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IMPACT OF TERRESTRIAL AND OCEANIC BIOTA ON THE MODERN CARBON AND OXYGEN CYCLES

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The biotic response to the anthropogenic perturbation of the environment strongly depends on the state of biota: virgin oceanic biota negatively feedbacks to the increase of the CO₂ concentration in the environment, now absorbing carbon from the atmosphere at the rate of up to 5 GtC yr⁻¹. This figure is obtained from two independent data sources: ¹⁴C/¹²C ratio measurements in the oceanic dissolved organic carbon and O₂/N₂ ratio measurements in the atmosphere. On the contrary, terrestrial biota strongly perturbed by anthropogenic activities, emits carbon to the atmosphere adding to the anthropogenic perturbation of the environment. The rate of carbon emission by terrestrial biota is found from the balance equation and is about 4 GtC yr⁻¹ which is consistent with the maximum direct modern estimates of carbon emissions due to land use. Restoration of virgin boreal and tropical forests could stop the growth of atmospheric carbon dioxide mass even if the present high rate of fossil fuel combustion remains the same.

Key words: *biota, anthropogenic perturbation, ocean, atmosphere, carbon and oxygen cycles.*

Introduction

Biological processes impose strong impact on the physical environment and this relationship may feature both negative (stabilizing) and positive (destabilizing) feedback. Possible response of biota to environmental perturbations has been so far considered either at the level of a single organism deprived of its natural ecological background or as a reaction of anthropogenically perturbed biological communities (e.g. exploited forests and pastures) [1,2]. Natural ecosystem, however, as a complex strongly correlated community of species, can have properties that cannot be predicted from studies of its isolated components and that can be lost when the ecosystem is perturbed. Being less destroyed by anthropogenic activities than terrestrial ecosystems, oceanic ecosystems represent a good object for studying feedback loops between the natural biota and the physical environment. If it proves that virgin unperturbed biota represents a considerable stabilizing force in respect to the growing atmospheric CO₂ concentration, then the task of conservation of virgin ecosystems may become of vital importance in stopping the global change processes of today.

Meanwhile possible role of oceanic biota has been ignored in most global change studies [3-6]. It has been assumed that productivity of oceanic biota cannot increase with increasing atmospheric CO₂ concentration, as it is limited by oceanic concentrations of nitrogen and phosphorus which now remain nearly constant [7,8]. Yet there are two ways for excessive atmospheric CO₂ to be absorbed by oceanic biota at constant concentrations of nitrogen and phosphorus. Firstly, phytoplankton may increase production of extracellular hydrocarbon-type excretions containing neither nitrogen nor phosphorus [9,10].

Total cellular production of the ocean remains then unchanged, and the newly synthesised dissolved organic matter will be characterised by a higher carbon/nitrogen and carbon/phosphorus ratios as compared to the preindustrial era. Secondly, it is possible to increase the share of the dissolved organic matter production in the remaining constant net oceanic primary production reducing, accordingly, the share of cellular production. Then the carbon/nitrogen and carbon/phosphorus ratios in the increased production of the dissolved organic matter will remain at the preindustrial level.

We use two independent data sources to evaluate the oceanic biotic sink of atmospheric carbon. The data used include the depth profile of $^{14}\text{C}/^{12}\text{C}$ ratio in the dissolved organic carbon (DOC) in the ocean [11] and recent data on atmospheric O_2 concentration change [6]. We find an oceanic biotic sink approximately two times larger than the known oceanic inorganic sink [6,12–14]. Oceanic biota is far less perturbed by anthropogenic activities than terrestrial biota which releases considerable CO_2 amounts adding to the anthropogenic perturbation of the environment [9,10,15]. Our results suggest that unperturbed ecosystems perform a stabilising function buffering detrimental environmental changes.

Equilibrium DOC state in the ocean

Concentrations of radiocarbon vary at an approximately constant gradient from the surface (s) to depth L , L is less than 1 km, and then at deeper levels (d) remain approximately constant down to the very seabed at depth H , see Fig.1. We denote all the surface preindustrial values by the additional “0” low index. Denoting the concentration of isotope ${}^{\vee}\text{C}$ by the same symbol in italics (and indicating additionally its localization in brackets in either organic (+) or inorganic (–) reservoir, e.g. ${}^{\vee}\text{C}(s_0^+)$ or ${}^{\vee}\text{C}(d^-)$), we arrive at the following equation of mass budget for carbon (either “+” or “–”) in the oceanic column of unit area and depth H :

$$H \frac{d{}^{\vee}\text{C}(d)}{dt} = {}^{\vee}F - \frac{{}^{\vee}\text{C}(d)H}{{}^{\vee}T} \quad (1),$$

where ${}^{\vee}F$ is the influx to that reservoir column, ${}^{\vee}T$ is the residence time for atoms of carbon isotope ${}^{\vee}\text{C}$ in the reservoir. The mass of carbon in the oceanic column is assumed equal to ${}^{\vee}\text{C}(d)H$. Deviation of preindustrial concentration in the surface layer from its value at depths biases the total mass of carbon by only several per cent, see Fig. 1.

The influx channel to reservoirs of both the organic and inorganic carbon is the same for both ^{12}C and ^{14}C . However, ^{14}C has an addition-

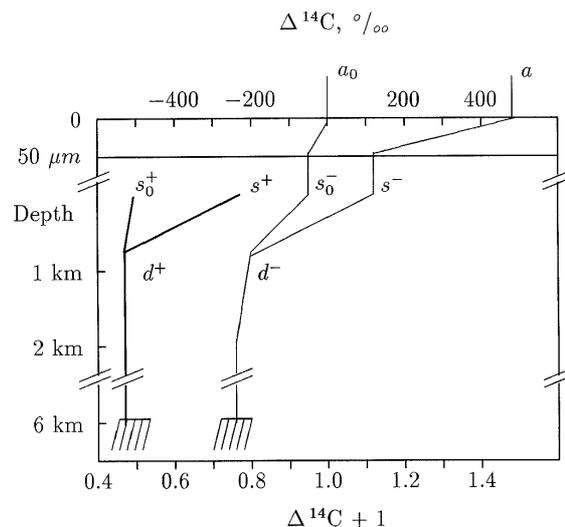


Fig.1. Profiles of the observed oceanic DOC (thick line) and DIC (thin line) $\Delta^{14}\text{C}$ [11]; s and d denote the surface and deep waters respectively. Upper indices + and – relate to DOC and DIC respectively. Low index “0” denotes the calculated equilibrium preindustrial state for DOC and observed prebomb state for DIC.

al escape channel from each reservoir — the decay channel. The residence time for ^{14}C in the reservoir is determined by the well-known relationship for any two parallel channels:

$${}^{14}T^{-1} = {}^{12}T^{-1} + T_c^{-1} \text{ or } \tau^{-1} = T^{-1} + 1, \quad (2)$$

$$\tau \equiv {}^{14}T/T_c, \quad T \equiv {}^{12}T/T_c,$$

where $T_c = 8267$ ·yrs is the average lifetime of ^{14}C [16,17].

The influx ${}^{\vee}F$ in both organic and inorganic reservoirs is proportional to surface water concentration of inorganic carbon ${}^{\vee}\text{C}(s^-)$:

$${}^{\vee}F_0 = {}^{\vee}\text{C}(s_0^-)v, \quad (3)$$

Here v is the proportionality coefficient that has the dimension of velocity. This coefficient is the same for all carbon isotopes with the accuracy of isotopic fractionation corrections, which are of the order of several percent.

The equation $d{}^{\vee}\text{C}/dt = 0$ had held for the stationary equilibrium preindustrial steady state, and the influxes ${}^{\vee}F_0$ into the reservoir coincided with the escape fluxes from the reservoir:

$${}^{\vee}F_0 = {}^{\vee}\text{C}(s_0^-)v = {}^{\vee}\text{C}(d)H/{}^{\vee}T \quad (4)$$

The fluxes ratio, ${}^{14}F_0^{\pm}/{}^{12}F_0^{\pm}$, is proportional to inorganic concentration ratio, ${}^{14}\text{C}/{}^{12}\text{C} =$

$^{14}R(s_0^-)$ in the (s_0^-) region. Certain chemical reactions take place there, so that carbon is channelled to both the organic and inorganic reservoirs from the dissolved gas (CO_2). We have from (3), (4) for both the “+” and “-” reservoirs:

$$\frac{\tau}{T} = \frac{^{14}R(d)}{^{14}R(s_0^-)} = \frac{\Delta^{14}C(d)+1}{\Delta^{14}C(s_0^-)+1} \quad (5)$$

$$\Delta^{14}(x) \equiv \frac{^{14}R(x)}{^{14}R(a_0)} - 1, \quad (6)$$

$$^{14}R(x) \equiv ^{14}C(x)/^{12}C(x),$$

where $^{14}R(x)$ and $^{14}R(a_0)$ are the $^{14}\text{C}/^{12}\text{C}$ ratios for one and the same reservoir (+ or -) in the observed region, x , and at the equilibrium with the preindustrial atmosphere, a_0 , respectively [17].

To simplify our calculations we use the following notation below:

$$\Delta^{14}C(x) \equiv \Delta(x), \quad ^{14}R(x) \equiv R(x). \quad (7)$$

Generating the ratio $^{14}F_0/^{12}F_0$ in (3) and (4) and using expressions (5) and (7) we obtain:

$$\begin{aligned} \tau^\pm &\equiv \frac{^{14}T^\pm}{T_c} = \frac{\Delta(s_0^\pm) - \Delta(d^\pm)}{1 + \Delta(s_0^\pm)} \\ T^\pm &\equiv \frac{^{12}T^\pm}{T_c} = \frac{\Delta(s_0^\pm) - \Delta(d^\pm)}{1 + \Delta(d^\pm)} \end{aligned} \quad (8)$$

Using data of Druffel and Suess [18] for $\Delta(s_0^-)$, data of Druffel and Williams [11] for $\Delta(d^-)$ and $\Delta(d^+)$, and (8) we have:

$$\begin{aligned} \Delta(s_0^-) &= -0.05, \quad \Delta(d^-) = -0.23; \quad \Delta(d^+) = -0.53; \\ \tau^+ &= 0.51, \quad T^+ = 1.02; \quad \tau^- = 0.19, \quad T^- = 0.23. \end{aligned} \quad (9)$$

The surface equilibrium value $\Delta(s_0^+)$ is not known. It may be calculated from the condition that the flux of carbon to oceanic depths due to eddy diffusion was equal to influxes and escape fluxes from the reservoir during the preindustrial era:

$$^{14}F_0^\pm = \frac{D}{L} [^{14}C(s_0^\pm) - ^{14}C(d^\pm)] = \frac{^{14}C(d^\pm)H}{^{14}T^\pm} \quad (10)$$

Here D is eddy diffusivity and L is the depth above which the gradient ^{14}C concentration is observed. The values D , L and H remains the same for both the organic and inorganic carbon. So using (8) and (10) we have:

$$\frac{\Delta(s_0^+) - \Delta(d^+)}{1 + \Delta(d^+)} \tau^+ = \frac{\Delta(s_0^-) - \Delta(d^-)}{1 + \Delta(d^-)} \tau^- \quad (11)$$

Finally, we have from (11) and (9):

$$D(s_0^+) = -0.49. \quad (12)$$

The surface equilibrium value $\Delta(s_0^+)$ (12) does not deviate considerably from the depth value $\Delta(d^+)$ (9). Therefore the residence time of surface equilibrium DOC fraction is approximately equal to the average DOC residence time T^+ (9).

Modern oceanic DOC production

The modern value $\Delta(s_0^+) = -0.18$ for surface DOC was found by Druffel and Williams [11]. Let us denote the increase (increment) of surface DOC concentration during the whole industrial era as $^v c(s^+)$. Then we have for $\Delta(s^+) + 1$ according to definitions (6) and (7):

$$\begin{aligned} \Delta(s^+) + 1 &\equiv \frac{R(s^+)}{R(a_0)} = 0.82; \\ R(s^+) &= \frac{^{14}C(s_0^+) + ^{14}c(s^+)}{^{12}C(s_0^+) + ^{12}c(s^+)} \end{aligned} \quad (13)$$

According to (12) during the preindustrial era the equilibrium value $\Delta(s_0^+) + 1$ was equal to

$$\Delta(s_0^+) + 1 = 0.51. \quad (14)$$

The ratio of (13) and (14) is equal to

$$\frac{\Delta(s^+) + 1}{\Delta(s_0^+) + 1} = 1.6. \quad (15)$$

If in (13) we took zero values for concentration increments of ^{12}C and ^{14}C , $^{12}c(s^+) = ^{14}c(s^+) = 0$, then, evidently, the value of $\Delta(s^+)$ would be equal to the equilibrium preindustrial value, $\Delta(s^+) = \Delta(s_0^+) = -0.49$ (12), and ratio (13) would be equal to unity. Bomb radiocarbon had been increasing ^{14}C concentration in surface DIC during the eight years (1955–

1963) of the most intensive atmospheric nuclear tests. The maximum relative increment of ^{14}C concentration in DIC never exceeded 10% [19], i.e. the average ratio $(\Delta(s^-) + 1)/(\Delta(s_0^-) + 1) \leq 1.1$. If the observed change in ^{14}C concentration had been due to bomb radiocarbon increment only and ^{12}C mass in DOC remained constant, ratio $[\Delta(s^+) + 1]/[\Delta(s_0^+) + 1]$ would have coincided with ratio $[\Delta(s^-) + 1]/[\Delta(s_0^-) + 1]$ and deviated from unity by the value of relative increment of surface DIC concentration, i.e. by less than 10% (Fig. 1). Thus, bomb radiocarbon cannot be the cause of the observed deviation (15).

The only possible explanation of the observed value of deviation of ratio (15) from unity is the assumption that during the industrial era (the last 200 years) DOC was driven away from the preindustrial equilibrium state and an additional mass of DOC has accumulated during this period. As the ^{14}C lifetime is of the order of ten thousand years (2), we can consider DOC that accumulated during the last 200 years as having zero age, i.e. $R(a_0) \equiv R(a)$. Then the industrial ^{14}C and ^{12}C increments $^{14}c(s^+)$ and $^{12}c(s^+)$ will be related as follows:

$$^{14}c(s^+) = R(a_0) ^{12}c(s^+), \quad (16)$$

where $R(a_0)$ is equal to ratio in the preindustrial atmosphere, see (6). Relation (16) reflects the evident fact that the $^{14}\text{C}/^{12}\text{C}$ ratio in the DOC increment will be the same as in the atmosphere, because the new DOC comes from DIC which is in equilibrium with atmospheric CO_2 . Using definitions (6) and (7) and expression (12) we have:

$$^{14}C(s_0^+) \equiv R(s_0^+) ^{12}C(s_0^+), \quad (17)$$

$$\Delta(s_0^+) = R(s_0^+)/R(a_0) - 1 = -0.49.$$

Using relations (13), (16) and (17) we obtain for the $^{12}c(s^+)/^{12}C(s_0^+)$ ratio:

$$^{12}c(s^+)/^{12}C(s_0^+) = \frac{\Delta(s_0^+) - \Delta(s^+)}{\Delta(s^+)} = 1.7, \quad (18)$$

More than 90% of the total atmospheric carbon increment has entered the atmosphere dur-

ing this century, i.e. time period of the order of 100 years. Bomb radiocarbon had been entering atmosphere during time period less than 10 years, i. e. less than 10% of the total period of DOC mass increase. Due to bomb radiocarbon emissions oceanic ^{14}C concentration increased by less than 10%, (Fig. 1). Consequently, the contribution of bomb radiocarbon into (18) is less than 1%.

The total relative increment of DOC mass is about one tenth of value (18), i.e. about 17%. Indeed, the total relative increment of DOC concentration above 800 m is about 1/2 of surface relative increment of DOC concentration, see (Fig.1). The average oceanic depth is about 4 km. So for total oceanic relative increment we have: $\gamma = 1.7 (1/2)(0.8/4) = 0.17$.

Note that we can find only relative DOC increment by such approach. In order to find the absolute mass of DOC increment we have to know the total oceanic DOC mass.

The traditional estimate of DOC mass, M_T^{DOC} , is equal to 700 GtC [4,20]. The maximum published estimate of DOC mass equals 2000 GtC [21,22]. To take into account the uncertainty in determination of the true DOC mass value, we enter coefficient Z , which is equal to the ratio of the true DOC mass to M^{DOC} its traditional estimate M_T^{DOC} . Then the absolute increment of DOC mass m^{DOC} added to the preindustrial equilibrium value M^{DOC} can be written as follows:

$$m^{\text{DOC}} = \gamma M^{\text{DOC}} = 120 \text{ GtC} \cdot Z; \\ Z \equiv \frac{M^{\text{DOC}}}{M_T^{\text{DOC}}} \geq 1. \quad (19)$$

The relative increment of DOC mass, γ , does not exceed 20%. Thus, with an accuracy of 20%, we can put in (19) the preindustrial DOC mass equal to its modern value.

The rate of change of DOC mass, $S^+ \equiv dM^{\text{DOC}}/dt$, depends on the known atmospheric increment of carbon mass, $m_a(t) = M_a(t) - M_{a0}$ (M_{a0} — carbon mass in the preindustrial atmosphere), and is proportional to this increment when its relative value is small, i.e. when $m_a(t)/M_{a0} \ll 1$ [4,23]. So we can write:

$$S^+ = dM^{DOC}/dt = k_s^+ m_a(t). \quad (20)$$

Here k_s^+ is the constant proportionality coefficient. Using the data for $m_a(t)$ [24] and integrating equation (20) over time t at constant k_s^+ we find

$$120 \text{ GtC} \cdot Z = m^{DOC} = \gamma M^{DOC} = k_s^+ \int_{1800}^{1986} m_a(t) dt = k_s^+ 9200 \text{ GtC} \cdot \text{yr}^{-1}, \quad (21)$$

$$k_s^+ = 0.013 \cdot \text{yr}^{-1} \cdot Z \quad (22)$$

The modern rate of DOC accumulation, dM^{DOC}/dt , can be obtained from (20) using the modern observed value of atmospheric carbon increment $m_a(1995) \cong 170 \text{ GtC}$, which is about 28% of equilibrium preindustrial atmospheric carbon mass [4]. Assuming that coefficient (22) remains constant at relatively small modern atmospheric carbon increment we have:

$$S^+ \cong dM^{DOC}/dt = k_s^+ m_a(1995) \cong 2.3 \text{ GtC}/\text{yr} \cdot Z. \quad (23)$$

The total mass of particulate organic carbon (POC) is of the order of several per cent of the total DOC mass [11,25]. The possible rate of POC mass change due to change of POC production, sinking or destruction is in any case negligibly small as compared with the obtained rate of DOC mass change (23).

Dissolved inorganic carbon change in the ocean

The rate of change of DIC mass, $S^- = dM^{DIC}/dt$, should be proportional to the atmospheric increment of carbon mass. So similarly to (20) we have:

$$S^- = dM^{DIC}/dt = k_s^- m_a(t). \quad (24)$$

Taking $dM^{DIC}/dt = 2.0 \text{ GtC}/\text{yr}$ [13,14] and $m_a(t) = 130 \text{ GtC}/\text{yr}$ [24] averaged for the years 1970-1990 we have:

$$k_s^- = \frac{\overline{dM^{DIC}/dt}}{\overline{m_a(t)}} = 0.015 \text{ yr}^{-1} \quad (25)$$

Similarly to (25) we can retrieve the modern figure for the rate of inorganic carbon change in the ocean, dM^{DIC}/dt :

$$S^- \equiv dM^{DIC}/dt = k_s^- m_a(1995) = 2.6 \text{ GtC}/\text{yr} \quad (26)$$

Global carbon and oxygen budget

According to the law of matter conservation we can represent the global budget for atmospheric carbon as

$$A + F + B + S = 0, S = S^- + S^+. \quad (27)$$

The budget terms are annual averaged rates of carbon content change in the main global carbon reservoirs A — atmosphere, F — fossil fuel, B — terrestrial biota, S — ocean. The oceanic term S has the inorganic component S^- (which describes change in DIC mass) and the organic term S^+ (which describes change in DOC mass due to activity of the oceanic biota). Sinks of carbon will enter (27) as positive values, sources of carbon as negative values. The corresponding oxygen changes can be represented in a similar equation of the global budget for atmospheric oxygen (low index O):

$$A_o + F_o + B_o + S_o = 0, S_o = S_o^- + S_o^+. \quad (28)$$

Processes like fossil fuel burning or destruction of biomatter are accompanied by binding of atmospheric oxygen and thus represent sources of atmospheric carbon but sinks of atmospheric oxygen. Therefore terms in (28) will have opposite signs as compared to the corresponding terms in (27).

The observed negative change in atmospheric oxygen A_o is known from measurements of atmospheric O_2/N_2 ratio changes [6]. The remaining terms in (28) are connected to the corresponding terms in (27) through stoichiometric coefficients $\alpha = O_2/CO_2$. For burning of fossil fuels $\alpha = \alpha_f = 1.38 \pm 0.04$ [6]. For synthesis and destruction of organic matter by terrestrial biota $\alpha = \alpha_B = 1.10 \pm 0.05$ [6]. For oceanic biota $\alpha = \alpha_{S^+} = 1.30 \pm 0.03$ (Redfield ratio) [15,26]. Oceanic uptake of carbon in inorganic form is assumed to have no considerable effect on atmospheric oxygen concentration [6], so $\alpha_{S^-} = 0$. Thus eq. (28) may be rewritten as

$$A_o \cdot (\alpha_f F / F_o) - \alpha_f F - \alpha_B B - \alpha_{S^+} S^+ = 0, \quad (29)$$

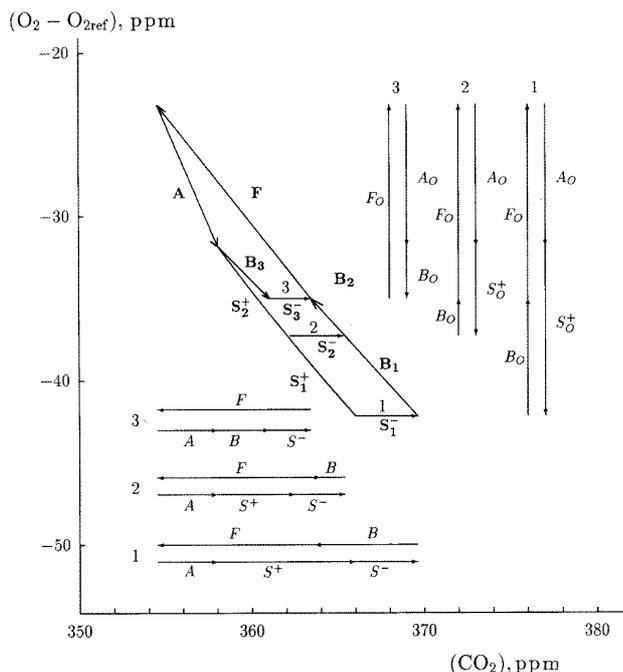


Fig. 2. Changes in the atmospheric O_2 concentration versus changes in CO_2 concentrations in the main global carbon reservoirs. A — atmosphere, F — fossil fuel, B — terrestrial biota, S^+ — oceanic biota (organic carbon), S^- — ocean (inorganic carbon). $O_{2,ref}$ is a fixed reference O_2 concentration [27]. For carbon 1 ppm = 2.1 GtC. Vectors and their projections on both axes connect data for July 1991 and July 1994. Slope of vector A is determined by the observed atmospheric CO_2 and O_2 changes [6]. Other slopes are stoichiometric O_2/CO_2 ratios for processes in different reservoirs, see eqs. (28) and (29). Balances of the vector projections on CO_2 and O_2 axes are given below and to the right of the closed diagram (low index «O» stands for oxygen). Vectors S_1^+ and B_1^+ are found in this paper solving (31), (32) at $S_1^- = 2.6 \text{ GtC yr}^{-1}$ (26), see (34). Vectors S_2^+ and S_2^- are found in this paper solving (31)–(32) at $B = -1.1 \text{ GtC yr}^{-1}$. Vectors B_3 and S_3^- are found by Keeling et al. [6] under the assumption that the oceanic biota plays no role in the budget (solution of system (31)–(32) at $S_3^+ = 0$), see also Table.

where atmospheric term A_0 is multiplied by transforming coefficient ($\alpha_F F / F_0$) which turns units of oxygen change measurements ($\text{ppm} \cdot \text{yr}^{-1}$) into units of carbon change measurements ($\text{GtC} \cdot \text{yr}^{-1}$), (Fig. 2).

For the 1991–1994 time period the annual averaged rate of fossil fuel burning F , the rate of atmospheric oxygen depletion due to fossil fuel burning F_0 and the observed rates of carbon accumulation and oxygen depletion in the atmosphere A and A_0 are as follows [6]*):

$$F = (-5.9 \pm 0.3) \text{ GtC} \cdot \text{yr}^{-1},$$

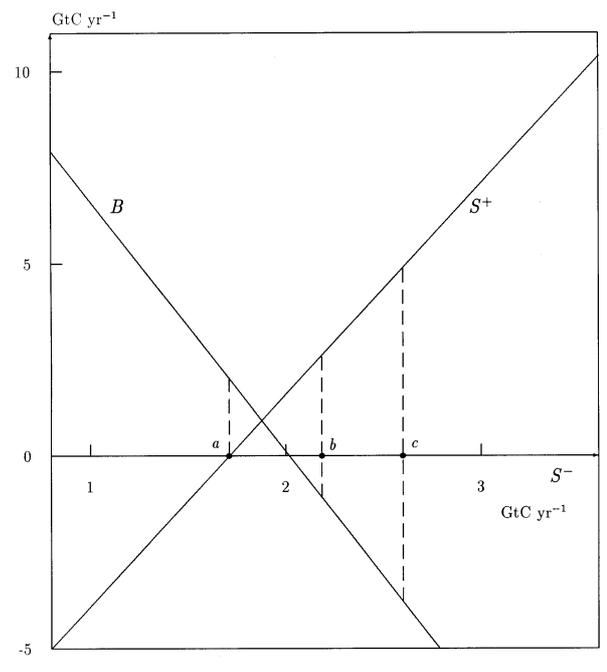


Fig. 3. Changes in organic carbon mass due to biotic impact in the ocean (S^+) and on land (B) versus changes in inorganic carbon mass in the ocean (S^-), see eqs. (31)–(32).

$$F_0 = (4.0 \pm 0.2) \text{ ppm} \cdot \text{yr}^{-1},$$

$$A = (2.2 \pm 0.2) \text{ GtC} \cdot \text{yr}^{-1}, \quad (30)$$

$$A_0 = (-2.9 \pm 1.3) \text{ ppm} \cdot \text{yr}^{-1}$$

(Keeling et al. [6] used per meg $\cdot \text{yr}^{-1}$ as units of O_2 change measurements. For oxygen 1 ppm = 4.8 per meg, [27]). Uncertainty ranges represent estimated standard errors. Taking these values and solving eqs. (27) and (29) for S^+ and B we have (Fig. 3):

$$S^+ = -9.4 \text{ GtC} \cdot \text{yr}^{-1} + 5.5 S^-, \quad (31)$$

$$B = 13.1 \text{ GtC} \cdot \text{yr}^{-1} + 6.5 S^-. \quad (32)$$

Relative uncertainties of all figures in (31) and (32) are about 10%. Excluding S^- from (31)–(32) yields relation between the rates of changes in mass of organic carbon on land (B) and in the ocean (S^+):

$$B = 2.0 \text{ GtC} \cdot \text{yr}^{-1} - 1.2 S^+. \quad (33)$$

We will now take in turn the available estimates for S^+ , B , S^- as the starting points and discuss possible solutions of system (31)–(32).

*We are indebted to M. Heimann for providing us with corrected figures.

Table

Available estimates of budget terms and different presentations of the global carbon budget as a whole (GtC yr⁻¹). All data are rates of carbon release (negative values) and uptake (positive values) averaged over time periods indicated in the table. Uncertainty ranges represent 90% confidence intervals as in refs. [1,6]

Sources of data	Ref. [29], 1980–1989	Ref. [6], 1991–1994	This paper, 1991–1994
INORGANIC CARBON (<i>IC</i>):			
Observed change in the atmosphere <i>A</i>	3.3 ± 0.2	2.2 ± 0.1	2.2 ± 0.1 (ref. 6)
Fossil fuel burning and cement roduction <i>F</i>	-5.5 ± 0.5	-5.9 ± 0.5	-5.9 ± 0.5 (ref. 6)
Change in the ocean <i>S</i> ⁻	2.0 ± 0.8	1.7 ± 0.9	2.6 ± 1.0 (eq. 26)
ORGANIC CARBON (<i>OC</i>): <i>OC</i> = - <i>IC</i> = - (<i>A</i> + <i>F</i> + <i>S</i> ⁻)			
Oceanic biota <i>S</i> ⁺	0	0	4.9 ± 2.1 (eq. 34)
Terrestrial biota <i>B</i> <i>B</i> = <i>OC</i> - <i>S</i> ⁺	0.2 ± 0.5	2.0 ± 0.9	-3.8 ± 1.9 (eq. 34)
tropical zone <i>B</i> _{<i>t</i>}	-1.6 ± 1.0	-1.6 ± 1.0 (ref. 29)	
boreal zone <i>B</i> _{<i>b</i>}	0.5 ± 0.5	0.5 ± 0.5 (ref. 29)	
remainder <i>B</i> _{<i>r</i>} <i>B</i> _{<i>r</i>} = <i>B</i> - <i>B</i> _{<i>t</i>} - <i>B</i> _{<i>b</i>}	1.3 ± 1.5	3.1 ± 1.5	0

Firstly, if we consider $S^- = 2.6 \text{ GtC}\cdot\text{yr}^{-1}$ (26), point *c* in Fig. 3, to be a good estimate for the modern rate of oceanic carbon uptake in inorganic form, we get for decrease of organic carbon mass on land and increase of it in the ocean, respectively:

$$B = -3.8 \text{ GtC}\cdot\text{yr}^{-1} \text{ and } S^+ = 4.9 \text{ GtC}\cdot\text{yr}^{-1}. \quad (34)$$

The former figure in (34) is about two times larger than the modern estimate of tropical terrestrial biotic carbon source [1,12], but is consistent with earlier estimates for the rate of decrease of the terrestrial organic carbon stock [28]. The latter figure in (34) corresponds to the following *Z* in (23), k_s^+ , in (22) and total DOC mass in (19):

$$Z = 2.1; k_s^+ = 0.028\cdot\text{yr}^{-1}; M^{\text{DOC}} = 1500 \text{ GtC}, \quad (35)$$

which is close to the upper estimates of total DOC mass (about 2000 GtC [21,22]). Though one of the authors [21] has retracted this value [22], our result can be thought of as pointing to the reliability of these maximum estimates of the total DOC mass.

Secondly, according to the available land use and forest statistics terrestrial biota as a whole represents a source of carbon at 1.1

GtC·yr⁻¹ (release of carbon dioxide due to land use at 1.6 GtC·yr⁻¹ (in tropics mostly) and accumulation of organic carbon due to regrowth of temperate latitude forests at 0.5 GtC·yr⁻¹) [1,6,12]. If we take $B = -1.1 \text{ GtC}\cdot\text{yr}^{-1}$ as a reliable estimate (point *b* in Fig. 3), solving (31)–(32) yields $S^+ = 2.6 \text{ GtC}\cdot\text{yr}^{-1}$ and $S^- = 2.2 \text{ GtC}\cdot\text{yr}^{-1}$ for the rates of oceanic uptake of carbon in organic and inorganic form, respectively (Fig. 2). These figures agree within uncertainty ranges with independent estimates of $S^+ = (2.3 \pm 1.0) \text{ GtC}\cdot\text{yr}^{-1}$ (23) and $S^- = (2.6 \pm 1.0) \text{ GtC}\cdot\text{yr}^{-1}$ (26) obtained from ¹⁴C/¹²C and ¹³C/¹²C measurements under the assumption that the rate of oceanic carbon uptake grows with time proportionally to the increase of atmospheric CO₂ concentration as compared to the preindustrial state (Table).

Uncertainty of the ¹⁴C/¹²C estimates is mostly determined by the uncertainty of the estimates of the total mass of DOC in the ocean (23). The traditional estimate of total DOC mass is 700 GtC [4,8] which now corresponds to $S^+ = 2.3 \text{ GtC}\cdot\text{yr}^{-1}$. Our result, $S^+ = 2.6 \text{ GtC}\cdot\text{yr}^{-1}$, corresponds to total DOC mass equal to 790 GtC, which falls within uncertainty ranges of traditional DOC mass estimates [4,8].

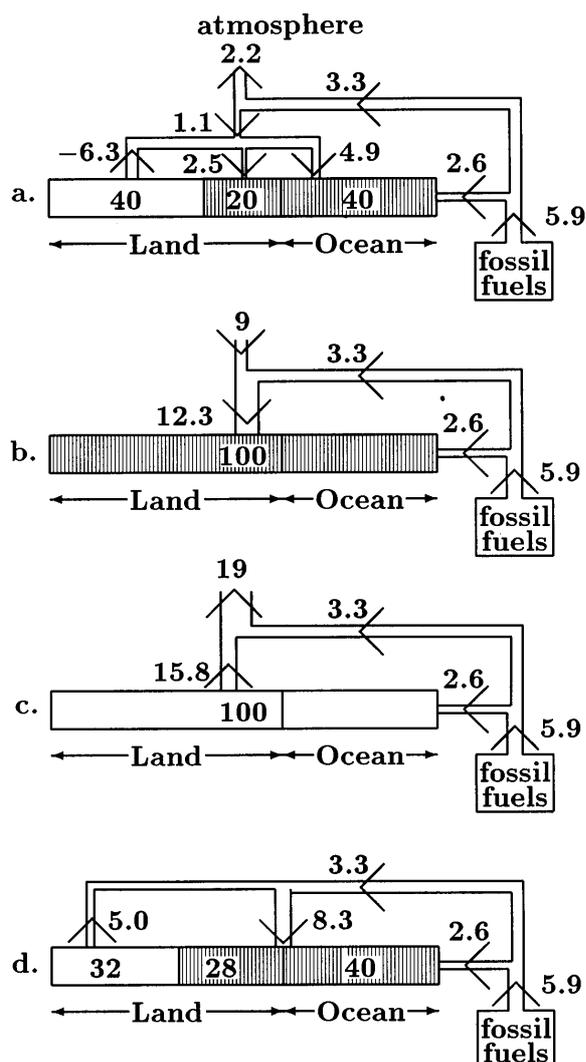


Fig. 4. The global change and virgin biota. The global carbon cycle change: a — for modern biosphere; b — for totally unperturbed land+ocean biota; c — for totally man perturbed land+ocean biota; d — for stopping of the global environmental change. Shaded area is the virgin biota, blank area is the perturbed biota. Figures near arrows are the carbon fluxes in GtC yr^{-1} . Figures in shaded and blank areas are net primary productions in GtC yr^{-1} .

Thirdly, if we assume that oceanic biota does not react to the increasing atmospheric CO_2 concentration at all ($S^+ = 0$, point a in Fig. 3), we arrive at the result obtained by Keeling et al. [6] They found a global oceanic inorganic sink $S^- = 1.7 \text{ GtC}\cdot\text{yr}^{-1}$ and a global terrestrial biotic sink $B = 2.0 \text{ GtC}\cdot\text{yr}^{-1}$ (Fig. 2). The former estimate obtained for the 1991-1994 time period appears to be less than the estimate $2.0 \text{ GtC}\cdot\text{yr}^{-1}$, which was obtained from $^{13}\text{C}/^{12}\text{C}$ ratio measurements for an earlier 1970-1990

time period [12,14] and which increases up to $2.6 \text{ GtC}\cdot\text{yr}^{-1}$ if the time growth of this rate is taken into account (Table, see also [30]).

Modern data on land use [1,6,12] tell that exploited lands (tropics mostly) represent a source of carbon at $1.6 \text{ GtC}\cdot\text{yr}^{-1}$. If the global biota as a whole represents a sink at $B = 2.0 \text{ GtC}\cdot\text{yr}^{-1}$ then the remaining part of terrestrial biota should remove from the atmosphere $3.6 \text{ GtC}\cdot\text{yr}^{-1}$, i.e. about two thirds of what is released from burning of fossil fuels. Among terrestrial ecosystems there are no possible candidates that could ensure such a powerful sink of carbon [1,5,31] (according to forest statistics, ecosystems of boreal forests which represent most part of the temperate latitude biota in the Northern hemisphere remove at best $(0.5 \pm 0.5) \text{ GtC}\cdot\text{yr}^{-1}$).

Conclusions

Summing up, the existence of a substantial oceanic biotic sink comparable by the order of magnitude with the modern fossil fuel source does not contradict any available data on major factors determining the global carbon cycle. The unperturbed oceanic biota has thus stabilising influence upon Earth's changing environment. On the contrary, terrestrial biota has probably lost the stabilising ability being substantially destroyed by anthropogenic activities and is now adding significantly to the detrimental environmental perturbations. Our results suggest that stabilising negative feedback loops between the physical environment and the natural biota constitute an essential feature of the biota-environment interrelationship.

The net primary production of the ocean is about $40 \text{ GtC}\cdot\text{yr}^{-1}$ and constitute about 40% of the net primary production of the whole biosphere which is about $100 \text{ GtC}\cdot\text{yr}^{-1}$ [4,23]. Anthropogenically perturbed land areas now amount to 60% of the total land surface area [4,23]. The areas of land wilderness constitute 40% of that area [4,23]. Thus the net primary production of the land wilderness and perturbed land area is about 20% and 40% of the net primary production of the whole bio-

sphere, respectively. It is naturally to assume that negative feedback reaction of virgin biota is proportional to the net primary production. If we assume that the virgin terrestrial biota reaction to the anthropogenic perturbation of the environment is similar to that of the virgin oceanic biota S^+ (34), we find that at present virgin terrestrial biota can absorb $(4.9 \text{ GtC}\cdot\text{yr}^{-1}/0.40)\cdot 0.20 \cong 2.5 \text{ GtC}\cdot\text{yr}^{-1}$. Overall, the land biota currently emits about $3.8 \text{ GtC}\cdot\text{yr}^{-1}$ to the atmosphere (34). That means that the biota in the perturbed land areas emits about $6.3 \text{ GtC}\cdot\text{yr}^{-1}$, see Table and Fig. 4a.

Were we able return the whole terrestrial biota, i. e. the whole biota of the biosphere, to its unperturbed state, it would absorb about $4.9 \text{ GtC}\cdot\text{yr}^{-1}/0.40 \cong 12.3 \text{ GtC}\cdot\text{yr}^{-1}$ from the modern atmosphere and could compensate total carbon emission from the present-day fossil fuels combustion. The atmospheric CO_2 concentration would decrease at a rate of $9 \text{ GtC}\cdot\text{yr}^{-1}$ due to carbon absorption by the ocean in inorganic form and the preindustrial equilibrium state could be restored in seven-eight decades, (Fig. 4b).

Were the whole terrestrial biota perturbed, it would emit about $6.3 \text{ GtC}\cdot\text{yr}^{-1}/0.69 \cong 10.5 \text{ GtC}\cdot\text{yr}^{-1}$, so that the atmospheric CO_2 concentration would increase at a rate of about $13.8 \text{ GtC}\cdot\text{yr}^{-1}$. Were the biota of the whole biosphere including ocean completely perturbed, it would emit about $6.3 \text{ GtC}\cdot\text{yr}^{-1}/0.40 \cong 15.8 \text{ GtC}\cdot\text{yr}^{-1}$ so that the atmospheric CO_2 concentration would increase at a catastrophic rate of about $19 \text{ GtC}\cdot\text{yr}^{-1}$, (Fig. 4c). The complete abandonment of the combustion of fossil fuels would reduce this value down to $6 \text{ GtC}\cdot\text{yr}^{-1}$ only. Thus the state of global environment strongly depends on the state of natural biota in the biosphere.

Using this fact, it is possible to stop the global environmental changes of today even if the modern rate of fossil fuel burning remains the same. This can be done by increasing the area occupied by virgin biota at the expense of anthropogenically perturbed territories. To ensure the desired effect, it is sufficient to reduce the anthropogenically perturbed area from 60% to 48%, which corresponds to a 30% reduction in exploitation of forests, (Fig. 4d).

References

- Melillo J.M., Prentice I.C., Farquhar G.D., Schulze E.-D. and Sala O.E. (1996) Terrestrial biotic responses to environmental change and feedbacks to climate. In: Houghton, J. T., Meira Filho, L. G., Callander, B. A. et al. (eds). *Climate Change 1995* IPCC, Cambridge: Cambridge Univ. Press, pp. 449–481.
- Walker B., Steffen W. (eds.) (1997) Terrestrial ecosystem interactions with global change, *IGBP Science*. 8–14.
- Falkowski P.G. and Wilson, C. (1992) Phytoplankton productivity in the North Pacific ocean since 1900 and implications for absorption of anthropogenic CO_2 . *Nature*. **358**, 741–743.
- Siegenthaler U. and Sarmiento J.L. (1993) Atmospheric carbon dioxide and the ocean. *Nature*. **365**, 119–125.
- Ciais P., Tans P.P., Trolier M., White J.W.C. and Francey R. (1995) A large northern hemisphere terrestrial CO_2 sink indicated by $^{13}\text{C}/^{12}\text{C}$ of atmospheric CO_2 . *Science*. **269**, 1098–1102.
- Keeling R.F., Piper S.C. and Heimann M. (1996) Global and hemispheric CO_2 -sinks deduced from changes in atmospheric O_2 concentration. *Nature*. **381**, 218–221.
- Hunter K.A., Kim J.P. and Croot P.L. (1997) Biological roles of trace metals in natural waters. *Environ. Monitoring and Assess.* **44**, 103–147.
- Denman K., Hofmann E. and Marchant H. (1996) Marine biotic responses to environmental change and feedbacks to climate. In: Houghton, J. T., L. G. Meira Filho, B. A. Callander et al. (eds). *Climate Change 1995*, IPCC, Cambridge: Cambridge Univ. Press, pp. 487–516.
- Fogg G.E. (1975) Biochemical pathways in unicellular plants. In: Cooper J.P. (ed). *Photosynthesis and Productivity in Different Environments*. Cambridge: Cambridge Univ. Press.
- De Voogs C.G.N. (1979) Primary production in aquatic environments. In: B.Bolin, E.T.Degens, S.Kempe and P.Ketner (eds). *The Global Carbon Cycle*. New York: J.Wiley, pp. 259–292.
- Druffel E.R.M. and Williams P.M. (1990) Identification of a deep marine source of particulate organic carbon using bomb ^{14}C . *Nature*. **347**, 172–174.
- Shimel D., Alves D., Enting I. et al. (1996) Radiative forcing of climate change. In: Houghton, J.T., L.G. Meira Filho, B.A. Callander et al., (eds) *Climate*

- Change 1995*, IPCC, Cambridge: Cambridge Univ. Press, pp. 65–131.
13. Quay P.D., Tilbrook B. and Wong C.S. (1992) Oceanic uptake of fossil fuel CO₂: carbon-13 evidence. *Science*. **256**, 74–79.
 14. Heimann M. and Maier-Reimer E. (1996) On the relations between the oceanic uptake of carbon dioxide and its carbon isotopes. *Global Biogeochem. Cycles*. **10**, 89–110.
 15. Redfield, A. (1958) The biological control of chemical factors in the environment. *Amer. Nat.* **46**, 205–221.
 16. Degens E.T., Kempe S. and Spitzzy A. (1984) Carbon Dioxide: a Biological Portrait. In: O.Hutziger (ed.) *The Handbook of Environmental Chemistry*, 1, C. Berlin: Springer-Verlag, pp. 127–215.
 17. Stuiver M. and Pollach H. (1977) Reporting of ¹⁴C data. *Radiocarbon*. **19**, 355–363.
 18. Druffel E.R.M. and Suess H.E. (1983) On the radiocarbon record in banded corals: Exchange parameters and net transport of ¹⁴CO₂ between atmosphere and surface ocean. *J. Geophys. Res.* **88**, 1271–1280.
 19. Nydal R., Gulliksen S., Lovseth K. and Skogseth F.H. (1984) Bomb ¹⁴C in the ocean surface 1966–1981. *Radiocarbon*. **26**, 7–45.
 20. Martin J. H. and Fitzwater S. E. (1992) Dissolved organic carbon in Atlantic, Southern and Pacific oceans. *Nature*. **256**, 699–700.
 21. Sugimura Y. and Suzuki Y. (1988) A high temperature catalytic oxidation method for determination of non-volatile dissolved organic carbon in seawater by direct injection of liquid samples. *Marine Chemistry*. **24**, 105–131.
 22. Suzuki J. (1993) On the measurement of DOC and DON in seawater. *Marine Chem.* **41**, 287–288.
 23. Gorshkov V.G. (1995) *Physical and Biological Bases of Life Stability. Man, Biota, Environment*. Berlin: Springer-Verlag.
 24. Staffelbach T., Stauffer B., Sigg A. and Oeschger H. (1991) CO₂ measurements from polar ice cores: more data from different sites. *Tellus*. **43**, 91–96.
 25. Mopper K. and Degens E.T. (1979) Organic carbon in the ocean: nature and cycling. In: B.Bolin, E.T.Degens, S.Kempe and P.Ketner (eds.). *The Global Carbon Cycle*, New York: Wiley, pp. 293–316.
 26. Chen C.-T.A., Gong G.C, Wang S.L. and Bychkov A.S. (1996) Redfield ratios and regeneration rates of particulate matter in the Sea of Japan as a model of closed system. *Geophys. Res. Let.* **23**, 1785–1790.
 27. Keeling R.F. and Shertz R. (1992) Seasonal and interannual variations in atmospheric oxygen and implications for the global carbon cycle. *Nature*. **358**, 723–726
 28. Houghton R.A., Hobbie J.E., Melillo J.M. et al. (1983) Changes in the content of terrestrial biota and soils between 1860 and 1980: a net release of CO₂ to the atmosphere. *Ecol. Monogr.* **53**, 235–262.
 29. IPCC WGI (1996) Technical Summary. In: Houghton J.T., L.G. Meira Filho, B.A. Callander et al., (eds) *Climate Change 1995* IPCC, Cambridge: Cambridge Univ. Press, pp. 9–49.
 30. Gorshkov V.G. (1996) Biogenic sources and sinks of atmospheric CO₂. *Geochimia*. **8**, 767–774 (in Russian).
 31. Schlesinger W. H. (1990) Evidence from chronosequence studies for a low carbon-storage potential of soils. *Nature*. **348**, 232–234.