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"Physical Factors Affecting the Cycles of Nutrients & Carbon in the Mediterranean Sea: Influence on the Carbon Mediated Global Change"

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Physical Factors Affecting the Cycles of Nutrients and Carbon in the Mediterranean Sea. Influence on the Carbon Mediated Global Change

by

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ABSTRACT

The striking departure from the mean climatic conditions all over the northern hemisphere during the last years has had a great impact on public opinion. On the other hand, the huge bloom of Chrysochromulina algae in the Skagerrak/Kategatt area in 1988, the frequent blooms of toxic dinoflagellates in shellfish growing areas and the chronic appearance of eutrophication events in the northern Adriatic Sea and other zones of the Mediterranean have alerted on the danger of altering the environment through human activities considered hitherto the key for the improvement of the quality of life through economic development.

One of the environmental factors having a global impact is the well documented increase in atmospheric CO2 concentrations occurred during the last century. Convincing evidence that such a steady state increase, due to the combustion of fossil fuels and the destruction of biomass, carried out by humans, may be leading through the greenhouse effect to a sensible change in the global climate has led governments to launch large international programmes to investigate the various aspects of the Earth system which may have a significant bearing on the climate.

The atmospheric CO2 concentrations happen to be in "equilibrium" with the partial pressure of CO2 in the surface seawater. However, the distribution of dissolved inorganic carbon (DIC) in the ocean is not homogeneous and depends to a great extent on the so-called biological pump. The vertical gradients of DIC and of pH, both factors determining the equilibrium solubility of CO2 in seawater, are to a great extent maintained against advection and diffusion by marine organisms. Photosynthesis produces most of the particulate organic carbon (POC) that after settling to the deep ocean supplies the pool of DIC.

The paper discusses the various factors controlling the carbon cycle, mainly those related to the availability of inorganic and organic nutrients, the production of extracellular organic substances and to the intracommunity exchanges.

INTRODUCTION

One of the environmental factors with a large global impact is the well documented increase in atmospheric CO2 concentrations occurred during the last century due to the combustion of fossil fuels and the destruction of biomass (Figure 1). There is convincing evidence that such a steady increase, is leading through the greenhouse effect to a sensible change in the global climate (Bolin, 1986).

The control of atmospheric CO2 concentrations is effected by a complex combination of processes involving various carbon reservoirs of which the atmosphere is far from being the largest (Figure 2). The biosphere and the soils contain three times as much carbon (2 000 Gt; 1 Gt=10^15 g) as is found in the atmosphere in the form of CO2 (727 Gt in 1983)(Figure 3). The oceans contain more than 50 times as much in the form of dissolved inorganic carbon (DIC) (38 000 Gt) in addition to significant amounts of particulate inorganic carbon (PIC) and of organic carbon, both dissolved (DOC) and particulate (POC). The sediments and sedimentary, rocks considered as the only real sink for carbon, contain 100 000 times as much (72 000 000 Gt).

The atmospheric CO2 concentrations are in "equilibrium" with the partial pressure of CO2 in the surface sea water (Figure 5). However, the horizontal distribution of DIC in the ocean is not homogeneous (Figure 6). Vertical and horizontal oceanic transport mechanisms give rise to geographical and seasonal changes in pCO2 that profoundly affect the exchanges of CO2 across the sea/air interface.

The maintenance of the vertical DIC gradients in the oceans, to a large extent through biological activity, is crucial for the stability of the present atmospheric CO2 concentrations. It is interesting to note that, if the oceans were well mixed, the surface concentrations of DIC would be about 15 % higher which in turn would imply an atmospheric CO2 concentration of about 700 ppm, more than twice the present values of about 325 ppm.

Total inorganic carbon concentrations are higher in the deep sea than in the surface layers. Alkalinity, a combination of all forms of negatively charged carbonic acid ions, has a vertical distribution similar to that of DIC (Takahashi et al., 1981) though with a range of variation considerably smaller (about 30 % of that of DIC).

There is a systematic change in deep sea DIC values rather low in the Arctic seas, higher in the Atlantic Ocean, even higher in the Antarctic Ocean and in the Indian Ocean reaching the maximum values in the older deep waters of the Pacific Ocean. This is the result of the mechanisms responsible for the formation of deep water in the oceans. At high latitudes, wintertime cooling generates mixing to considerable depths, which in limited areas and during short periods of time may extend all the way to the bottom as is the case in the Greenland Sea, in the Weddell Sea or in the NW Mediterranean Sea where formation of deep water takes place with winter cooling and evaporation.

The deep waters formed in this way have relatively low DIC concentrations which correspond to a near "equilibrium" with the atmosphere. Quasi equilibrium between atmospheric CO2 and sea water is established in a matter of days (about 20 according to ....). As these water masses spread over the bottom of the ocean basins, they maintain their original low values though their DIC progressively increases because of decomposition of DOC contained in the original water mass and of POC settling from the layers above. This increase in DIC at the expense of POC and DOC brings about a simultaneous consumption of oxygen concentrations and mineralization of nutrients.

The surface layers of the ocean with high temperatures and low salinities generate vertical gradients due to the influence of these two physical properties on CO2 solubility. However, it is widely accepted that physiological activity of marine organisms is indeed a major factor for the maintenance of these gradients against the diffusive transport of DIC into the surface layers. Photosynthesis in the photic zone, takes up part of the DIC diffused and/or advected from the deeper layers. On the contrary, respiration by heterotrophic organisms, mostly in the top 1000 m of the sea, transforms the organic carbon into DIC. A relatively small amount of detrital organic matter largely in the form of faecal pellets and flocs reaches deeper layers and less than 1 % of the organic material produced through photosynthesis is actually deposited on the ocean floor (Honjo, 1980).

Photosynthesis depends on nutrient and light availability. Low nitrate concentrations often limit the rate of photosynthesis while, at high latitudes, light may be the limiting factor for primary production. The rate of photosynthesis per unit area has large variations from more than 0.5 g C m-2 d-1 in areas of intense upwelling to less than 1% of this value in the oligotrophic regions of the ocean (Figure 6). The largest part of carbon and nutrients fixed in organic structures circulate in the ecosystem until they are metabolized and returned to the dissolved inorganic forms (Figure 7).

Part of the carbon photosynthesized finds its way to the pool of DOC. Recent studies with analytical techniques (Sugimura and Suzuki, 1988) seems to indicate that DOC in surface waters is higher than previously thought by a factor of 3 to 4, while they double previous deep ocean values in good agreement with the apparent oxygen utilization values.

NUTRIENT FLUXES AND THE CARBON CYCLE

The quasy steady state distribution of DIC in the oceans represents an approximate balance between the downward flux of POC mostly with the detritus produced in the upper layers which is then mineralized at greater depths and upward transport by diffusion and advection from deeper layers with greater DIC corcentrations. This cycling of carbon corresponds to the nitrogen cycling scheme proposed by Dugdale and Goering (1967) and by Eppley (   ) since nitrate is required for the production of POC through photosynthesis.

The upward eddy diffusive flux of nitrate from deep water into the euphotic zone is considered to be the primary mechanism for the supply of nitrate to the euphotic layer with atmospheric input and nitrogen fixation traditionally considered small sources (Dugdale and Goering, 1967). Vertical advection is only associated with especially active upwelling areas, mainly in coastal areas and along the Equator. Extremely active eddy diffusion links with advection when large inputs of turbulent kinetic energy brings about the establishment of convective cells.

The vertical nitrogen fluxes in the nearsurface waters of the ocean have been computed by McCarthy and Carpenter (1983). These authors assume that vertical eddy diffusion of nitrate (the same could be applied to phosphate) is the main transport mechanism in oligotrophic zones of the worlds oceans and that both atmospheric supply of nitrogen compounds as well as nitrogen fixation and advection are only of minor importance in oceanic waters.

The vertical flux of a substance N can be assumed to take place by advection and by eddy diffusion (McCarthy and Carpenter, 1983; Duce, 1984). Using the very simple model proposed by Sverdrup et al. (1943) and considering the simplest case of only one dimension in the vertical direction, the flux can be represented by

$$F = w N + Kz \ (dN/dz)$$

where F is the combined flux of N from deeper water
N is the concentration of that substance
w is the vertical velocity of the water
Kz is the coefficient of eddy diffusivity
dN/dz is the vertical gradient of N

Through discretization of the vertical profile of N and with proper assessment of the vertical eddy diffusion coeficient Kz, values of both advection and diffusion terms may be computed. There are various ways in which Kz may be determined taking into account the vertical density distribution, the dissipation of turbulent kinetic energy or the vertical flux of heat and salt (see Varela for a review of techniques).

Duce (1984) critically evaluated the estimates made by McCarthy and Carpenter taking into account the data sets produced by the GEOSECS programme. For their estimates of the nitrate fluxes in the water column, McCarthy and Carpenter used a value of 0.7 mmol m-3 for the concentration of NO3 at a depth of 100 m in the Sargasso Sea. A range of 0.5 - 1.0 mmol m-3 is observed in the GEOSECS (1981) data near Bermuda. For the North Pacific Gyre region McCarthy and Carpenter (1983) use a range from 0.8 (Gundersen et al., 1976) to 1.8 mmol m-3 (Kiefer et al., 1976) observed at 150-175 m. Similar values are observed from the GEOSECS (1982) data. With the values of w given previously, these values result in an advective flux ranging from 0 to 40 umol m-2d-1 in the Sargasso Sea and 10-22 umol m-2d-1 for the North Pacific.

For the calculation of the eddy diffusion flux in the Sargasso Sea region, the nitrate gradient, dN/dz used by McCarthy and Carpenter was 0.013 mmol m-4. A value of 0.021 mmol m-4 is found from the GEOSECS (1981) data in this region. In the North Pacific Gyre, McCarthy and Carpenter used a range for dN/dz of 0.035 and 0.05 mmol m-4 as determined by Gundersen et al (1976) and Kiefer et al (1976) respectively. Somewhat higher values of 0.061 to 0.073 mmol m-4 are determined from the GEOSECS (1982) data.

The eddy diffusion fluxes calculated using the correct range of values of Kz given in table 1 are 22 to 210 umol m-2 d-1 in the Sargasso Sea and 60 to 730 umol m-2 d-1 in the North Pacific. Platt et al. (1982) also estimated a maximum diffusive flux in this region of 800 umol m-2 d-1.

Duce (1984) also reassessed the nitrogen fluxes from the atmosphere into the ocean. Wet deposition rates for nitrate in the Sargasso Sea region are apparently on the order of 10 - 20 umol m-2 d-1. This is a factor of 2 to 3 higher than most estimates of a global background input rate for nitrate in rain, and this is almost certainly related to the higher nitrate concentrations over the western North Atlantic as a result of pollution sources in North America (Duce,) Input rates for nitrate in the Central North Pacific are in the range of 3-6 umol m-2 d-1. Note that there is some suggestion that input rates may be lower in remote regions of the South Pacific.

For the wet deposition of ammonium the input rate to the Sargasso Sea area is approximately 10-15 umol m-2 d-1, a factor of 2 to 3 less than that estimated from the data of Menzel and Aspaeth (1962b). There are few data on NH4 over the North Pacific but ammonium does not appear to be as variable as nitrate, in agreement with the lower anthropogenic source strength for NH3. The wet NH4 input rate in the North Pacific region is likely in the range of 2 to 10 umol m-2 d-1.

The dry deposition of aerosol nitrate is apparently higher over the Sargasso Sea region than over the North Pacific, again due to the proximity of North America. A best guess is that the dry nitrate deposition to the Sargasso Sea is 5 - 15 umol m-2 d-1 while that to the North Pacific is 2 - 6 umol m-2 d-1.

Aeorosol ammonium data over remote regions are very sparse and an estimate of 0.2 - 1 umol m-2 d-1 at both locations is very uncertain. Similarly, there are few data for HNO3 and the estimate of 0.5-2 umol m-2 d-1 at each site is also extremely tenuous and perhaps too high (Prospero et al., 1985). Ammonia measurements are also rare in marine areas and it is uncertain whether the sea is a source or a sink for atmospheric ammonia. The ocean is likely to function both as source and sink depending upon oceanic and atmospheric conditions.

The dry deposition of nitrogen species obtained from our own aerosol measurements at Blanes (Alarcon and Cruzado, 1988; Alarcon and Cruzado, 1989) as well as wet deposition at two locations in southern France and Corsica (Magon et al., 1989) show results comparable to those of Duce (Table --).

The total atmospheric input rate to the Sargasso Sea ranges from 26 to 54 umol m-2 d-1 while that for the North Pacific ranges from 8 - 26 umol m-2 d-1. The fluxes for the NW Mediterranean Sea are much higher, in the order of 300 umol m-2 d-1 as corresponds to locations that are on the shore and they have been considered a major source of eutrophicants in the Skagerrak causing the large Chrysochromulina bloom in 1988.

There are few data on organic nitrogen in rain over the ocean. However, Williams (1967) found 33 % and 40 % respectively of the total dissolved nitrogen to beorganic in rain samples collected off the California coast and near Samoa. Knap et al. (1985) suggest that significant quantities of N in rain near Bermuda are organically associated. Clearly this is a fraction of atmospheric nitrogen which requires considerable additional study.

In very large areas of the world oceans, particularly where strong seasonal thermoclines (or pycnoclines) inhibit penetration of turbulent energy, the surface layers, poor in both phytoplankton and nutrients, are often underlain by layers with relatively important concentrations of chlorophyll pigments bound to the suspended particulate matter (Steele, 1964; Anderson, 1969; Venrick et al., 1973; Kiefer et al., 1976; Velasquez, 1980; Cullen, 1982; Cullen et al., 1983; Herbland, 1983; Herbland and Le Bouteiller, 1983; Estrada, 1984; Takahashi and Hori, 1984). Deep chlorophyll maxima (DCM) and Subsurface chlorophyll maxima (SCM), terms often used in this connection, cover a number of phenomena similar in appearance but probably diverse in the mechanisms producing and maintaining them.

The descriptions that the various authors make of the DCMs and SCMs are far from being coincident. The depth at which chlorophyll maxima are found varies widely and so does the consideration of "deep" by the various authors. While most of the papers published refer to DCM and SCM found below 30 m of depth and almost always above 75 m depth, a few authors have indicated greater depths between 75 and 125 m. Chlorophyll maxima at depths greater than 50 m are often found in tropical and subtropical areas of the sea such as the Gulf of Mexico (Steele, 1964) and the Mediterranean (Estrada, 1984; Cruzado and Velasquez, 1988) or in highly oligotrophic oceanic gyres (Anderson, 1969). However, some of the maxima described occur at much shallower depths between 20 and 50 m (Cullen and Eppley, 1981; Cullen 1982; Cullen et al., 1983) and are closely related to vertical discontinuities caused by superpositioning water layers having different origins.

Since chlorophyll a in the filtered fraction is a good indicator of phytoplankton biomass, DCMs may be related to phytoplankton biomass maxima. However, the fact that DCMs often occur at depths were light intensity may be near or below 1 % of that incident at the surface, makes it difficult to relate the chlorophyll maxima to photosynthesis maxima. Discrepancies are mainly based upon the fact that oxygen maxima are usually found at slightly shallower depths than chlorophyll maxima. The assumption being that as a result of the greater light intensity reaching phytoplankton cells located in the layers above the DCM they may be able to maintain larger levels of photosynthetic activity than those at the DCM.

This however, has a drawback in the existence of low nutrient concentrations, particularly nitrate, above the chlorophyll maximum since chlorophyll maxima are always located at the "nutricline" and more specifically "nitracline", a region in which nitrate concentrations increase from values at or below the 0.1 umol l-1 (often below detection level) to values ranging from 1 to 5 umol l-1. However, not always uptake of nitrate ions is considered enough to maintain photosynthesis and some authors consider that vertical changes in the ratio phytoplankton_carbon/chlorophyll_a may account for some of the maxima (Steele, 1964). Another assumption seldom contested is that nutrient-rich environments allow a build up of chlorophyll_a in phytoplankton cells.

Observations made in the Mediterranean Sea and in the Sargassos Sea, with very close sampling, allow a much finer tuning among the various signals than had been reported before. They show a stable DCM located at 70 m (in the NW Mediterranean Sea), at 100-120 m (in the Sargassos Sea) and even deeper in the Eastern Mediterranean with oxygen and nitrite maxima less than 10 m above it and always linked to the presence of an extremely steep nitracline in spite of the smooth variation of temperature at such depths.

The environmental causes of the DCMs appear to be the great stability of the surface water layers created by a strong thermocline. In regions such as the Mediterranean Sea, away from coastal areas, the only fertilizing mechanism is vertical turbulent diffusion. However, the existence of a pronounced thermocline sets up a barrier to the surface-generated turbulence limiting the upward flux of nutrients into the surface layers.

Photosynthetic organisms grow through the fresh supply of nutrients diffused upwards from the lower layers. In oligotrophic waters they need to be at depths in which a vertical gradient of nutrient concentrations or "nitracline" exists. Since the uptake of nutrients depends on the availability of light as well as nutrients, the narrow layer at which both light intensity and nutrient concentrations are optimally equilibrated is the place where photosynthetic organisms find their best growing conditions and produce a DCM. On the other hand, the nitracline is formed at the depth in which photosynthetic growth is sufficiently intense to take up all the nutrient transported upwards by the turbulent diffusion process.

The phytoplankton population might be inactive when the DCM occurs at depths exceeding those at which the light itensity is well below 1 % of that at the surface values. However, Geider et al. (1985) have shown that, at extremely low light intensity, a number of species may still be capable of maintaining a positive growth rate. Picoplankton species found in the DCM of the Sargassos Sea during the CHLOMAX cruise were very small (0.6 um) protochlorophyceae present in large numbers (100 000 /ml, according to Vaulot, 1988)

In addition, observed vertical motions of the maximum as a whole induced by internal waves have been postulated as a way in which the population can be temporarily exposed to higher light levels. Yet, the changes in light intensity with depth at those levels are insignificant when the vertical motions are small (10 m excursions). Besides, internal waves are not a transport mechanism except if they break.

Oxygen and nitrite maxima are closely connected to the chlorophyll maxima but at somewhat shallower depths. Photosynthesis at the DCM may be promoted by temporary excursions of the organisms to slightly shallower waters, by internal waves. These motions may provide the time needed to deplete the nitrate and other nutrients from the water while producing oxygen and nitrite. The latter is excreted when nitrate is reduced in excess of the amount that can be assimilated by phytoplankton at low light levels (Blasco, 1972) but is taken up again when nitrate becomes lower and sufficient light is available. Once the internal pool of nutrient is exhausted, the cells would sink again to the nutrient-rich waters.

It may be concluded that the DCM is formed by a healthy and active phytoplankton population that gives raise to an important part of the "new" production and maintains through grazing a good deal if not all of the "regenerated" production of the surface layers. It also affects the chemical characteristics of sea water in such a way as to produce a "nutricline" and an oxygen maximum that due to its location is difficult to be exported to the atmosphere. Therefore there seems to be ample evidence that the DCM practically controls the entire pelagic ecosystem in the vast oligotrophic areas of the oceans.

OXYGEN FLUXES

With regard to the oxygen fluxes it is important to note that air/sea exchange processes are not instantaneous (Simpson, 1984) and dense water being formed at the surface by wind action may sink before oxygen has reached equilibrium with the atmosphere.

In order to make a proper assessment of the role of the ocean in the global carbon cycle it is of extreme importance to precisely determine the vertical fluxes of nutrients, carbon and oxygen. The bulk of the ocean water is far from showing any general trend towards increasing nutrient or decreasing oxygen concentrations in the deep layers. This is particularly true in oligotrophic areas of the oceans or, to take an example, in the bulk of the Mediterranean Sea.

Oxygen, like nutrients, is vertically diffused though in the opposite direction. In a recent paper, Bethoux (1989) computed the oxygen, phosphate and organic carbon budgets for the Mediterranean Sea, concluding that "... in the early 21st century, the Mediterranean Sea may begin to present some of the characteristics of the Black Sea and the Baltic Sea, such as having anoxic deep layers". The hypotheses followed in Bethoux computations are:

1. Oxygenation of the water column takes place only through the formation of deep water masses and there is equilibrium between the dissolved oxygen in the dense water being formed and that in the atmosphere.
2. Fertilization of the euphotic zone takes place by upward advection and external inputs (atmospheric, terrestrial, etc.).
3. The particulate organic matter settling away from the euphotic zone is balanced by the "new" production based on upward advection of phosphate.

The uncertainties in these assumptions and the very low precision of the data used in the calculations deserve serious criticism.

The Mediterranean Sea being a concentration basin, the overall vertical water movement is directed downwards since the surface Atlantic water entering through the Straits of Gibraltar, estimated at 41.6 10^{12} m3/yr (after making allowance for a 6 % of this amount that evaporates) flows downward and back into the Atlantic Ocean. Irrespectively of local upwellings, downwellings and convection cells, the average downwards flux of water is thus of about 16.6 m/yr (the area of the Mediterranean Sea is about 2.5 10^{12} m2). As a consequence, vertical advection is not a general fertilizing mechanism although it may contribute to transport some oxygen from surface to deeper layers.

These features are found all over the Mediterranean Sea during most of the year except in very few areas affected by upwelling (Tintore et al., 1988), convection (particularly in winter and early spring) or discharges from rivers. The extension of all Mediterranean areas relatively well fertilized, according to the pigment levels in the spring bloom, the most productive time of the year, is not more than about 5 % of the basin surface (Fig. 1).

Leaving aside for a moment the nutrient inputs from external sources and the losses to the Atlantic Ocean, the nutrient/carbon/oxygen balance can be represented by a one-dimension system.

The vertical distribution of nitrate and oxygen as well as other hydrographic parameters in two deep stations of the NW Mediterranean Sea (Fig. 2) are shown in Figure 3. One of the stations (MA/ZA 3) is right in the area in which winter deep water formation takes place and the other (MA/ZA 1) in the middle of the western basin, in a very stable area due to the low salinity and relatively high temperature of the Modified Atlantic Ocean Water present in the surface layer. This fact is reflected in the extremely homogeneous distributions of temperature, salinity and density of the former which contrast with those of the latter. Surface nitrate values also indicate a large difference in the vertical extent of mixing in the surface mixed layer of the two stations.

Table I shows values computed for the upward eddy diffusion fluxes of nitrate and "new" production of particulate organic carbon in the two stations in comparison with the downward eddy diffusion fluxes of oxygen and respiration of organic carbon at the oxygen minimum layer. The fluxes have been computed taking into account the vertical gradients of nitrate and oxygen and the eddy diffusivity values of 1.7 and 8.5 m2/d (0.2 and 1 cm2/s) used previously by McCarthy and Carpenter (1983), Duce (1986) and Cruzado and Velasquez (1986) among others and the Redfield molar ratio of C:O2:N of 106:138:16.

The nitrate fluxes range from 53 - 265 umol NO3/m2 d in the more homogeneous station to 100 - 510 umol NO3/m2 d in the offshore station. Such a difference is probably showing the weaker uptake in the MA/ZA 3 station with a deeper mixing depth than that in the more stable MA/ZA 1 station.

The production of particulate organic carbon which may be sustained with these nitrate fluxes ranges between about 4 - 20 mg C/m2 d (1.5 - 7.5 g C/m2 yr) at the MA/ZA 3 station and 8 - 40 mg C/m2 d (2.9 - 14.5 g C/m2 yr) at the MA/ZA 1 offshore station. These are much lower values than those mentioned by Bethoux for the "new" primary production and correspond better with an oligotrophic system.

The vertical distribution of oxygen in the two stations in Figure 3 shows the existence, in the top 100 m layer, of undersaturated values (85 - 90 %), only slightly above those of the deep waters at station MA/ZA 3, in the deep water formation area and around saturation (99 - 102 %) at the offshore station MA/ZA 1. The deep waters have values above 80 % saturation and an oxygen minimum zone is found at depths between 200 and 400 m, in both stations, with values not lower than 70 % saturation.

The values computed for the eddy diffusion fluxes of oxygen in both stations (Table I) indicate that such a flux would be more than sufficient (130 %) for the respiration requirements at the oxygen minimum layer in the deep-water formation station MA/ZA 3, assuming all the newly formed particulate organic matter would be respired in this layer, while it would only cover a small part of such requirements (23 %) in the more fertile offshore station MA/ZA 1. Since respiration of particulate organic matter takes place while detrital matter settles through the water column, the layer affected by respiration is thicker than the light controlled nutricline and, therefore, the oxygen gradient is smaller than that computed for nitrate.

Therefore, an important part of the oxygen requirements in the oxygen minimum layers (300-1000 m depth), in which most of the respiration of the newly formed organic matter takes place, is provided through eddy diffusion. This mechanism has a nice feature, that the larger the oxygen consumption is in the lower layers, the larger is the vertical gradient and thus the flux of oxygen.

Atmospheric nutrient inputs, an important source of particulate organic matter, have been neglected in the phosphate budgets computed by Bethoux (1981, 1989) as well as by Coste et al. (1988). Photosynthesis, based on nutrients deposited, takes place in the surface layer and part of the oxygen produced is probably lost to the atmosphere. If deposition values determined at coastal stations in Cap Ferrat and Corsica (Migon et al., 1989) or in Blanes (Alarcon and Cruzado, 1988) were representative of the open sea, this source would provide an important fraction of the nutrients required to support the "new" production, comparable to those supplied by diffusion.

A secular increase in production through enhanced atmospheric nutrient deposition would increase the oxygen consumption in the intermediate layers (300 - 1000 m depth) making the oxygen minimum stronger and correspondingly increasing the oxygen gradient (surface and subsurface values are always at or above saturation) and thus the diffusive flux of oxygen.

Before any sign of anoxia can appear in the deep layers of the Mediterranean Sea, a strong minimum oxygen zone and much higher nutrient concentrations at intermediate depths should appear, followed eventually by denitrification, like in the Peru upwelling system (Packard et al., 1983) and other highly fertile regions. The oxygen minimum zone observed in parts of the Alboran Sea (Packard et al., 1988) is more pronounced than in the rest of the Mediterranean Sea as corresponds to the dynamic effect of the jet stream entering from the Atlantic Ocean through the Straits of Gibraltar.

On the other hand, nutrients discharged from terrestrial sources are not necessarily going to increase the deep water dissolved nutrient pool. Those nutrients incorporated into the particulate organic matter settle to the bottom, over the shelf, before the water mass has had time to reach the open sea (Cruzado and Velasquez, 1989). Longshore currents such as the Algerian current or the Liguro-provencal current are good traps for particulate matter settling onto the shelf. Turbidity currents from the slopes may of course export resuspended particulate organic matter to the deep sea but their role as a transport mechanism for nutrients into the deep waters of the Mediterranean has not been properly documented.

Monitoring the time change of deep water phosphorus concentrations is very difficult due to the large imprecision of the dissolved inorganic and organic phosphorus measurements. Nitrate determinations are much more precise and, even so, temporal changes in nutrient concentrations either in the deep layers or in the nutrient maximum layers, or changes in the depth of the nutricline or the nutrient maximum, even with the best available analytical techniques cannot be precisely determined (even worst at such low orthophosphate concentrations found in the Mediterranean waters). Even in other areas of the oceans with higher nutrient levels, such calculations are meaningless in view of their low precision.

In conclusion, nutrients, oxygen and carbon biogeochemical cycles are only known with a large degree of inaccuracy. The areal distribution of the vertical fluxes of these elements is practically unknown for large areas of the Mediterranean Sea. Most of the basin has nutrient distributions typical of the most oligotrophic regions in the world. The capacity for oxygen to diffusively flow downwards to cope with the respiration requirements is at times larger than the fertilizing nutrient fluxes.

To balance the nutrient losses through Gibraltar with the nutrient inputs from terrestrial and atmospheric sources is not easy with the present degree of measurement accuracy. Therefore, the nutrient reservoir of the Mediterranean Sea could equally well be decreasing or increasing.

Earlier voices claiming that the Mediterranean Sea was dying (Cousteau and Paccalet, 1987) have been suffocated by serious scientific work undertaken in the last decades. Efforts are being made by individual teams to study, both in the western and eastern basins the biogeochemical cycles of substances, particularly nitrogen, carbon, oxygen and other elements which may have a bearing on these ones. However, available resources cannot cover in depth the problems identified today. It is therefore urgent that Mediterranean and non Mediterranean scientists undertake a concerted action attracting to this region the interest of international bodies such as JGOFS. The future evolution of the Mediterranean Basin is too important to be taken lightly.

EUTROPHICATION

Eutrophication, meaning "enhanced nourishment", refers to the stimulation of aquatic plant growth by nutrients, particularly the combined inorganic forms of phosphorus, nitrogen and to a lesser extent of silicon (UNEP, 1990). It is usually the result of introducing large amounts of inorganic and organic nutrients in the aquatic environment. The processes that concur to its existence are fundamentally natural though the result may be an ecosystem quite far from natural equilibrium when its causes are anthropogenic. Many areas of the worlds oceans are naturally eutrophic due to intense energy inputs bringing about greater fertilization of the surface waters than average. Upwelling, whether coastal or open sea, generates such eutrophic conditions. Many estuaries, because of the nutrients and organic matter discharged by the rivers, are naturally eutrophic.

Algal blooms, qualified as exceptional when they are noticeable, particularly to the general public directly or indirectly through their effects are often associated with eutrophication. They are common in coastal and estuarine areas as well as in open seas in the temperate regions (Fisher et al., 1988; Valiela and Costa, 1988). However, because of recent changes in most rivers of the water flows and sediment and nutrient loads, eutrophication related algal blooms are now occurring in areas previously not affected.

Eutrophication is seen by the GESAMP in its latest report on The State of the Marine Environment as the most important threat caused by pollution and second to "Coastal and Hinterland Development" (UNEP, 1990). This change in appreciation since earlier reports prepared by the same group has been the consequence of dramatic algal blooms that have occurred during the last years in various coastal areas and enclosed seas around the world not to mention coastal lagoons (UNESCO, 1984) and inland lakes and reservoirs (OECD, 1982).

The almost permanent algal blooming of the North Adriatic waters, particularly along the coast of the Emilia-Romagna region (Rinaldi and Montanari, 1988), the algal bloom occurred in the Skagerrak during the spring of 1988 (Barth and Nielsen, 1989) and the occurrence of mucillagenous materials in the North Adriatic Sea (Rinaldi et al., 1990; Herndl and Peduzzi, 1988; Herndl, 1988) and in the North Sea and English Channel (Lancelot et al., 1987) are all serious eutrophication events deserving close attention. Embayments such as Saronikos Gulf, Ismirn, Iskanderun and Alexandria bays, in the Mediterranean (UNEP, 1989; 1990), not to mention inland seas such as the Black and the Baltic seas (Lancelot et al., 1990), the Sea of Marmara and the Sea of Japan, suffering from chronic dystrophia but often considered to be in their normal state.

The present paper discusses the major processes that constitute the direct causes of coastal eutrophication, eventually leading to dystrophia, some of which are amenable to reduction by proper management of the systems.

Processes Controlling Eutrophication

Eutrophication phenomena are always related to discharges of nutrient laden freshwater either from rivers or sewers nourishing dense algal and microbial populations (Marchetti et al., 1989; Halim et al., 1976; Muller-Karger et al., 1988). In these areas nutrients and organic matter, controlled by their influxes as well as by local physical and biogeochemical processes, establish a delicate balance with oxygen that can be upset whenever meteorological or climatic conditions limit vertical mixing, eventually producing catastrophic events (Barth and Nielsen, 1989).

The aquatic ecosystems may be seen as a two-component machine powered by light and nutrient (Figure 1):

1. An upper layer, receiving the sunlight, in which photosynthetic biomass production dominates over respiration. Oxygen is produced and nutrients taken up in roughly proportional amounts to the biomass produced.

2. A lower layer, in the dark, in which respiratory destruction of biomass dominates. Oxygen is consumed and nutrients are regenerated in proportional amounts to the biomass consumed.

In oligotrophic systems (tropical and subtropical oceans and thermally stratified temperate seas), the nutrients flow to the euphotic layers from below mostly through turbulent diffusion and give rise to a nutrient limited ecosystem that often shows a biomass maximum below the thermocline. An oxygen maximum is produced at approximately the same depth (50 to 120 m). In winter time temperate seas, vertical mixing is strong and brings nutrients to the upper layers. However, because of it, the autotrophic populations do not have time to produce enough biomass to maintain themselves and nutrient-rich surface water exist temporarily giving rise to a bloom at the surface (at the onset of spring when thermal stratification begins) that soon develops into a more oligotrophic system. When this happens, oxygen comes close to equilibrium with the atmosphere.

The net biomass produced in the upper layer settles through the lower layers and onto the sediments where nutrients are regenerated and reconstitute, in non-eutrophic conditions, the deep-water nutrient pool. Since oxygen is produced mostly below the thermocline, the upper layer is in equilibrium with the atmosphere and therefore most of it remains in the water column. Due to advection and vertical turbulent diffusion the oxygen produced flows down to the oxygen consuming layers (minimum oxygen zone) and no deficit is produced. The cycle closes with the upward flux of inorganic nutrients in amounts equivalent to the downward flux of oxygen. Recycling also takes place in the euphotic layer without changing the net fluxes.

In eutrophic conditions, nutrients are added from external sources (terrestrial). Biomass is produced in the uppermost layers (a very shallow euphotic zone), particularly if, as usual, freshwater is the carrier of these nutrients. Oxygen is also produced in the top layers but, in this case, in large amounts causing the water to be oversaturated with respect of the atmosphere and a net loss of oxygen takes place. Biomass settles through the lower layers in excess of the upward nutrient flux. Actually, the net nutrient flux is often in reverse (from the upper onto the lower layer). Downward oxygen flux is hindered by strong stratification and it never matches the downward flux of biomass. As a consequence, respiration would exceed oxygen supply were it not by lateral advection that brings oxygenated water into the eutrophic region.

The lower layer and the sediments become an additional source of nutrients for the upper layer when, during a bloom, the surface nutrients become exhausted contributing to the maintenance of the blooming community.

The above machine is extremely active in eutrophic ecosystems and may be dislocated by a conjunction of environmental factors resulting in consequences undesirable from many points of view. In particular, it may reduce (in spite of a greater primary production) the fish stocks and the ammenities through the turbidity of the waters and fouling of the sediments. The area may eventually go anoxic, causing fish kills and smell H_2S and mercaptans.

A number of physical, chemical and biological processes participating in this basic cycling contribute to the appearance and intensity of eutrophic conditions (Cruzado, 1974; 1985). First of all is the excessive nutrient loads (both inorganic and organic) in areas with restricted water circulation. Second, is a limited rate of turnover of the water column seriously restricting the dispersion of the nutrients and plankton populations. Third comes the availability of oxygen for respiration in deep water layers (Frascari, et al., 1979; Frignani et al., 1978; Wassmann, 1986). Other processes may also contribute substantially to the degree of eutrophication. In the following sections, some of them, deemed to be important, are discussed.

Circulation of water in coastal areas

An important condition for the existence of eutrophication is the persistence without mixing with neighbouring waters of nutrient rich freshwater. Enclosed sea areas, particularly estuaries and bays, receiving important amounts of freshwater from rivers, sewers or runoff are prone to develop eutrophic conditions. Water exchange with the open sea through boundaries or openings depends on various factors such as tidal cycles, wind events, etc.. Accumulation of organic matter within the sediments takes place in areas with restricted exchange with the open ocean. Sills at the entrance of marginal seas contribute to the maintenance of two water layers with very different densities. The Black Sea, the Baltic Sea or the Norwegian fjords are examples of two-layered systems in which the deeper water is seldom exchanged with the open ocean (Figure 2).

Open coastal areas may also have restricted exchanges with the open sea due to the existence of hydrodynamic fronts and eddies (Cruzado et al., 1990). Current systems, if properly directed, may act as barriers for dissolved and/or suspended matter. Shelf areas, such as the Gulf of Lions in the NW Mediterranean Sea, behave as a trap not only for terrigenous nutrients but also for those having a marine origin (Cruzado and Velasquez, 1989). Longshore advection of water trapped between the coast and slope currents (Figure 3) does not disperse the nutrients and organisms efficiently (Cruzado et al., 1990). Therefore, the chances of having dense biomass concentrations are not reduced by horizontal circulation. On the contrary, horizontal recirculation, producing an entrapment of the nutrients and organisms responsible for biomass production, increases these chances by supplying seeds for phytoplankton development (Cruzado and Velasquez, 1990).

Mixing of water layers

The surface layer in the sea is stirred by wind to depths depending on both wind force and water stratification. The surface mixed layer, normally extends down to a depth of about 10 meters but may be as thick as 40 m or more in winter conditions, or even limited to a few centimeters in a river plume (Figure 4).

Vertical mixing between surface and subsurface waters favours the exchange of dissolved and particulate materials and therefore their dispersion. Wind- or tide-generated turbulence promotes mixing, bringing to the lower layers the oxygen required for the respiration of the organic matter either locally produced or discharged by rivers. On the contrary, stagnant conditions may lead to the onset of a very strong pycnocline that separates the two layers, promoting the development of very dense planktonic populations and hindering the transport of oxygen to the lower layers.

Onset of anoxic conditions is avoided by the wind-forced ventilation of the deep layers. Surface water, under stormy conditions in shallow areas or in very cold weather events, may penetrate down to the bottom. Anoxia takes place only with calm winds lasting for some time.

The deep ocean can hardly be anoxic because of the low fraction of the surface produced organic matter that reaches the bottom layers and of the deep water renewal by ventilated deep water formed in high latitude areas. Deep basins, only if the bottom layers have restricted circulation because of strong haloclines (some fjords and enclosed seas such as the Black Sea, Baltic Sea, Cariaco Trench, etc.), may have large bodies of anoxic water due to lack of deep water renewal coupled to large rates of organic matter falling down. If deep circulation is not restricted but the amount of organic matter produced in the euphotic zone is large (Southeast Pacific, Alboran Sea, etc.), minimum oxygen layers develop over large areas of the oceans (Packard et al., 1985).

On the contrary, shallow systems have a strong coupling between the water column and the bottom communities, the latter acting as a reservoir for organic as well as inorganic nutrients. Large shallow areas such as the North Adriatic Sea show important restrictions to the inflow of open sea water, thus to dispersion of the freshwater.

Light

Light availability is "the" factor controlling photosynthesis in the ocean. Vertical mixing in the open ocean brings the organisms down to depths in which there may be insufficient light intensity for their sustainable growth. Blooming of plankton organisms can only take place if the depth of the surface mixed layer is above the critical depth at which algal photosynthesis and respiration compensate each other.

Turbidity, generated by mineral particles, detritus and organisms may limit photosynthesis to only the upper few meters and heavily hinder the nutrient utilization and oxygen production in the lower water layers. Therefore, net oxygen consumption may take place in the largest part of the water column even in shallow areas.

Sedimentation

Rates of sedimentation and nature of the settling particles also play a very important role in the control of eutrophication. The concentration of organic matter in sediments may be kept low in coastal areas, by the continuous or episodic accumulation of mineral materials or by the rapid growth of shell bearing molluscs in shallow environments. When the sediments are rich in organic matter, they are the support of a large microbial community, whose respiratory activity may produce anoxia and hydrogen sulphide typical of highly eutrophic environments.

Watershed management in many river basins has led to important reductions in the amount of mineral particles covering the organic detritus in coastal areas. On the contrary, deforestation and soil erosion may cause a decrease in the trophic potential of some third world estuarine areas.

Nutrients

Sources of nutrients for coastal areas are diverse (rivers, sewage disposal, agriculture, stock rearing, industry, atmospheric deposition, etc.) and their importance varies from place to place. Rivers provide to the coastal areas large nutrient loads (inorganic and organic) as well as fresh water that leads to the stability of the water column. This sets the ground for a particular structure of the pelagic and benthic communities leading to eutrophication of estuarine systems (Mahoney, 1989).

Vertical advection of nutrients through upwelling, generated either by frontal dynamics or through interaction between currents and the bottom topography, is the exception rather than the rule and only occurs in a relatively small number of locations around the worlds oceans.

Atmospheric nutrient inputs, an important source for the production of particulate matter, have been neglected until recently (Duce, 1986). Photosynthesis based on deposited nutrients takes place in the surface layer and part of the oxygen produced may be lost to the atmosphere. If deposition values determined at coastal stations were representative of the open sea (Migon et al., 1989; Alarcon and Cruzado, 1988;1989) this source would provide an important fraction of the "new" production, comparable to that supplied by eddy diffusion (McCarthy and Carpenter, 1984).

The vertical fluxes of nutrients, taking place in most parts of the ocean by eddy diffusion, are very weak only supporting low biomass densities in the upper layers (Cruzado and Velasquez, 1988). Most of the diffusive fluxes take place at the "nutricline", a zone of steep gradient, linked to the maximum penetration of light and to the turbulent kinetic energy available for eddy diffusivity. For similar eddy diffusivities, the steeper the nutricline slope is, the greater will be the vertical nutrient flux (McCarthy and Carpenter, 1984). When eddy diffusivity is the only mechanism, such a flux gives rise to a deep chlorophyll maximum at the top of the nutricline (Varela, 1991). In two-layered systems, diffusion through a strong pycnocline is heavily reduced. Only wind stirring may bring about important nutrient exchanges across such interfaces.

The composition of the surface water (nutrients, organic matter, trace metals, vitamins, etc.) is of key importance in determining the potential for eutrophication. The nutrient load and the chemical and biochemical oxygen demand of the waters discharged onto the sea are extremely important in this regard. The amount and composition of the organic matter dissolved and/or dispersed in the fresh water discharged determine the oxygen requirements of the microbial communities that metabolise it.

Dissolved nutrients and organic matter remain in the pelagic domain until they are taken up by pelagic organisms and included in their own body or aggregated into large particles through biological or physicochemical processes. Nutrients and organic matter found in the particulate fraction settle, disappearing from the pelagic domain. However, such disappearance is only transitory since the particles may be resuspended by storms and dissolved nutrients and organic matter in interstitial water diffuse back into the water column after biochemical transformations in the bottom sediments. Therefore, allochtonous as well as autochtonous organic matter promote the growth of local microbial populations in the sediments.

The redox potential in the water column and in the sediment pore waters, linked to the concentration of organic matter and microbial activity, determines whether the system will be fixing nitrogen or losing it through denitrification. It is now accepted that highly eutrophic areas lose organic nitrogen through denitrification while not so eutrophic marshes may be important sources of nitrogen.

Our understanding of how changes in N-loads affect N cycling in these systems is qualitative at best. Low N-loads tend to produce systems in which biomass is dominated by benthic plants and their associated predators, whereas high N-loads favour dense plankton concentrations and a benthic community dominated by filter feeders. The pattern of N cycling in these two types of system differs greatly.

Phosphate, reputedly responsible for most eutrophication phenomena in fresh water and suggested by many as responsible in marine eutrophication as well may not be the key issue in view of both the phosphate requirements of planktonic organisms and of the geochemical processes that regulate phosphate concentrations in fresh and seawater. The present believe is that phosphorus geochemistry allows for exchanges to take place with the sediments, particularly those in the lower river bed, rich in calcium and magnesium carbonates and in iron oxides (Kelderman, 1988).

Phosphate concentrations, relatively lower than those of nitrate and ammonium, may determine the extent of phytoplankton growth, thus controlling the maximum algal densities attainable in a given water parcel. It should, however, be noticed that this role is different from that of controlling the overall production of biomass in a given area, determined by a combination of factors of which the most important ones could be nitrogen loads and dispersion.

The disequilibrium in the relative nutrient concentrations has been held responsible for the occurrence during the last years of mucillagenous materials in parts of the North Adriatic Sea, especially along the Emilia-Romagna region (Rinaldi et al., 1990) but not limited to this area as seen from remote sensing (ESA, 1989). In spite of the fact that the occurrence of algal blooms and other such phenomena referred to as the "mare sporco" are not new in the North Adriatic (Fonda Umani et al., 1989), the frequency of occurrence of such phenomena in recent years gives weight to the fear that change in climate, taking place at the global scale, might be promoting their occurrence (Houghton and Woodwell, 1989; UNEP, 1989). This question, however, is still rather speculative and should not provide justification for delays in active control of nutrient loads in the river Po.

Photosynthesis

The major consequences of eutrophication are observed both at the sea bottom (high organic matter contents in the sediments, lack of oxygen, very poor bottom fauna, etc.) and in the pelagic domain (phytoplankton blooms, high oxygen concentrations in surface layers and low in bottom layers, high turbidity, etc.).

Whether inorganic nutrients or nutrient-containing organic matter are introduced in the marine environment, the result is an enhancement of the cycle shown in Figure 5. The left to right process takes place within the surface layers with overproduction of oxygen which partly remains dissolved in the water and partly is expelled to the atmosphere. The right to left process takes place in the low layers and in the sediments pore water where oxygen consumption often exceeds the flux of oxygen brought in through diffusion and advection.

Algal blooms are an important part of eutrophication. They develop as a consequence of rapid reproduction of phytoplanktonic organisms (algae, flagellates, photosynthetic bacteria, etc.) unbound while nutrients are available in the water parcel in which the population grows.

In some cases, there is evidence of competition for nutrients between phytoplankton and seaweeds (for example in the Venice Lagoon). Turbidity and sedimentation rates may be, with nitrogen availability, the most important parameters to decide which community takes over the other (Sfriso et al., 1988).

The initial concentration of nutrients in the water determines the final biomass density of a phytoplankton bloom although various factors may affect the stoichiometry of such a relationship (Figure 6). The departure of the nutrient concentrations from the straight line defined by the mixtures of freshwater (salinity Sfw and nutrient concentration Nfw) with seawater (salinity Ssw and nutrient concentration Nsw) indicates the biomass each water parcel has produced in the immediate past. On the other hand, the nutrient concentrations observed in each salinity zone stretching from the river mouth to the open sea indicate the potential production of the zone.

The "blooming potential" depends also on the time spent by a water parcel in the low-salinity zone. Well mixed estuaries like that of the Rhone river do not produce large blooms because salinity changes rapidly with time as a water parcel flows through the plume. Most of the biomass is produced at salinities above 30 psu (4/5 Ssw) when the nutrient concentrations are less than the theoretical mixing value of 1/5 Nfw.

On the contrary, enclosed estuaries or highly stratified seas may retain the water parcels for a long time before they are mixed with nutrient poor surrounding seawater. In this case, the density of the biomass being formed is much larger. For example in the Sacca di Goro (one of the Po river mouths) chlorophyll a values above 140 ug/l correspond to a salinity about 11 psu and nitrate concentrations of about 7 ug-at/l (29 june 1988, st. 39; Regione Emilia Romagna, 1988). However, for a salinity of 17 psu, the chlorophyll a was only 3 ug/l when the nitrate concentration was greater than 100 ug-at/l (3 december 1987, st 39).

On one hand, all the factors contributing to the export of nutrients away from the phytoplankton population (dispersion, grazing, settling, etc.). On the other, all those that contribute to the in situ production of nutrients (excretion, nitrogen fixation, remineralization of organic matter, etc.).

Several of these factors act in combination (e.g. grazing and excretion). Turbulent motion causes both diffusion of nutrients and dispersion of phytoplankton cells. However, since the gradients of both parameters are usually inverse, phytoplankton dispersion out of a patch is concurrent with nutrient diffusion into the patch contributing to the maintenance of the bloom.

Respiration

The right to left process in Figure 5, requires oxygen. Oxygen is a gas and should, in principle, be in equilibrium with the atmosphere. However, the oxygen cycle is linked to the carbon and nutrient cycles and oxygen is produced by photosynthesis in the euphotic zone and consumed by respiration in the entire water column and in the sediments.

The normal distribution of oxygen concentrations in the sea shows a maximum at or near the surface, a minimum at intermediate depths (300-1000 m) and greater values (always below saturation) in the deep water. Oversaturation is often observed in eutrophic and non eutrophic systems alike. This is particularly true in the surface mixed layer during the spring phytoplankton bloom in temperate seas and under the thermocline where the phytoplankton biomass is present during summer stratification periods and in tropical and subtropical seas.

When oversaturation occurs at the surface, oxygen tends to escape to the atmosphere through the air/sea interface. If oversaturation occurs below the thermocline, oxygen diffuses up and down. Actually, very little oxygen is lost to the atmosphere in oceanic oligotrophic systems because the oversaturated water does not normally establish contact with the atmosphere and the gradient between the concentration at the oxygen maximum and those at equilibrium (surface water) is relatively small. On the contrary, in eutrophic systems the amount of oxygen lost to the atmosphere may be very important due to the high activity of phytoplankton blooms and the difficulty with which oxygen is transferred from the upper well-lit water layers to the lower dark layers.

Undersaturation may occur at different depths and in the interstitial waters of the sediments. In particular, an oxygen minimum zone occurs at intermediate depths caused by the metabolic activity of microorganisms bound to slowly settling detrital matter. The intensity of the oxygen minimum depends on the amount of organic matter falling down from the euphotic zone. Oxygen flows into the oxygen minimum layer by diffusion from above and from below at a rate sufficient (at steady state) to compensate for the metabolic consumption. A non-negligible amount of oxygen (actually equivalent to the "new" production) is slowly diffusing downwards, more important the greater the vertical oxygen gradient is. Oxygen consumption under upwelling areas is well known to bring about undersaturation and even partial anoxia in open ocean areas or total anoxia such as in the Cariaco Trench where pelagic production of organic matter supported by upwelling is complemented by stagnation in the deep layers.

The values computed for the eddy diffusive fluxes of oxygen in two stations in the NW Mediterranean Sea (Table I) indicate that such a flux would be more than sufficient (130 %) for the respiration requirements at the oxygen minimum layer in one station (MA/ZA 3), assuming all the newly formed particulate organic matter would be respired in this layer, while it would only cover a small part of such requirements (23 %) in another, more offshore, station (MA/ZA 1). Since respiration of particulate organic matter takes place while detrital matter settles through the water column, the layer affected by respiration is thicker than the light controlled nutricline and, therefore, the oxygen gradient is smaller than that of nitrate.

An important part of the oxygen requirements in the oxygen minimum layers (300-1000 m depth), in which most of the respiration of the newly formed organic matter takes place, is provided through eddy diffusion. The larger the oxygen consumption is in the lower layers, the larger is the vertical gradient and thus the flux of oxygen.

Excessive consumption of the oxygen dissolved in the water column, in cases of extreme eutrophication, causes anoxia to develop at the bottom and sometimes in the entire water column. In such cases, Dystrophia, a term used to define these ecological conditions, is characterized by a surface layer loaded with photosynthetic organisms underlain by a bottom layer devoid of oxygen causing the death of bottom living organisms and sometimes of the entire fauna (Cruzado, 1978).

Before any sign of anoxia appears in the deep layers, a strong minimum oxygen zone and very high nutrient concentrations at intermediate depths exist, followed eventually by denitrification, like in the Peru upwelling system (Packard et al., 1983) and other fertile regions. The oxygen minimum zone observed in parts of the Alboran Sea (Packard et al., 1988) is more pronounced than in the rest of the Mediterranean Sea as corresponds to the dynamic effect of the jet stream entering from the Atlantic Ocean through the Straits of Gibraltar.

A secular increase in production through enhanced atmospheric nutrient deposition would increase the oxygen consumption in the intermediate layers (300-1000 m depth) making the oxygen minimum stronger and correspondingly increasing the oxygen gradient and thus the diffusive flux of oxygen.

In conclusion:

1. Nutrient loads in estuarine and coastal areas are the main driving force for eutrophication. Inorganic and organic nutrients stored in the sediments provide an important part of the nutrient inputs to the euphotic zone.

2. Phytoplankton populations develop denser the higher the initial nutrient concentrations in the water and the longer the persistence of this water unmixed in the euphotic zone.

3. The nutrient controlling the maximum density attainable by phytoplankton in a water parcel is that with the lower relative concentration, namely phosphate in some estuarine areas.

4. The nutrient controlling the maximum biomass produced over a given area is probably nitrogen in most estuaries.

5. Oxygen transport from the top highly productive layers to the bottom consuming layer is mainly through diffusion. Anoxia develops when stratification cannot be broken by weak wind regimes.

6. Dispersion of nutrient rich freshwater by mixing with surrounding nutrient poor seawater decreases the potential for eutrophication. On the contrary, recirculation increases the chances of algal blooms being formed.

7. Wind or tidal mixing decreases the potential for acute eutrophication phenomena. Longshore transport does not effectively mix the water and therefore does not decrease such a potential.

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VERTICAL NITRATE FLUXES AT THE BASE OF THE EUPHOTIC ZONE IN OLIGOTROPHIC AREAS OF THE OCEANS

LOCATION	GRADIENT mmol m-4	DEPTH RANGE m	FLUXES(*) umol m-2 d-1	SOURCE
Sargassos Sea			22 - 135	(1)
North Pacific Gyre			60 - 520	(1)
Costa Rica Dome			380 - 1760	(2)
Trop. N. Pacific			180	(3)
SW Sargassos Sea	0.05	100 - 160	80 - 400	(4)
NW Mediterranean	0.09	70 - 120	160 - 800	(4)
Ionian Sea	0.03	100 - 200	50 - 250	(5)
Levantine Sea	0.005	100 - 200	8 - 40	(6)

(1) McCarthy and Carpenter (1983)
(2) King and Devol ( )
(3) Anderson (19)
(4) Cruzado and Velasquez (1986)
(5) Souvermezzoglu (pers. com.)
(6) Yilmax et al. (1989)

(*) Minimum and maximum values computed with Kz=1.7 and 8.5 m2 d-1 equivalent to 0.2 and 1.0 cm2 d-1

CONCENTRATIONS OF CARBON SPECIES IN THE OCEANS

Adapted from Degens et al. (1984)

CARBON SPECIES	SURFACE		DEEP	
	WARM	COLD	ATLANTIC	PACIFIC
Gaseous CO2	0.01	0.01	0.015	0.02
HCO3-	1.65	1.95	2.10	2.35
CO3=	0.35	0.20	0.15	0.10
T-CO2	2.01	2.16	2.26	2.47
Alkalinity *	2.35	2.35	2.40	2.55
pCO2 **	285	405	400	420

* Carbonate Alkalinity (meq kg-1) = (HCO3-) + 2 (CO3=)

** Atmospheric "equilibrium" value 325 (uatm)

FLUXES AT STATION EROS 2000 MA 1

(umol m-2 d-1)

Computed from:

	PROFILES *	REDFIELD **
Nitrate	115	115
Oxygen	-592	990
T-CO2	3300	760

* Kz = 1.7 m2 d-1

** Computed from NO3 flux with Redfield ratios 106 C:138 O2:16 NO3-

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NITROGEN FLUXES FROM THE ATMOSPHERE INTO THE OCEAN

(umol N/m2 d)

	NO_3^-	NH_4^+	NO_3^-	NH_4^+	NO_3^-	NH_4^+	T-N
Sargassos Sea 1)	10-20	10-15	5-15	0.2-1	15-35	10-16	25-51
Sargassos Sea 2)							
Pacific Ocean 1)	3- 6	2-10	2-6	0.2-1	5-12	2-11	7-23
Pacific Ocean 2)							
NW Mediterran.3)							

1) Duce (1984)
2) McCarthy and Carpenter (1984)
3) Magon et al. (1988) Wet deposition, Alarcon and Cruzado (1989) dry deposition

TABLE I

Eddy Diffusion Fluxes of Nitrate, Oxygen and Carbon in the Mediterranean Sea

Station:	MA/ZA 1	MA/ZA 3
Nitrate gradient: umol N/m4	60	31
Nitrate flux: umol N/m2 d	102 - 510	53 - 265
Carbon flux: (new production) mg C/m2 d	8 - 40	4.2 - 21
g C/m2 yr	2.9 - 14.5	1.5 - 7.5
Oxygen gradient: umol O2/m4	120	358
Oxygen flux: umol O2/m2 d	205 - 1025	600 - 3000
Carbon flux: (respiration) mg C/m2 d	1.8 - 9	5.4 - 27
g C/m2 yr	0.7 - 3.5	2 - 10
Respiration/New production: ("new" organic matter that can be respired with the oxygen flux)	23 %	130 %

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