.23	199.8				
106	0725	22	26.43	36.24	200.5	2045	2380	—	10.42
107	1340	16	26.63	36.24	198.7				
108	1830	20	26.88	36.25	199.9	2038	2371	394	10.07
109	2330	20	26.80	36.31	200.0				
110	0730	8	26.76	36.30	199.5	2043	2375	396	10.20
111	1330	4	27.33	36.21	197.8				
112	1900	2	27.53	26.28	199.9	2034	2369	393	10.25
113	2320	5	27.43	36.30	196.7				
114	0720	12	26.85	36.25	199.1	2056	2394	392	10.16
115	1330	17	26.90	36.30	198.2				
116	1830	18	26.88	36.29	200.0	2063	2400	394	10.16
Mean		14	26.68	36.24					
SD		± 6	± 0.30	± 0.03					

N_2 (μmol kg^{-1}): concentration of dissolved molecular nitrogen
Ar (μmol kg^{-1}): concentration of dissolved argon

The mean evasion rate of O_2 from the ocean into the atmosphere (ΔO_2^g) is equal to 2.4 ± 2.7 mmol m^{-2} d^{-1}. The mean CO_2 flux escaping from the ocean into the atmosphere $\Delta \Sigma CO_2^g$ is equal to 1.2 ± 1.1 mmol m^{-2} d^{-1}. The large uncertainties on ΔO_2^g and $\Delta \Sigma CO_2^g$ determinations are mainly due to the wide range of variations of O_2 and CO_2 exchange coefficients; besides that ΔO_2^g is increased by the variation of the saturation anomaly ($O_2 - O_2^*$).

Table 3. Variations of CO_2 and O_2 fluxes across the air-sea interface over the 10-day period (29 July–7 August 1986) in the Guinea Dome (12N, 21W)

Station	V_{10} (ms^{-1})	T (°C)	$k_t(CO_2)$ (cm h^{-1})	$PCO_2 - pCO_2$ (μatm)	$\Delta\Sigma CO_2^a$ (mmol m^{-2}d^{-1})	$k_t(O_2)$ (cm h^{-1})	$O_2 - O_2^*$ (μmol kg^{-1})	ΔO_2^a (mmol m^{-2}d^{-1})
119	9.5	27.02	20.5	32.7	4.4	23.2	−4.2	(−23.4)
120	12.0	27.13	29.0			32.8	0.3	2.4
121	6.5	27.12	10.5	35.5	2.5	11.9	1.7	4.9
122	5.0	27.06	5.4			6.1	1.3	1.9
123	4.3	26.98	3.1	44.3	0.9	3.5	−2.2	−1.8
124	4.0	27.32	2.1			2.4	1.3	0.7
125	3.6	27.34	0.8	36.2	0.2	0.9	4.1	0.9
126	1.5	27.20	0.3			0.4	7.2	0.7
127	1.6	27.21	0.3	30.8	0.1	0.4	−0.6	−0.1
128	1.5	27.41	0.3			0.4	0.8	0.1
129	3.1	27.77	0.7	41.5	0.2	0.8	3.2	0.6
130	2.0	27.48	0.4			0.5	5.2	0.6
131	2.8	27.24	0.6	39.1	0.2	0.7	1.4	0.2
132	2.0	27.37	0.4			0.5	4.0	0.5
133	5.1	27.53	5.8	40.9	1.6	6.6	3.9	6.2
134	7.0	27.33	12.2			13.8	2.6	8.6
135	4.9	27.31	5.1	36.6	1.2	5.8	0.4	0.6
136	4.0	27.55	2.1			2.4	3.8	2.2
137	5.0	27.55	5.5	36.3	1.3	6.2	2.7	4.0
138	3.0	27.44	0.6			0.7	3.4	0.6
139	4.8	27.37	4.8	38.0	1.2	5.4	0.0	0.0
140	2.5	27.59	0.5			0.6	1.5	0.2
141	4.1	27.58	2.4	41.1	0.7	2.7	3.5	2.3
142	7.0	27.45	12.3			13.9	3.0	10.0
143	6.1	27.35	9.2	43.1	2.6	10.4	0.6	1.5
144	7.0	27.31	12.2			13.8	1.7	5.6
145	4.3	27.33	3.1	32.0	0.7	3.5	3.8	3.2
146	3.0	27.36	0.6			0.7	4.8	0.8
147	5.1	27.17	5.8	30.5	1.2	6.6	1.2	1.9
148	5.0	27.50	5.5			6.2	2.5	3.7
149	4.0	27.60	2.1	36.7	0.5	2.4	3.4	2.0
150	4.0	27.25	2.1			2.4	2.6	1.5
151	1.4	27.32	0.3	29.3	0.1	0.4	−0.4	−0.04
152	1.5	27.79	0.3			0.4	3.1	0.3
153	4.4	27.67	3.5	34.4	0.8	4.0	5.3	5.1
154	5.0	27.45	5.5			6.2	2.6	3.9
155	4.9	27.22	5.1	36.3	1.2	5.8	1.5	2.1
156	7.0	27.39	12.2			13.8	2.3	7.6
157	6.1	27.46	9.2	44.1	2.7	10.4	2.7	6.7
Mean	4.6	27.36	4.9	37.0	1.2	5.9	2.2	2.4
SD	± 1.8	± 0.21	± 4.8	± 4.6	± 1.1	± 6.9	± 2.1	± 2.7

V_{10}: wind speed, measured at 10 m height
$k_t(CO_2)$ and $k_t(O_2)$: CO_2 and O_2 exchange coefficients
PCO_2 and pCO_2: CO_2 partial pressures in seawater and in air, respectively
O_2 and O_2^*: dissolved oxygen concentration and oxygen solubility, respectively
ΔCO_2^a and ΔO_2^a: CO_2 and O_2 fluxes across the air-sea interface
The value of ΔO_2^a in parentheses is omitted in the calculation of the mean

Table 4 shows the calculation of the vertical diffusivity K_z and the vertical diffusional flux F^d. Over the 10-day period, the mean O_2 vertical flux, ΔO_2^d, is 0.7 ± 0.3 mmol m^{-2} d^{-1} and the mean ΣCO_2 vertical flux, $\Delta \Sigma CO_2^d$, is 1.6 ± 0.6 mmol m^{-2} d^{-1}.

6. Interpretation

During the 10-day period, the depth of the mixed layer hardly changed: 24 ± 3 m (Table 1). The O_2 change (ΔO_2) over the depth of the mixed layer is:

$$(0.35 \pm 0.14 \ \mu\text{mol kg}^{-1} \text{d}^{-1}) \times (24 \pm 3 \text{ m}) = 8.6 \ (4.5 \text{ to } 13.5) \text{ mmol m}^{-2} \text{d}^{-1} O_2$$

Likewise, the ΣCO_2 change, $\Delta \Sigma CO_2$, is:

$$-(1.35 \pm 0.6 \ \mu\text{mol kg}^{-1} \text{d}^{-1}) \times (24 \pm 3 \text{ m})$$
$$= -33.2 \ (-16.2 \text{ to } -53.9) \text{ mmol m}^{-2} \text{d}^{-1} \Sigma CO_2 \text{ and } \Delta TA$$
$$= -39.3 \ (-29.3 \text{ to } -46.7) \text{ meq m}^{-2} \text{d}^{-1}$$

Magnitudes and signs of all terms of relations (1), (2) and (18) are known except ΔO_2^b and $\Delta \Sigma CO_2^b$ and we can deduce the production and consumption of O_2 and CO_2 in the mixed layer in the Guinea Dome area. Rearranging expressions (1), (2) and (18) we can write:

$$\Delta O_2^b = \Delta O_2 - \Delta O_2^a - \Delta O_2^d \tag{19}$$

$$\Delta \Sigma CO_2^b = 0.935 \ (\Delta \Sigma CO_2 - \Delta \Sigma CO_2^a - 0.5 \ \Delta TA - \Delta \Sigma CO_2^d). \tag{20}$$

The fluxes escaping from the mixed layer are counted negative and the re-entering ones positive. Finally it is possible to evaluate ΔO_2^b and $\Delta \Sigma CO_2^b$ (day to day changes of dissolved oxygen and inorganic carbon due to biological processes), i.e., the net apparent production of the total system:

$$\Delta O_2^b = 10.3 \ (1.5 \text{ to } 16.5) \text{ mmol m}^{-2} \text{d}^{-1} O_2$$
$$\Delta \Sigma CO_2^b = -13.0 \ (-1.2 \text{ to } -30.5) \text{ mmol m}^{-2} \text{d}^{-1} \Sigma CO_2.$$

The first finding is the expected inverse variation in the ΔO_2^b and $\Delta \Sigma CO_2^b$ changes: the production of oxygen through photosynthesis in the mixed layer is accompanied by the removal of dissolved inorganic carbon during the period of our study. However we notice that the ΣCO_2 quantity removed exceeds the O_2 quantity produced, when in the classic scheme of production and decomposition the production to respiration ratio is -1.3 (Redfield *et al.*, 1963) or -1.4 (± 0.2) (Takahashi *et al.*, 1985). This discrepancy could be attributed to the margin of error in the results due to the uncertainty contained in each term of expressions (19) and (20). The largest errors are without doubt made in determinations of ΔO_2 (4.5 to 13.5 mmol m^{-2} d^{-1}) and $\Delta \Sigma CO_2$ (-16.2 to -53.9 mmol m^{-2} d^{-1}). Given those large uncertainties, one might think that the results derived from our approach are semi-quantitative at best. Yet we are quite

Table 4. Computation of the vertical diffusional fluxes of O_2 and ΣCO_2 (ΔO_2^d and $\Delta \Sigma CO_2^d$) at the bottom of the mixed layer during the 10-day period (29 July–7 Aug. 1986) in the Guinea Dome (12N, 21W)

Station no.	$d\sigma t/dz$	$K_z \times 10^6$ ($m^2\,s^{-1}$)	dO_2/dz (mmol m⁻⁴)	ΔO_2^d (mmol m⁻²d⁻¹)	$d\Sigma CO_2/dz$ (mmol m⁻⁴)	$\Delta \Sigma CO_2^d$ (mmol m⁻²d⁻¹)
119	0.220	2.38	1.25	0.26	10.2	2.10
120	0.023	(22.7)	0.71	(1.40)		
121	0.088	5.94	2.23	1.15	3.4	1.74
122	0.053	9.87	0.85	0.72		
123	0.028	(18.7)	0.58	0.94	1.7	2.75
124	0.114	4.59				
125	0.200	2.62	2.10	0.47	5.8	1.81
126	0.108	4.84	1.03	0.43		
127	0.123	4.25	2.99	1.10	1.3	0.48
128	0.101	5.18	2.50	1.12		
129	0.079	6.62	1.92	1.10	3.2	1.83
130	0.175	2.99	0.49	0.13		
131	0.288	1.82	3.57	0.56	3.8	0.60
132	0.071	7.37	0.80	0.51		
133	0.131	3.99	1.29	0.45	4.0	1.38
134	0.154	3.40	0.45	0.13		
135	0.148	3.53	1.25	0.38	7.2	2.20
136	0.112	4.67	0.98	0.40		
137	0.052	10.1	2.19	(1.91)	4.0	(3.49)
138	0.088	5.94	1.43	0.73		
139	0.332	1.58	3.79	0.52	13.8	1.88
140	0.109	4.80	2.77	1.15		
141	0.143	3.66	2.32	0.73	4.8	1.52
142	0.234	2.24	4.02	0.78		
143	0.150	3.49	2.63	0.79	6.0	1.81
144	0.124	4.22	2.32	0.85		
145	0.233	2.24	4.46	0.86	11.5	2.23
146	0.138	3.79	2.86	0.94		
147	0.245	2.06	4.55	0.81	11.2	1.99
148	0.115	4.55	1.79	0.70		
149	0.250	2.09	4.78	0.86	9.8	1.77
150	0.198	2.64	4.60	1.05		
151	0.126	4.15	2.41	0.86	4.6	1.65
152	0.073	7.16	1.47	0.91		
153	0.166	3.15	3.44	0.94	4.4	1.20
154	0.266	1.97	4.15	0.71		
155	0.377	1.39	5.18	0.62	5.3	0.64
156	0.038	(13.8)	0.63	0.75		
157	0.198	2.64	2.32	0.53	7.0	1.60
Mean	0.149 ± 0.084	4.1 ± 2.1	2.37 ± 1.38	0.72 ± 0.28	6.2 ± 3.5	1.61 ± 0.60

The vertical turbulent eddy diffusivity K_z($m^2\,s^{-1}$) is calculated from $K_z = 0.25\,\epsilon\,N^{-2}$ where turbulent energy dissipation ϵ is chosen equal to $2.10^{-8}\ m^2\,s^{-3}$ (Dillon and Caldwell, 1980) and the buoyancy frequency N is given by the relationship $N^2 = (g/\rho)\,(d\rho/dz)$ with $d\rho/dz = 10^{-3}\,d\sigma t/dz$.

dO_2/dz and $d\Sigma CO_2/dz$ mean vertical gradients of O_2 and ΣCO_2 concentrations at the top of the thermocline.

The values in parentheses are omitted in the calculation of the means.

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Table 5. Budget of dissolved oxygen and total inorganic carbon during the 10-day period in the mixed layer in the Guinea Dome (12N, 21W)

Variable	*In situ* variation over time mmol m^{-2}d^{-1}	Evasion to the atmosphere mmol m^{-2}d^{-1}	Vertical diffusional flux mmol m^{-2}d^{-1}	Biological production mmol m^{-2}d^{-1}
Q	ΔQ	ΔQ^a	ΔQ^d	ΔQ^b
O_2	8.6 (4.5 to 13.5)	-2.4 ± 2.7	0.7 ± 0.3	10.3 (1.5 to 16.5)
ΣCO_2	-12.6 (-2.9 to -28.8)*	-1.2 ± 1.1	1.6 ± 0.6	-13.0 (-1.2 to -30.5)

$\Delta Q = \Delta Q^a + \Delta Q^b + \Delta Q^d$

*For ΣCO_2, ΔQ is corrected from the changes due to precipitation or dissolution of carbonate minerals.

confident in our results which agree with the values of net community production in other tropical areas (Table 6). Minas *et al.* (1986) reported in the Costa Rica Dome a production of 0.14 gCm^{-2} d^{-1}, i.e., 11.7 mmol m^{-2} d^{-1} C. Over a 15-day period in the Sargasso Sea, Langdon *et al.* (1987) measured an increase of the O_2 content of the 20 m mixed layer by 36 mmol m^{-2} for an evasive loss of 122 mmol m^{-2}; such an increase in oxygen content is equivalent to a net community production of 10.5 mmol m^{-2} d^{-1} O_2.

On the time scale of our study (10 days) the O_2 and ΣCO_2 changes in the mixed layer due to biological processes are much greater than the ones due to physical processes (exchange with the atmosphere and vertical diffusional flux from the thermocline) (Table 5). The $\Delta \Sigma CO_2^a$ change is only 9.5% of $\Delta \Sigma CO_2^b$: this ratio is relatively low in an area where CO_2 is largely supersaturated (PCO$_2$ $-$ pCO$_2$ = 37 μatm). The result confirms the conclusion of Johnson *et al.* (1979) that the ΣCO_2 changes in the ocean on a daily basis are primarily due to biological production. The ΔO_2^a change is more important: 22.6% of ΔO_2^b. This result is appreciably different from the 70% ratio of Johnson *et al.* (1979), but these authors dealt with a much larger supersaturation in O_2. The vertical diffusional flux of O_2 as well as of ΣCO_2 is very low (practically an order of magnitude) compared with the biological processes. The calculated vertical eddy diffusivity at the top of the thermocline is indeed very low (Table 4: $K_z = 4 \times 10^{-6}$ m^2 s^{-1}) and typical of the layers with a strong stratification (Denman and Gargett, 1983).

7. Discussion

We previously saw that during the first period of our observations (8 days), close to the center of the dome, the dissolved oxygen content had a tendency to decrease: O_2 decrease = -0.19 ± 0.13 μmol kg^{-1} d^{-1}. During this period we notice that the sea surface temperature increases (Fig. 7) while it does not change later (second period of 10 days). The regression line fitted to the temperature data observed in the morning

Table 6. Comparison of different estimates of production rates from results of PIRAL cruises and from similar studies areas

References	Location	Methods	Results
PIRAL cruise	Guinea Dome 12N, 21W	Net production of O_2 from O_2 change over time	10.3 mmol m^{-2}d^{-1}O_2
—	—	Net uptake of CO_2 from ΣCO_2 change over time	13.0 mmol m^{-2}d^{-1}CO_2
Minas *et al.* (1986)	Costa Rica Dome 8N, 90W	Net community production from Broenkow's oxygen model and Bowden's residence time	11.7 mgat m^{-2}d^{-1}C
Langdon *et al.* (1987)	Sargasso Sea 34N, 70W	Net community production from O_2 change over time	10.5 mmol m^{-2}d^{-1}O_2
PIRAL cruise	Guinea Dome 12N, 21W	Diurnal *in situ* O_2 increase	0.33 μmol kg^{-1}h^{-1}
Johnson *et al.* (1983)	Caribbean Sea 18N, 80W	Net O_2 production during the light period (diel variation of *in situ* O_2)	0.40 μmol l^{-1}h^{-1}
PIRAL cruise	Guinea Dome 12N, 21W	O_2 production from diurnal *in situ* increase of O_2(*)	98.9 mmol m^{-2}(1/2d)$^{-1}$$O_2$
—	—	O_2 production from diurnal *in vitro* increase of O_2	47.1 mmol m^{-2}(1/2d)$^{-1}$$O_2$

(*) The diurnal *in situ* increase of O_2 is corrected for the evasion into the atmosphere and the supply from the vertical diffusion (thermocline).

(07.30 h) has a slope of 0.076 ± 0.017 (°C) d^{-1}, significantly different from 0 ($t = 4.54$ for $n = 8$). This temperature increase involves a diminution in the solubility of the dissolved gases (oxygen — nitrogen — argon). For such a temperature increase the decreases in the solubility of the gases are -0.27 ± 0.06 μmol kg^{-1} d^{-1} for O_2, -0.42 ± 0.10 μmol kg^{-1} d^{-1} for N_2 and -0.013 ± 0.004 μmol kg^{-1} d^{-1} for Ar. During the same time lapse the decrease in dissolved molecular nitrogen concentration is -0.46 ± 0.26 μmol kg^{-1} d^{-1} and that of argon -0.044 ± 0.013 μmol kg^{-1} d^{-1} (Fig. 7). If we take the limits of precision into account the tendency of O_2 concentration in the mixed layer to decrease agrees with the decrease in solubility as do the variations of N_2 and Ar, which are conservative gases. This tendency of dissolved O_2 during this period means that the
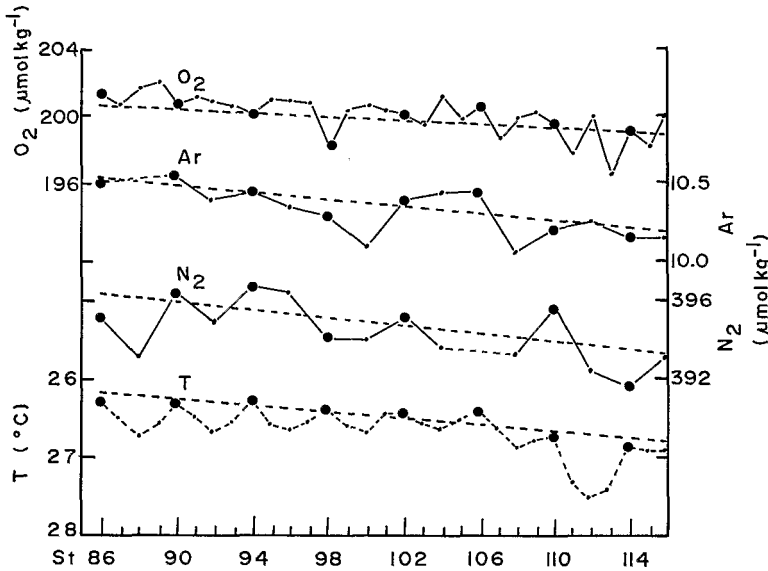
Figure 7. Variations of dissolved oxygen O_2, argon Ar, nitrogen N_2 and temperature T in the mixed layer over the 8-day period at 11°50N, 22W. The broken lines indicate the rates of change in gases, quantities and temperature over the 8-day period: they are built as linear regression equations fitted to the morning station (07.30 h) data (solid dots on the graphs).

net O_2 biological production from day to day is very low then and is masked by evasion into the atmosphere. This conclusion is confirmed by the low increase of *in situ* O_2 between the morning station (07.30 h) and the evening station (18.30 h), normally due to photosynthesis. Over the 8 diel cycles the mean increase of the *in situ* O_2 concentration (between 07.30 h and 18.30 h) is 0.72 ± 0.85 μmol kg^{-1} d^{-1}, i.e. 0.07 ± 0.08 μmol kg^{-1} h^{-1} (according to the results shown in Table 2). During the first period of the observations (close to the center of the dome) the daylight O_2 increase is consequently 5 times less than one (0.33 ± 0.15 μmol kg^{-1} h^{-1}) estimated during the second period of the observations at a short distance away from the center of the dome. We must moreover point out that during the first period the oxygen supersaturation $(O_2 - O_2^*)$ in the mixed layer remained twice as low as during the second period (respectively 1.0 and 2.2 μmol kg^{-1}). Finally we should remember the vanishing of the subsurface maximum of oxygen in the center of the dome, as previously mentioned (Fig. 3). Consequently there are grounds for supposing that the biological production of oxygen could be lower in the center of the dome than at a short distance from it. It is a fact that the chlorophyll biomass in the mixed layer was on average significantly lower during the 8-day period (in the exact center) than during the following period. This feature could be due to a reduction in stability in the center of the dome related to the upward motion of the subsurface layers, rather like the onset of an upwelling (the mixed layer never disappears). In fact the vertical eddy diffusivity at the top of the

thermocline is higher (7.3×10^{-6} m² s⁻¹) in the exact center of the dome than at a short distance away from it (Table 3: 4.1×10^{-6} m² s⁻¹). Some indications could give evidence of more subsurface water in the mixed layer in the exact center of the dome: lower temperature (Tables 1 and 2), higher difference of CO_2 partial pressure in the seawater and in the air (Oudot and Andrié, 1989). Minas *et al.* (1974) pointed out, by comparing the 1959 and 1973 situations in the Costa Rica Dome, that oxygen concentrations and photosynthetic activity are lower when the "upwelling" is stronger and the dome more emphasized (isotherms nearer the surface).

Without going into details of an exhaustive study of alternative methods for determining primary production (which will be the subject of a later paper), we shall just add a few remarks about the comparison of the results of the O_2 changes (Table 6). The diurnal *in situ* increase in oxygen content observed during the 10-day period in the mixed layer is 0.33 ± 0.15 μmol kg⁻¹ h⁻¹: Johnson *et al.* (1983) measured, during a photoperiod in the mixed layer of the Caribbean Sea (18N, 80W), a net O_2 production of 0.4 μmol l⁻¹ h⁻¹, a value which agrees exactly with our observations. Our daylight O_2 increase, corrected for evasion into the atmosphere and supply from vertical diffusion, and converted for the duration of the photoperiod (12 h), leads us to estimate the oxygen production during the photoperiod as 98.9 ± 46.1 mmol m⁻² (1/2 d)⁻¹. Experiments of *in vitro* incubations were also performed during the PIRAL cruise to determine primary production by the standard ¹⁴C method and by the oxygen method. The oxygen release in the light incubation bottles (300 ml) was on average 1.9 ± 0.8 μmol kg⁻¹ (12 h)⁻¹ (Le Bouteiller, pers. commun.), which means a net O_2 production of 47.1 (24.4 to 74.7) mmol m⁻² (1/2 d)⁻¹ over the depth of mixed layer (24 ± 3 m). The apparent production rate based on *in situ* O_2 change is therefore about twice as large as the *in vitro* O_2 change. Recently Fahnenstiel *et al.* (1987) found that in the Great Lakes the *in situ* O_2 changes were twice as great as *in vitro* O_2 changes. The last remark we shall make concerns the low ratio (11%) between the net oxygen production from day to day and that of the daylight period: there is practically an order of magnitude between these two production rates.

The frequency of ΣCO_2 observations (twice a day) is too low to attempt a comparative study of diurnal ΣCO_2 and O_2 changes as made by Johnson *et al.* (1981) and thus to confirm the disparity between O_2 and ΣCO_2 dynamics. Henceforth it will be necessary to connect more closely the ΣCO_2 observations with the O_2 observations and continue to improve the precision of the ΣCO_2 measurements to get a level similar to that of O_2 measurements.

8. Conclusion

Relating physical and biological changes in oxygen and dissolved inorganic carbon in the mixed layer on a daily basis provides a useful tool for estimating a net apparent production of the marine ecosystem, even for oligotrophic oceanic waters. The ΣCO_2-based production (13.0 mmol m⁻² d⁻¹) seems to be of the same order as the

O_2-based production (10.3 mmol m^{-2} d^{-1}). Our results agree well with the evaluation of the "net community production" in other tropical areas (Costa Rica Dome, Minas *et al.,* 1986; Sargasso Sea, Langdon *et al.,* 1987).

On a daily basis the O_2 and ΣCO_2 changes in the mixed layer due to biological processes greatly exceed the ones due to exchanges (evasion in the present case) through the air-sea interface and turbulent diffusion with the upper layers of the thermocline.

The net apparent production of O_2 from day to day is only 11% of the gross diurnal production. The determination of primary production by the *in situ* diurnal oxygen rhythm provides results which are higher than those of classical *in vitro* incubation methods (oxygen method).

Lastly the biological production of oxygen seems to be higher a short distance away from the center of the Guinea Dome than in the exact center, where the hydrological stability of the mixed layer does not seem to be so great.

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