

The global carbon cycle and the role of the ocean

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Abstract: Only about half of all the CO₂ that has been produced by the burning of fossil fuels now remains in the atmosphere. The CO₂ “missing” from the atmosphere is the subject of an important debate. It was thought that the great majority of the missing CO₂ has invaded the ocean, for this system naturally acts as a giant chemical regulator of the atmosphere. Although it is clear that ocean processes have a major role in the regulation of the carbon dioxide content of the atmosphere through air-sea exchange processes, recent studies of the oceanic carbon cycle and air-sea interaction indicate that oceanic carbon is in a quasi-steady state via the system of biological and physical processes in the ocean interior. It is difficult to determine whether the ocean has the capacity to take up the increasing air-born CO₂ released by human activities over the past five or six decades. To understand this enigma, we need a better understanding of the natural variability of the oceanic carbon cycle.

Key words: carbon dioxide, oceanic carbon cycle, biogeochemical cycling, DOC, greenhouse effect.

INTRODUCTION

One of the principal reasons for the launching of the World Climate Program was the concern over the efforts of the inexorable rise in the partial pressure of CO₂ gas and other trace gases in the atmosphere. Carbon dioxide is one of the most important “greenhouse” gases in the atmosphere. During the past century roughly 200 Gt of carbon as carbon dioxide have been released to the atmosphere from the combustion of fossil fuels and the clearing of forests and other ecosystems for agricultural land. The corresponding increase of carbon dioxide in the air has been from 280 ppm (in the preindustrial age) to 355 ppm. Presumably, most of the anthropogenic carbon dioxide is not retained in the atmosphere.

The greenhouse effect is no new phenomenon. It was at its most efficient in the early days of the earth. The planet would have been frozen without natural concentrations of carbon dioxide and water vapor in the atmosphere, at a time

when the estimated average temperature was minus 20°C, 35°C colder than today.

Life on Earth has experienced countless climate changes in the past 3,000 million years. For example, since the demise of the dinosaurs around 65 million years ago, average surface temperatures have been as high as 10°C warmer, and as low as 5°C cooler, than today. While many factors have contributed to climate trends and fluctuations, variability in the strength of the natural greenhouse effect seems to have been of particular importance.

Thus the overall range of climate extremes has apparently been set by the upper and lower limits in the abundance of carbon dioxide and other greenhouse gases.

Analysis of air bubbles trapped in polar ice has shown the extent and rate of changes in atmospheric carbon dioxide associated with the Earth’s geologically-recent climate oscillations that resulted in ice age cycles. Figure 1 shows an atmospheric CO₂ record over the past 160 kyr

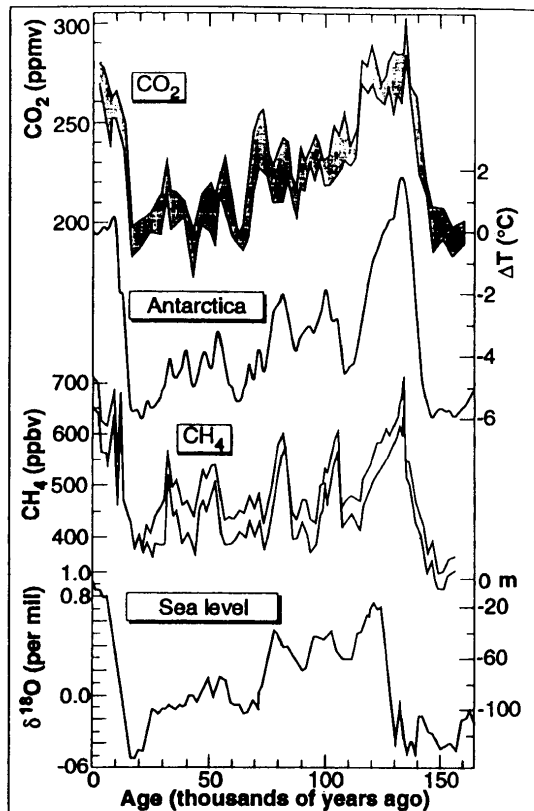


Fig. 1 Changes in atmospheric CO_2 over the past 160,000 years have been closely correlated with the atmospheric temperature (after BARNOLA *et al.* 1987).

obtained from the Vostok ice core in Antarctica (BARNOLA *et al.*, 1987). This is one of the most exciting discoveries of the past decade about the global carbon cycle. The ice cores revealed that the carbon dioxide content of the glacial-age atmosphere was about one-third (180 ppm) lower than typical Holocene values (280 ppm). Anomalies in planetary orbit (affecting the amount and distribution of energy received from the sun) provide the pacemaker for glacial-interglacial temperature changes, but can not account for the magnitude of those changes.

Where did that CO_2 go to and come from? The answer may lie in the ocean, with its great supply of exchangeable carbon. It is widely accepted that oceans have a moderating effect on climate. The effect is due mainly to the special characteristics of water: its large heat-carrying capacity allows ocean currents to redistribute heat at the surface of the earth. In addition, the chemical nature of water plays an important role

in the course of the Earth's climate. The oceans are a huge reservoir for carbon dioxide, containing 50 times more than the atmosphere and 20 times more than the terrestrial biosphere. The oceans are the major natural factor determining the carbon dioxide content of the atmosphere, with a critical effect on the global heat balance.

However, little is known about the oceanic responses to rapid changes in climate, the way in which the oceanic carbon balance is related to physical and biological processes, and the role of the ocean in taking up excess CO_2 (and factors that affect it such as air / sea CO_2 exchange and organic carbon concentration).

MAJOR CARBON RESERVOIRS AND FLUXES

Carbon is the key element of life. The carbon cycle is, therefore, of fundamental importance when trying to understand the biosphere and its basic mechanisms and, conversely, when determining the importance of the role of the biosphere in controlling the carbon cycle. Carbon is present on the earth in many different chemical compounds: as carbon dioxide, carbon monoxide and organic compounds (hydrocarbons, organic acids, chlorofluorocarbons, etc.) in the air, as carbonate ions, carbon dioxide, carbon monoxide, and organic compounds in sea water, as carbonates and organic compounds in the soil, and as organic compounds in the biota.

One example of the global carbon cycle is shown in Figure 2. The sea water reservoir contains about $40,000 \times 10^{15}$ grams or 40,000 gigatons of carbon (largely as the bicarbonate ion HCO_3^-), and its annual exchange rate with the atmosphere is on the order of 80 Gt per year. The atmospheric reservoir itself contains only about 700 Gt as carbon dioxide, and exchanges at a rate of about 70 Gt C per year with the terrestrial biosphere. The latter contains about 2000 Gt C, of which some 70 percent, say 1400 Gt C, is in the soil (humus), with the remaining 600 Gt C above ground as plants (DEBAAR *et al.* 1988). The annual carbon budget is shown in Table 1 (SUGIMURA 1990). The annual budget does not balance, and as yet unknown sinks, on land or in the oceans, must play a role. The imbalance may be diminished in several ways: by reductions in the estimate of the rate of deforestation or increases in the regrowth; by larger estimates of oceanic uptake; by significantly shorter term natural variations in the atmosphere that override

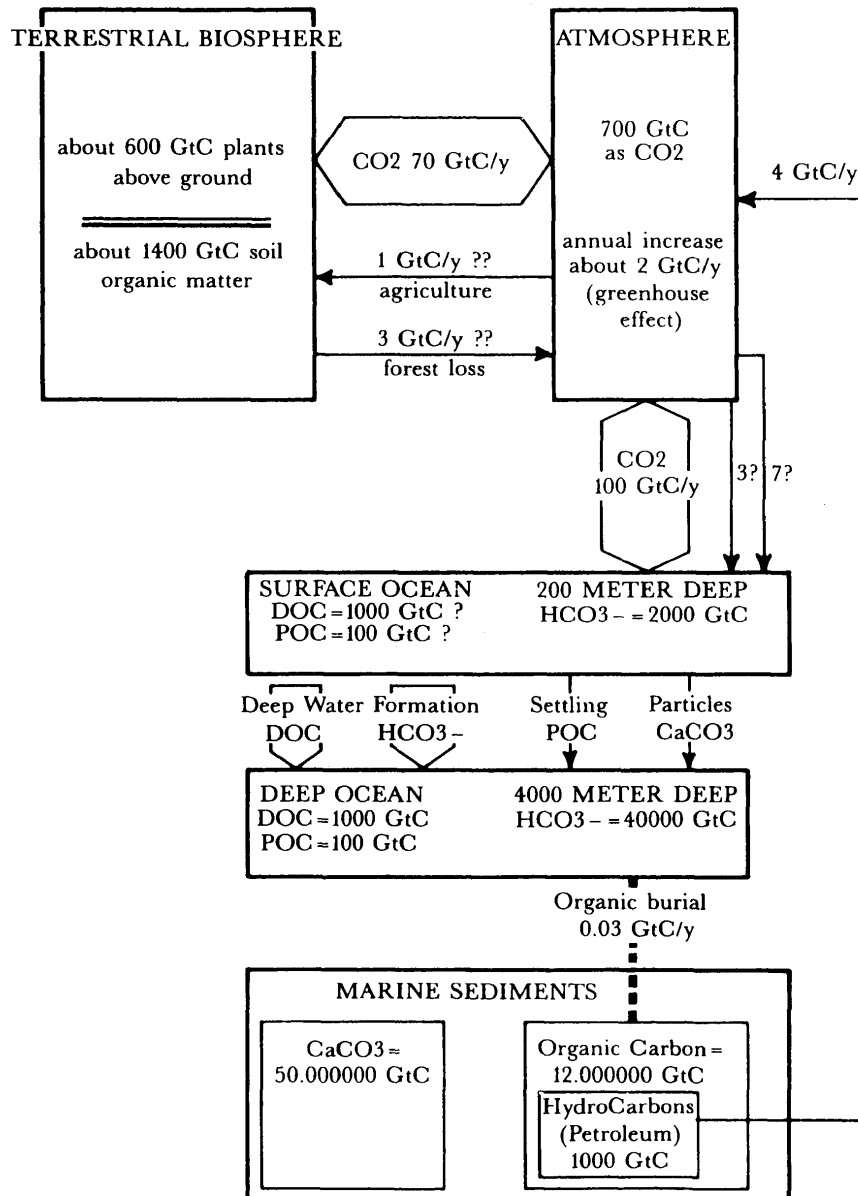


Fig. 2 The global carbon budget, after H. DEBAAR *et al.* (1989). (1 gigaton – 10¹⁵ grams C). Large open arrows represent natural fluxes, thin arrows represent human activities.

the imbalances; by increased estimates of uptake by oceanic productivity, through the “biological pump”; or by increased estimates of the downward flux of organic carbon. The question of which combination of these possibilities is more likely is both important and fascinating.

The present average concentration of carbon dioxide in the atmosphere is about 353 ppm (parts per million). The residence time (residence time = budget / annual flux) of atmospheric carbon dioxide is about 4 years. As shown in Figure 2, there is more carbon in the atmosphere than there

is in all the world's forests – a very surprising fact. There are, of course, other “pools” of terrestrial carbon. The amount of carbon released globally from vegetation and soils as a result of deforestation remains uncertain, though much progress has been made in recent years. The difficulties in calculating this release is due to uncertainties in two factors: one is the rate of land – use change and another is the response of

the biota to disturbances.

Calculations of the net carbon loss from the global biotic inventory attempt to take into account the uncertainties concerning disturbance rates and biotic response. Table 1 shows that at present the contribution of fossil fuel combustion to atmospheric CO₂ is about 5.5 Gt C, of which about 50% is not retained in the atmosphere. As shown in Table 1, the ocean plays an important

Table 1 Major global carbon reservoirs and fluxes.

<u>Reservoirs</u>		<u>10¹⁵ g C</u>
Atmosphere: CO ₂ :	Before 1850 ca. 280 ppm	594
	1980	717
	Other gases: CH ₄ :	4
	CO:	0.1 ppm
(Troposphere: 80 %, stratosphere: 20 % of atmospheric mass)		
Oceans:	Inorganic C (ECO ₂)	37,400
	Dissolved organic matter	ca. 1,000
	Biomass	3
Land biosphere:	Living	560
	Soil, humus (pre-historic: 200-500 · 10 ¹⁵ g C more)	1,500
Groundwater:		450
Sediments:	Inorganic C	ca. 60,000,000
	Organic C	ca. 12,000,000
Fossil fuels:		ca. 5,500
<u>Fluxes (gross)</u>		<u>10¹⁵ g C/yr</u>
Atmosphere-ocean, CO ₂ exchange		78
Atmosphere-land biota, photosynthesis/respiration (NPP)		65
Marine photosynthesis		45
Sedimentation in oceans		0.2
Volcanism		ca. 0.9
Fossil fuel combustion, 1980		5.3
<u>Residence times: τ = mass/flux</u>		
Atmosphere (pre-industrial): total exchange		4 yr
	exchange with ocean only	8 yr
	exchange with biosphere only	9 yr
Living land/biosphere: photosynthesis/respiration		11 yr
Marine biosphere: photosynthesis/respiration		0.07 yr
Oceans: exchange with atmosphere, total flux		490 yr
	sedimentation only	180,000 yr
Atmosphere + biosphere + oceans: sedimentation		210,000 yr

role as a huge reservoir related to the regulation of atmospheric carbon dioxide. However, there is a great deal of uncertainty regarding the rate of oceanic uptake or release of carbon dioxide.

INCREASE OF CO₂ IN THE ATMOSPHERE

Long term variations in the concentration of atmospheric CO₂ at Mauna Loa Observatory, Hawaii are shown in Figure 3. The smooth curve represents a fit of the data to an annual cycle; this annual cycle shows increases linearly increases and decreases with time. The dots indicate monthly average concentrations (BACASTOW & KEELING 1985). The atmospheric carbon dioxide concentration gives a remarkable portrait of the "natural" or, if you will, the ecological world, overlain by the "industrial" or economic world. An oscillation is seen, which is principally a reflection of the Northern Hemisphere's seasonal pattern of biological activity. Carbon dioxide decreases in the spring time as plants begin to "wake up" and take in carbon dioxide through photosynthesis, which exceeds the return of carbon dioxide as a result of decomposition of organic matter in the soil. This continues through the summer, but with fall there is a reduction

in photosynthetic activity, and carbon dioxide cycles upward. This continues further as plants drop their leaves, and the plant material begins to decompose and releases additional carbon dioxide. Consequently, the atmospheric carbon dioxide peaks, only to decline again the following spring. The carbon from burning fossil fuels, harvesting forests, and converting land to agriculture is reflected in the trend of increasing carbon dioxide concentrations that is superimposed on the seasonal pattern. The upward trend in atmospheric CO₂ is significant, especially if it continues.

SUGIMURA (1990) discussed the relationship among cumulative emission, CO₂ in air, and the airborne fraction of the emissions. He concluded that even if CO₂ release decreases, the CO₂ in the air will continue to increase; the CO₂ concentration in the air is linearly related with the cumulative emission of CO₂, assuming a nearly constant airborne fraction of 60% as shown in Figure 4.

The airborne fraction of carbon dioxide is determined by the exchange of CO₂ between the atmosphere and biosphere and by the exchange of CO₂ between atmosphere and ocean. A more important factor is to know the net flux of carbon dioxide among the different reservoirs.

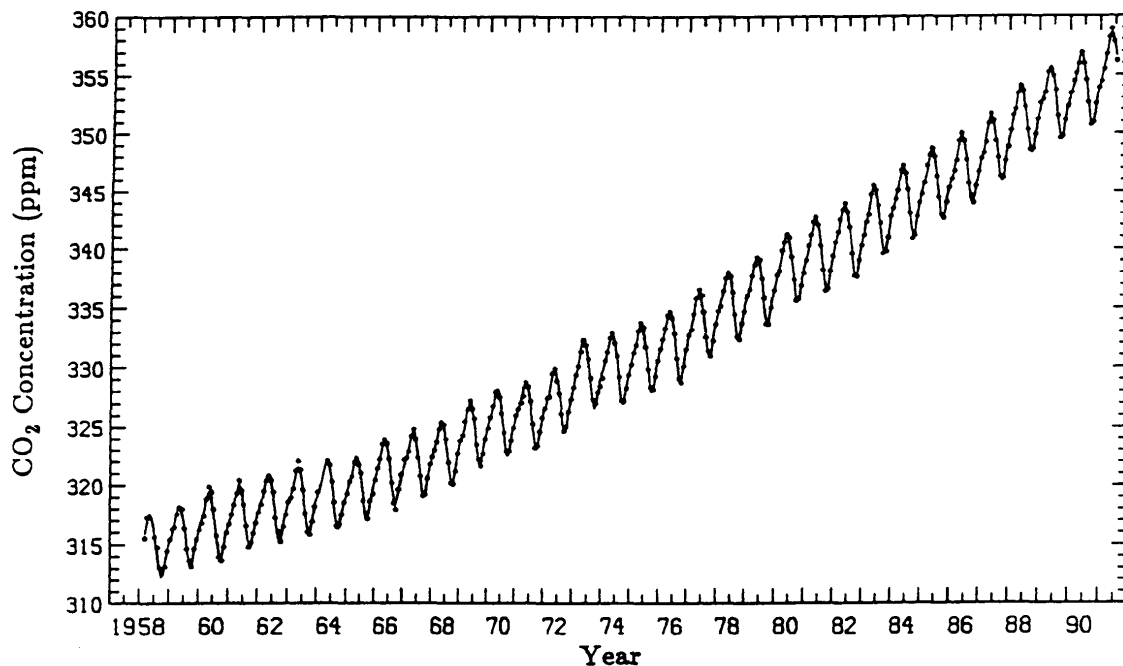


Fig. 3 The concentration of atmospheric CO₂ (ppm) in dry air volume at Mauna Loa Observatory Hawaii versus time. The dots denote monthly mean concentrations (after KEELING 1993).

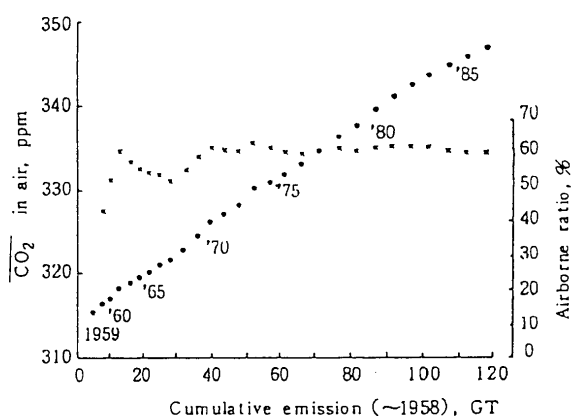


Fig. 4 The relationship between mean CO_2 concentration and cumulative emission of CO_2 , and airborne ratio in the atmosphere (after SUGIMURA 1990).

AIR / SEA INTERACTION

The direction of CO_2 transfer between air and water may be determined by the difference between the partial pressure of CO_2 in the two reservoirs. Since partial pressure ($p\text{CO}_2$) in the atmosphere varies by no more than five percent over the globe at any given time (excluding the areas affected by local industrial emissions), and since the $p\text{CO}_2$ in the surface waters of the open seas varies from about one half to twice the atmospheric value, any variation in the difference is due mainly to changes in the $p\text{CO}_2$ of sea water. It in turn is governed primarily by water temperature, pH, the total CO_2 concentration, and primary production in sea water.

Water temperature changes geographically – 1.9 degrees in the polar seas to 30 degrees in the equatorial oceans – and also seasonally by as much as 15 degrees at a given location. CO_2 concentration may be reduced by the photosynthetic utilization of carbon, and increased by the release of CO_2 due to respiration and oxidation of organic matter. To a lesser extent, it is also affected by the CO_2 gas exchange with the atmosphere and by the formation and dissolution of calcareous shells. Therefore, the $p\text{CO}_2$ in sea water depends on a complex web of interactions involving physical, chemical, and biological processes occurring in the surface layer of the oceans.

Figure 5 shows the distribution of the air / sea CO_2 partial pressure difference ($\Delta p\text{CO}_2 = p\text{CO}_2$ of sea water - $p\text{CO}_2$ of air) in the surface water over the Pacific Ocean (MIYAKE & SUGIMURA 1974;

INOUE & SUGIMURA 1987, 1988). The positive values indicate regions where the surface ocean is a source of atmospheric CO_2 , while the negative values show those parts of the ocean where it is a “sink” for atmospheric CO_2 . Results indicate that the atmosphere over the Pacific Ocean is out of equilibrium with the surface ocean CO_2 , because the $p\text{CO}_2$ values are not zero. This suggests that the ocean water exhibits a large $p\text{CO}_2$ difference from the air because the gas transfer rate of CO_2 to and from the atmosphere is much slower than the rates of change due to temperature change or due to biological processes, both of which have a large effect on the $p\text{CO}_2$ in sea water. Figure 5 shows that the equatorial Pacific water is supersaturated with CO_2 and that the middle latitude Pacific sea water is undersaturated with CO_2 .

The $p\text{CO}_2$ variation of the equatorial Pacific is affected by the El Nino / Southern Oscillation event (FEELY et al 1987; INOUE & SUGIMURA 1988). During the El Nino event of 1982, the $p\text{CO}_2$ of the equatorial waters in the central Pacific was reduced to nearly equal to that of the atmospheric value, suggesting that the high $p\text{CO}_2$ equatorial water of normal years was overridden by and covered with a layer of the warmer low $p\text{CO}_2$ water that rushed eastward from the western equatorial Pacific (FEELY et al 1987). The $p\text{CO}_2$ of the equatorial water, during the ENSO event of 1987, was also reduced, but the reduction of the $p\text{CO}_2$ concentration is smaller than the 1982 / 1983 events (INOUE & SUGIMURA 1988). The most surprising matter is that the $p\text{CO}_2$ in the surface water of the middle and higher latitudes of the Pacific was nearly constant for more than 20 years (INOUE & SUGIMURA 1988). During the same period, CO_2 in the air increased at an average rate of 1.5 ppm / y. Similar results were reported in the North Atlantic (TAKAHASHI *et al.* 1983).

The total mass balance of carbon dioxide between air and ocean is not clearly understood, but according to recent studies using ocean carbon models and observations, it seems that the physical mixing processes influencing absorption of the excess carbon dioxide may not play an important role.

THE OCEAN CARBON CYCLE

The rate of carbon dioxide uptake by the oceans is controlled by sea water temperature, surface chemistry, and biology, and by the various patterns of mixing and circulation that determine

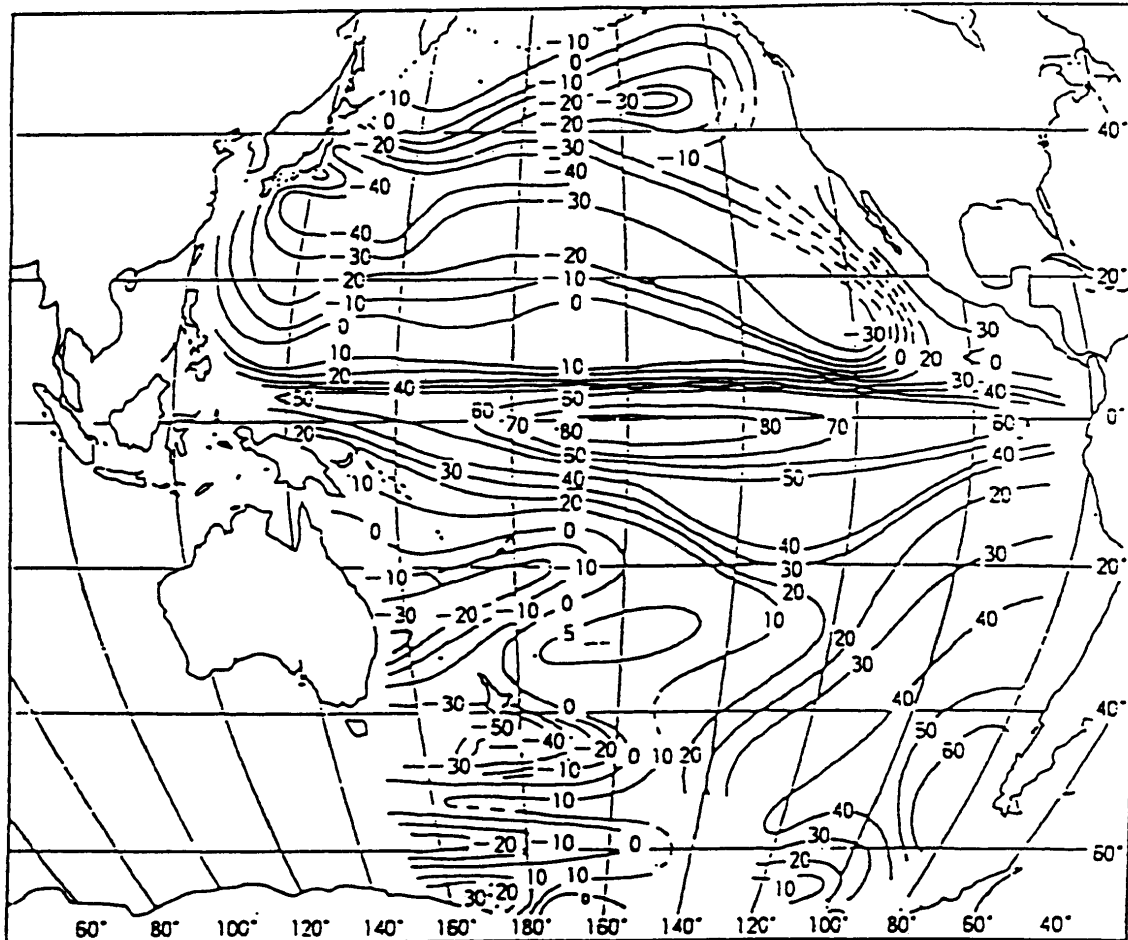


Fig. 5 The surface distribution of the difference between $p\text{CO}_2$ of the atmosphere and surface ocean (i.e., the disequilibrium) in the Pacific Ocean (ppm) (after MIYAKE & SUGIMURA 1974).

the amount of carbon transported from surface waters to the deep ocean as organic matter (dissolved and particulate forms) and carbonate. Carbon distribution in the ocean is influenced by the following phenomena: 1) the consumption of carbon dioxide in primary production in biologically active surface waters; 2) the enrichment of deep water in carbon dioxide as a result of the decomposition and dissolution of carbonate and organic matter (dissolved and particulate forms) produced in the surface water; 3) the sinking of water in polar regions, particularly in the North Atlantic, taking CO_2 with it, followed by a general bottom water flow toward the equator; 4) the upwelling of some of this water in equatorial regions with corresponding outgassing of carbon dioxide to the atmosphere and a general poleward flow of surface waters; 5) general turbulent mixing processes accompanying the meridional circula-

tion, whereby the carbon-rich water at intermediate depths is continuously exchanged with surface water of lower carbon content.

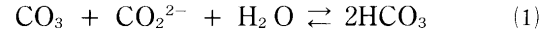
The first two features taken together are related to the biogeochemical processes in the ocean, the so called "biological pump" and "biodynamic converter" (SUZUKI 1993b). The "pump" in biological pump involves the incorporation of CO_2 , in tissue or as carbonate in shells, into living organisms that had been dissolved in surface waters; this is followed by the transportation of dead marine organic particles, mostly from the euphotic zone. The biodynamic converter relates to the flux and rate constant of production and decomposition of DOC within the upper ocean layer. As discussed in the above section, it is likely that the "physical pump," the injection of carbon dioxide from the air into the sea, is not effective. The "biological pump" has

been given much attention as a possible missing path of atmospheric carbon dioxide.

In general the photosynthetic C - fixation (for primary production) by algae is responsible for the massive conversion of dissolved inorganic carbon into plant organic matter. This leads to lower values of DIC in the surface ocean water. If this creates a negative disequilibrium between sea water and air, further physicochemical dissolution of atmospheric CO_2 into the sea becomes possible; if the disequilibrium is positive, but lessened, escape of CO_2 to the atmosphere slows down. To evaluate how the biological pump

is related to the injection of atmospheric carbon dioxide into the ocean is to understand the oceanic carbon cycle quantitatively.

Dissolved inorganic carbon (DIC) is divided between CO_2 and the much larger pool of bicarbonate and carbonate: the balance is determined by the reversible reaction



Simulation of this chemical system requires knowledge of the apparent dissociation constants (K_1 , K_2) of carbonic acid and boronic acid (KB) in sea water, the gas solubility, and equations linking the variables pH, pCO_2 , total CO_2 and

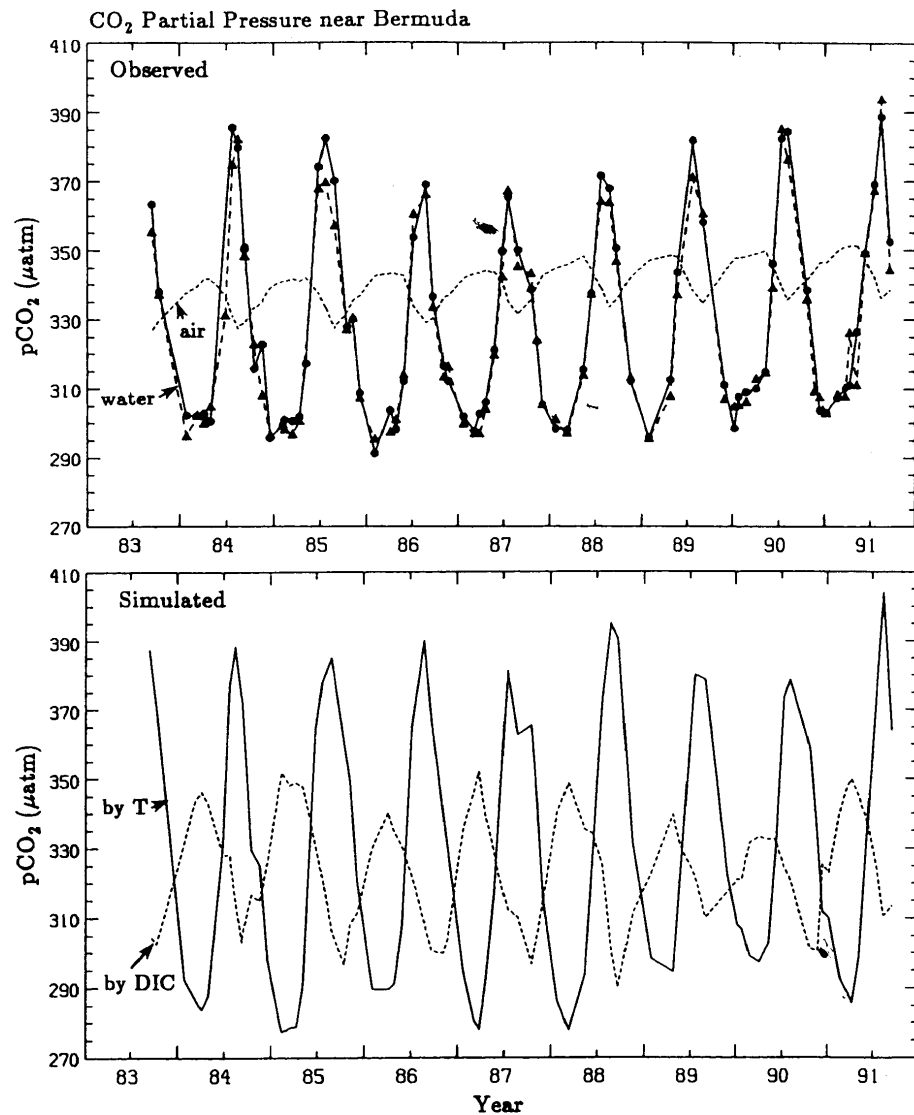


Fig. 6 Upper: Time series of pCO_2 in the surface water near Bermuda. Lower: pCO_2 which would occur if the variation were produced solely by the temperature (circles and solid line segments) or by variations in DIC (plus marks and dashed lines) (after KEELING 1993).

alkalinity. The changes in $p\text{CO}_2$ and DIC concentration are controlled by physical (advection, mixing, and air-sea exchange), biogeochemical (production and regeneration), and inorganic chemical (mineralization and dissolution of calcium carbonate) processes. Of these parameters it is most important to evaluate the biological contribution in the oceanic carbon cycle. The time series data of DIC and organic carbon (dissolved and particulate forms) are very useful for understanding the biogeochemical cycle in the ocean.

KEELING (1993) has reported time variations of DIC and $p\text{CO}_2$ near Bermuda since mid-1983. Their data show a pronounced seasonal oscillation in DIC of approximately $30 \mu\text{mol kg}^{-1}$, but no regular oscillation in alkalinity (Fig. 6). This oscillation implies a seasonally varying uptake of inorganic carbon by soft tissue that explains approximately half of the seasonal cycle in DIC. The remaining variation is mainly from the combination of air-sea exchange and vertical transport. High DIC concentrations occur in early spring when surface water, at its yearly minimum temperature, overturns with deeper water. Low DIC concentrations occur in late summer, shortly after the temperature has reached its seasonal maximum. This time lag in minimum DIC suggests that primary production of organic carbon continues to reduce DIC after the water has started to cool. On the basis of the physical oceanographic data, the coefficient of vertical diffusion, the mixed layer depth, the difference of $p\text{CO}_2$ between air and sea, and the gas exchange rate, they calculated the Bermuda seasonal cycle in DIC fluxes (Fig. 7). As shown in Figure 7, the exchange of carbon between DIC and organic carbon (dissolved and particulate forms) is the dominant flux. When this flux is negative, net primary production exceeds respiration. This suggests that of the processes controlling the oceanic carbon cycle, the study of the carbon cycle in relation to the biogeochemical processes, in particular, the conversion of inorganic carbon to organic carbon by photosynthesis and organic carbon to inorganic carbon by decomposition, appears most important. However, the flux and turnover time of carbon between both reservoirs are still not well known. The study of organic carbon is a key for understanding the biogeochemical cycle of carbon in the ocean.

The major part of organic carbon in the ocean is dissolved organic carbon (DOC), accounting for more than 90% of the total. Until recent studies by Suzuki and co-workers, it was believed that DOC played an unimportant role in the oceanic

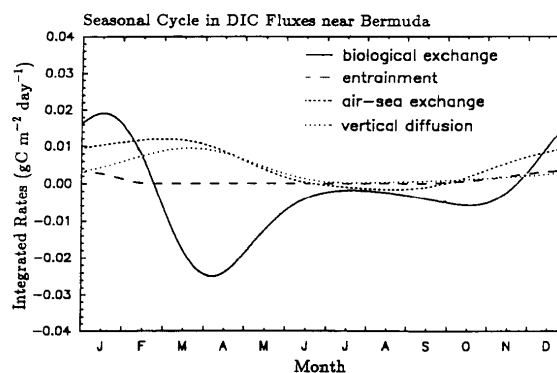


Fig. 7 Seasonal variations in the rates of change of DIC, in ($\text{gCm}^2\text{day}^{-1}$) integrated over the surface mixed layer (after KEELING 1993).

carbon cycle. This was because most DOC in sea water was thought to consist of “refractory” compounds with an “old” apparent age of more than 3000 years (WILLIAMS & DRUFFEL 1987).

Recent new measurements of DOC in sea water using a high temperature catalytic oxidation (HTCO) method unveiled new aspects of DOC. These studies suggested that DOC plays an important role in biogeochemical cycles of carbon, as evidenced by its relationships to oxygen consumption and remineralization to DIC, compared with particulate organic carbon (sinking particulates) (SUGIMURA & SUZUKI 1988; SUZUKI & TANOUE 1991, SUZUKI *et al.* 1992, 1993, KIRCHMAN *et al.* 1991). The typical vertical profile of DOC in the western North Pacific is illustrated in figure 8 with that obtained by the traditional method. New measurements of DOC indicate that this compound is three to four more times abundant in surface than deep water, decreasing rapidly with depth; it is twice as high in deep water as previously thought, and is highly correlated with oxygen consumption and the DIC profile (Fig. 9).

According to the traditional view, organic matter particles sink or by other means are removed from the euphotic zone; they were considered to be a source of DIC through remineralization, associated with oxygen consumption after sinking into deep water. However, several decades of sediment trap experiments and large volume sea water filtrations did not provide any satisfactory explanations for the mechanisms by which the oxygen is consumed and inorganic nutrients are remineralized in the deep water column. Most of organic matter produced is recycled within the upper 200 m of the surface. This imbalance can be explained by the

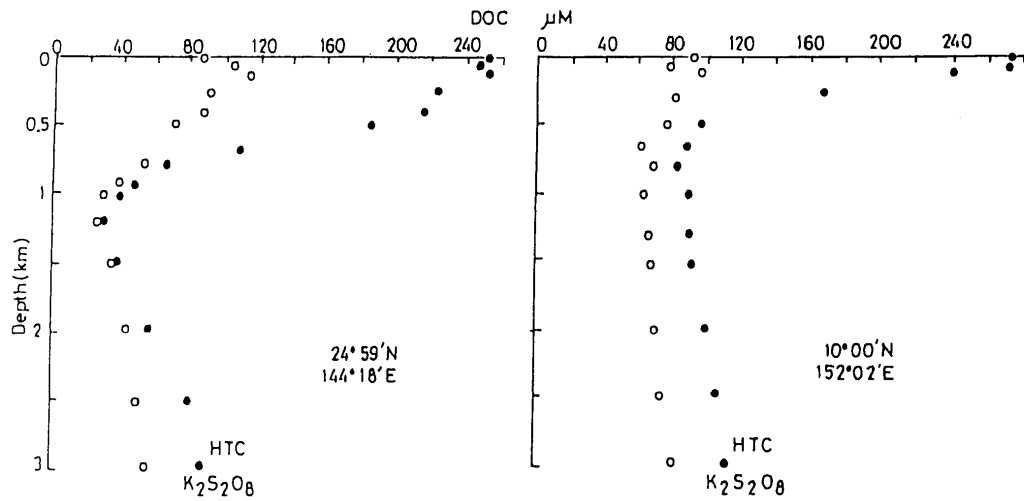


Fig. 8 The vertical distribution of DOC in the western North Pacific ocean; open circles: wet chemical oxidation, closed circles: high temperature catalytic oxidation (HTCO) (after SUZUKI & TANOUE 1991).

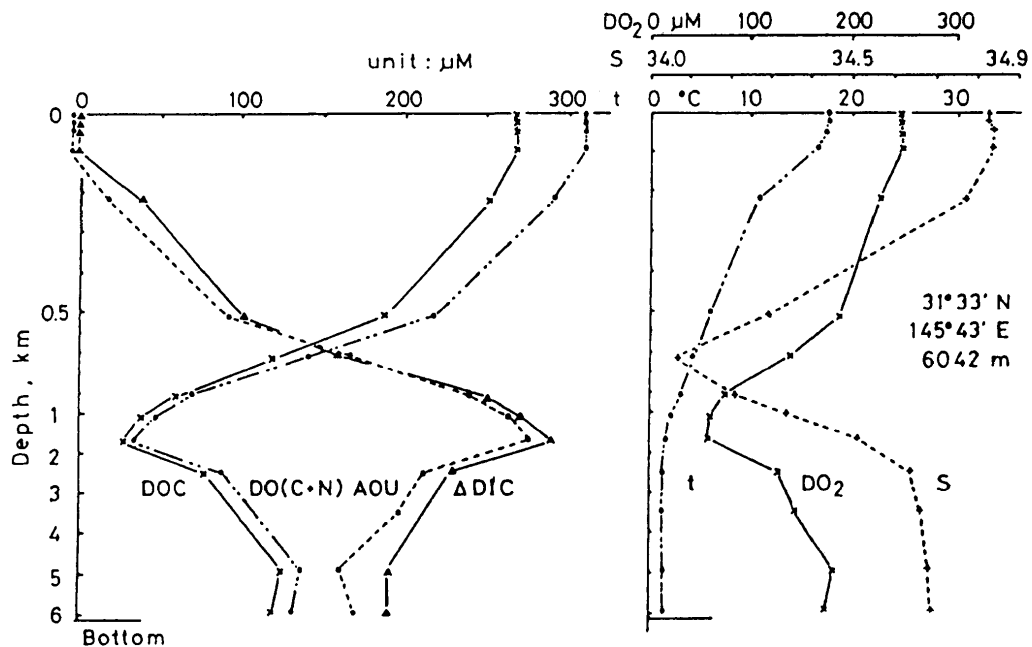


Fig. 9 Vertical distribution of DOC, AOU, dissolved inorganic carbon in the western North Pacific (AOU: apparent oxygen utilization) (after SUGIMURA & SUZUKI 1988).

decomposition of DOC in deep water. DOC is thus a key ingredient to understanding the ocean carbon cycle.

Most of the previous models of the oceanic carbon cycle are mainly concerned with the

inorganic carbon cycle, but the oceanic carbon cycle may be controlled by the exchange and transport of DIC and DOC (Fig. 10), thus DOC would be related in an essential way to the air-sea exchange of carbon dioxide. Figure 11 shows

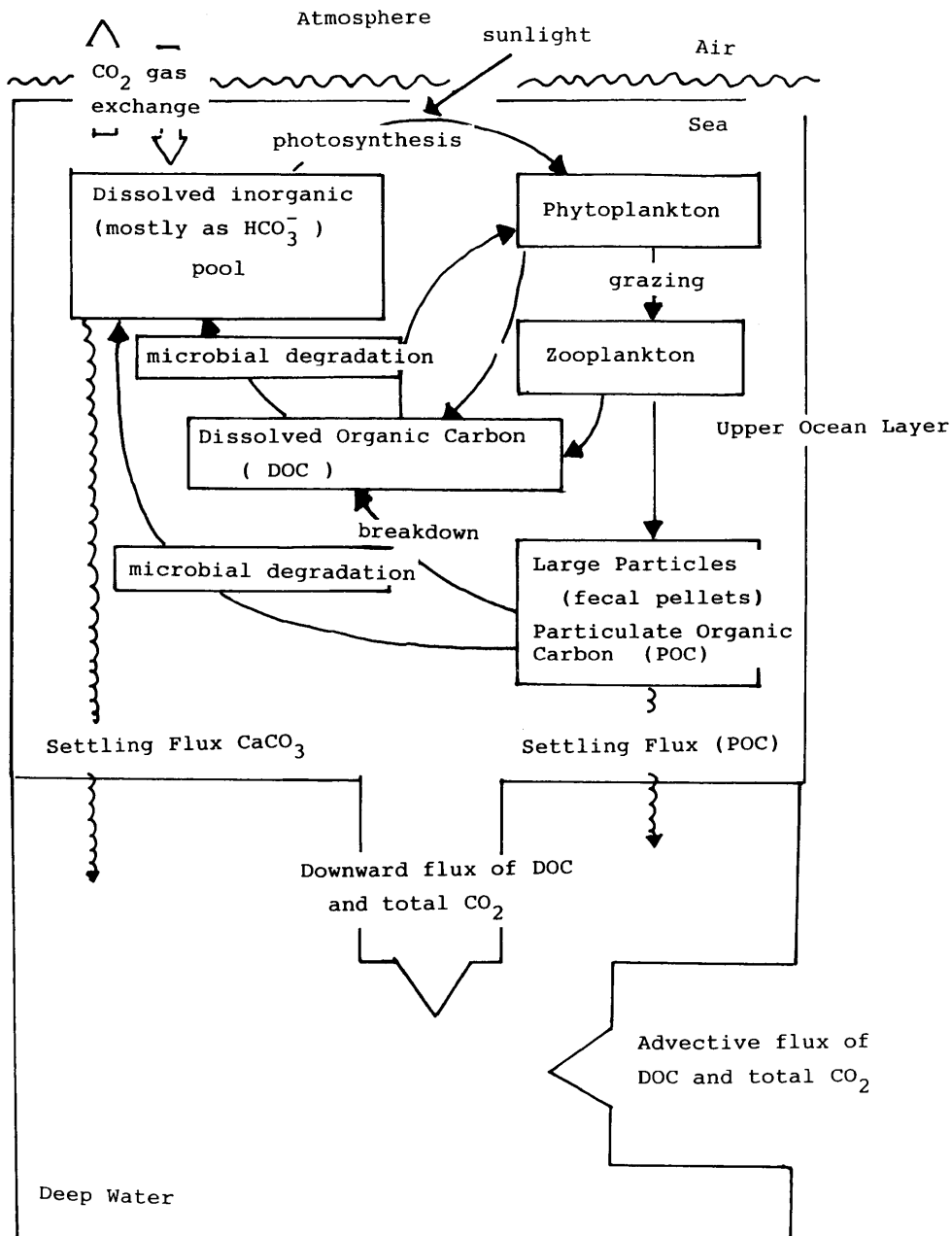


Fig. 10 Simplified scheme of oceanic carbon cycle (compartments and fluxes) in the upper ocean (after H. DEBAAR *et al.* 1989).

the net flux of CO₂ in the world ocean; the Pacific Ocean is generally a source and the Atlantic Ocean a sink, suggesting that there is a balance between source and sink in sea - sea interactions. However, the CO₂ flux has a seasonality based on the change of water temperature and plankton blooming. A 4% change in pCO₂ is caused by a change of 1 degree C in water temperature. This means that, for example, in the case of surface

sea water with 300 ppm of pCO₂, a change of pCO₂ of 60 ppm will occur if there is a change of 5 degrees C water temperature (for example in the middle of the Pacific Ocean), and turn a sink into a source for an atmospheric pCO₂ of 350 ppm. In addition, if the change of CO₂ caused by the plankton blooming could be estimated quantitatively, possibly a sink of "excess" CO₂ into the sea could be found.

A two - box model of the oceanic carbon cycle is shown in Figure 12. The annual primary production (DIC to DOC) is 50 Gt yr^{-1} , of which 80% of the DOC is regenerated within the upper 100 m depth. The downward flux of DOC with vertical mixing of the water mass is 11.6 Gt yr^{-1} , the upward flux is 6.8 Gt yr^{-1} , and the net downward flux of DOC is 5 Gt yr^{-1} . This is greater than previously thought. Recently SUZUKI (1993b), BACASTOW & MAIER-REIMER (1991) and KUMAR *et al.* (1991) reported carbon cycle models of the ocean, and came to the conclusion that DOC plays an important role in the oceanic carbon cycle. In particular, the DOC cycle in the ocean is more dynamic, with a short turnover time of 50 yr to 250 yr, than previously believed. The oceanic carbon cycle is possibly controlled by a “quasi” steady state of production and decomposition and by water circulation.

Is there any possibility that “excess” carbon dioxide is injected from the atmosphere into the ocean? Recent studies of the oceanic carbon cycle and air - sea interactions indicate that it is difficult to determine whether the ocean has the capacity to take up the increasing atmospheric carbon dioxide stemming from human activities over the past five decades.

CONCLUSION

What has changed in the gradient between the atmosphere and the ocean? The concentration of carbon dioxide in the atmosphere has been increasing; certainly the concentration is greater than, for example, 200 years ago. The oceanic flux of carbon dioxide in the ocean interior is probably balanced by the biological and physical processes occurring there. Although the increase of atmospheric carbon dioxide is proportional to the increase of $p\text{CO}_2$ in the surface sea water, with an average of 1.5 ppm yr^{-1} , the maximum amount of injection of CO_2 into the ocean from the atmosphere is about 0.5 Gt yr^{-1} , which is only 10% of the annual emission of CO_2 into the atmosphere (4.0 to 5.0 Gt yr^{-1}). Do the “biological pump” or “biodynamic converter” give us a satisfactory explanation of the ocean carbon cycle’s relation to the air - sea interaction of CO_2 ? It is likely that a key to solving this question is in making clear the mechanisms behind the “quasi” steady state of the upper ocean in the carbon cycle, using an explanation based on the biodynamic converter. In other words, the net downward flux of DOC from the upper ocean to deep water must be evaluated. A value of about 5 Gt yr^{-1} of net downward flux of DOC, calculated by a two box model (SUZUKI 1993), is much greater than previously thought. We think that a quantitative explanation of the missing carbon sink

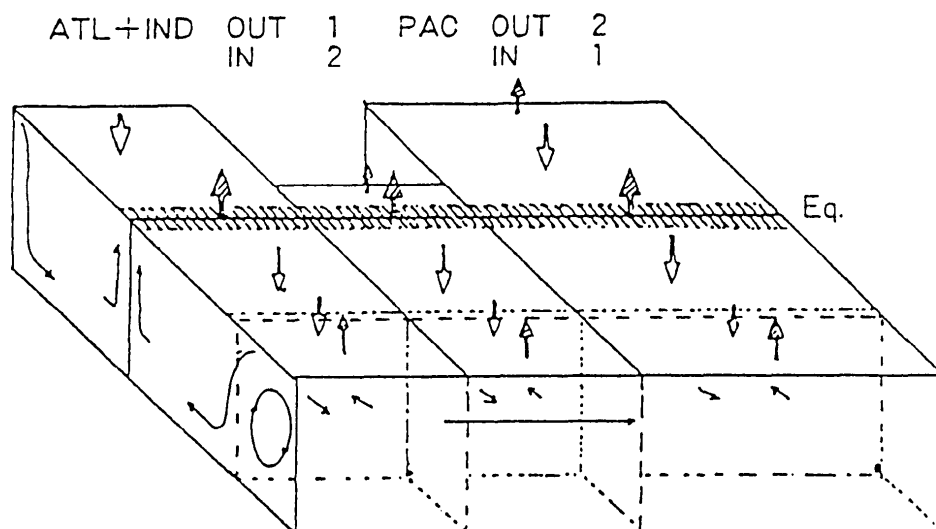


Fig. 11 CO_2 fluxes in the world ocean (after SUGIMURA 1990).

