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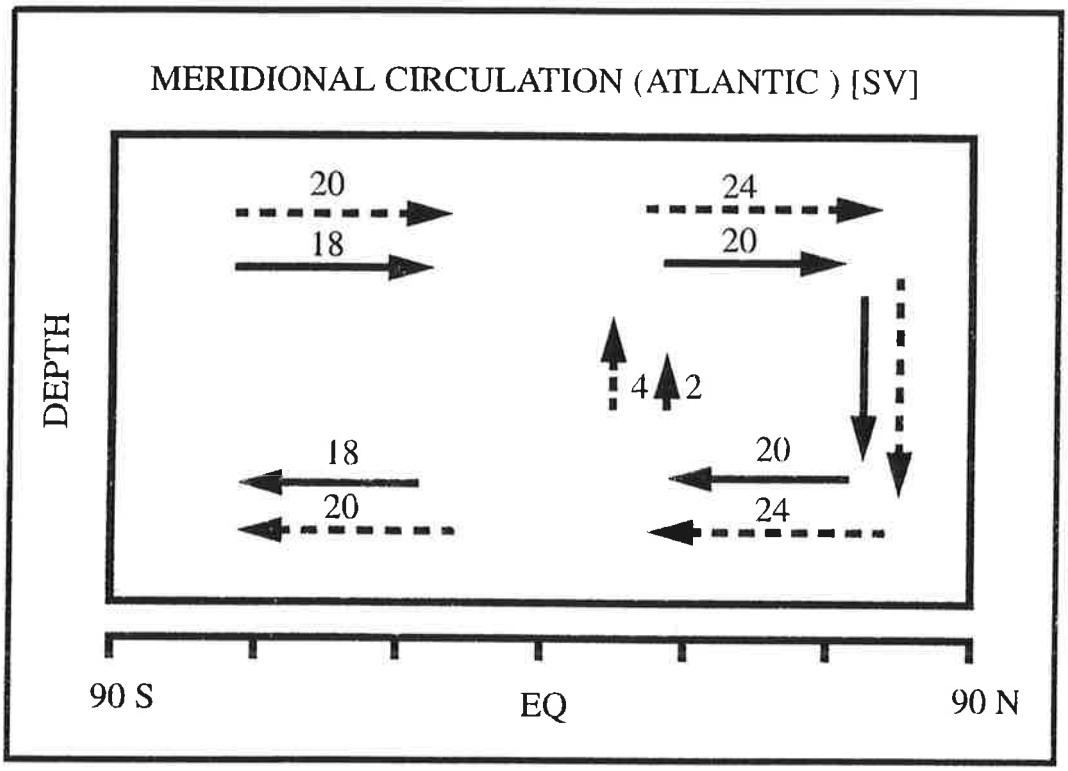


# INTERACTIONS BETWEEN OCEAN CIRCULATION AND THE BIOLOGICAL PUMPS IN THE GLOBAL WARMING

by

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# Interactions between ocean circulation and the biological pumps in the global warming

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Abstract

We discuss the potential variations of the biological pump that can be expected from a change in the oceanic circulation in the ongoing global warming. The biogeochemical model is based on the assumption of a perfect stoichiometric composition of organic material. Upwelling nutrients are transformed into organic particles that sink to the deep ocean according observed profiles. The physical circulation model is driven by the warming pattern as derived from scenario computations of a fully coupled ocean - atmosphere model. The amplitude of the warming is determined from the varying concentration of atmospheric $CO_2$.

The model predicts a pronounced weakening of the thermohaline overturning. This is connected with a reduction of the transient uptake capacity of the ocean. It yields also a more effective removal of organic material from the surface which partly compensates the physical effects of solubility. Both effects are rather marginal for the evolution of atmospheric $pCO_2$. Running climate models and carbon cycle models separately seems to be justified.

## I. Introduction

The increase of carbon dioxide in the atmosphere due to the combustion of fossil fuels has stimulated (Revelle and Suess, 1957) a broad spectrum of research. One of the key question concerns the amount of excess $CO_2$ that can be absorbed by the oceans. The combination of elementary inorganic chemistry with the oceanic volume gives an upper estimate of 5/6 for the storage capacity of the oceans, as already pointed out by Arrhenius at the end of the last century. Due to the slowness of the oceanic deep circulation this asymptotic value may be expected after several thousand years only. At present, it is estimated that approximately one third of the carbon dioxide emissions enter the ocean (Houghton et al., 1990). This number represents a compromise between model studies and estimates from direct observations.

Simplified box models to study the oceanic $CO_2$ uptake were developed already 20 years ago (Oeschger et al., 1975). More recently, three dimensional circulation models were used for such studies, (Maier-Reimer and Hasselmann, 1987 and Sarmiento et al., 1992). The results confirm more or less the previous estimates. The main shortcoming of these models (apart from an insufficient formation of North Atlantic deep water) was the complete neglect of all biological processes. The strength of the gradients of dissolved carbon dioxide (i.e. the sum of the species $CO_2$, $H_2CO_3$, $HCO_3^-$, and $CO_3^{--}$ which frequently is abbreviated as TIC or DIC(total or dissolved inorganic carbon) or $\Sigma CO_2$) in the ocean, is primarily due to biological processes; thermodynamical effects alone could explain only a quarter of the observed gradients (Maier-Reimer and Hasselmann, 1987). Without the biological pumps, the ocean would be chemically quite different from today's real ocean even in the abiotic components. Consequently, the basic state on which the anthropogenic perturbation was imposed, could not be compared with the distribution of carbon dioxide in the real ocean.

The role of biological processes in the uptake of fossil fuel carbon dioxide by the ocean is often misinterpreted. The naive sketch that the excess carbon dioxide in the atmosphere could be brought into the deep ocean by the biological pump is in striking contradiction to the fact that the carbon dioxide -

content of the atmosphere has remained almost perfectly constant during the ten thousand years preceding the onset of anthropogenic emissions (Neftel et al., 1982). The ongoing sedimentation of calcium carbonate at the sea floor and of organic matter on some shelf areas is in balance with river inputs. More quantitative arguments against the role of biology in the sequestering of additional $CO_2$ is given by Broecker (1991). The natural biological pumps are to first order in equilibrium with the upwelling products of the inverse processes (nutrient remineralization and calcite dissolution). In regions of high productivity, the carbon dioxide content of the surface water is depleted by only 15% of the deep sea value. It is hard to imagine that 10%increase of dissolved carbon dioxide (the increase that would be in equilibrium with a doubling in the atmosphere (cf. Revelle and Suess, 1 957)) could substantially improve the utilization of nutrients. This argument appears to be not valid for some diatom species which depend on neutral $CO_2$ whose contribution to the total dissolved $CO_2$ is in the order of 1% only. Riebesell et al. (1993) have indeed shown in laboratory experiments that under optimal growth conditions a reduction of $CO_2$ slows down the cell doubling rate. They did not show a pronounced increase in the growth rate under artificially enhanced $pCO_2$ and they did not show either any changes in the carbon fixation per nutrient unit (Wolf-Gladrow, 1994).

A positive feedback of marine biology to the $CO_2$-induced global warming has been, in contrary, identified by Frankignoulle et al. (1994) from the fact that in warmer water there is a higher likelihood for algae to form calcareous shells; the accompanying removal of alkalinity should then tend to increase the atmospheric $pCO_2$.

An indirect effect of biology on the uptake of fossil fuel carbon dioxide is provided by the existence of sediment layers of which the upper ten or twenty centimeters are bioturbated and, thus, still in close contact with the overlying water. Incidentally, this pool is estimated to contain the same amount of carbon in the form of $CaCO_3$ as the present estimates of coal resources (Broecker and Takahashi, 1977). During the uptake of excess carbon dioxide the surface water becomes more acid. When it reaches

contact with the sediment layer, it tends to dissolve the $CaCO_3$ sediment and, consequently, to neutralize the acidification and, thus, the excess carbon dioxide. In the modern ocean, carbonate sediments are preserved - in the global average - at depths shallower than 4 km. In equilibrium with a doubled carbon dioxide content in the atmosphere, the concentration of carbonate ions in the seawater will be halved and the calcite preservation level ("lysocline") will rise by several kilometers and thus expose the corresponding sediment layers to dissolution. The resulting effect of sediment dissolution on the atmospheric $pCO_2$ is a seemingly paradoxical reduction, caused by the fact that with the dissolved $CO_3^{--}$ (not $CO_2$ !) the double amount of alkalinity is added to the system. Fig.1 shows the response of atmospheric pCO2 of different versions of the HAMOCC (Hamburg model of oceanic carbon cycle) to a sudden doubling of atmospheric pCO2. The inorganic version (Maier-Reimer and Hasselmann, 1987) and the first order biota version without the interactive sediment pool (Bacastow and Maier-Reimer, 1990) behave very similar despite the completely different structure of $\Sigma CO_2$ in the undisturbed state. In both models, the inventories of alkalinity were tuned to yield a preindustrial atmospheric $pCO_2$ of 278 ppm. With the assumption of a constant carbon to nutrient (Redfield-) ratio, the behaviour should be to first order identical. The response curves diverge only within the strength of the line drawing, due to the nonlinearity of the inorganic chemistry. The inclusion of an interactive sediment pool (Heinze et al., 1991) modifies the uptake characteristics substantially after a few centuries. The figure displays also the response of one of the simplified models (Siegenthaler, 1983), incidently the model with the highest uptake rate.

A direct effect of the biological pump on the uptake of carbon dioxide can be expected only from changes in the pumping mechanisms. Such changes could result in a modified thermodynamic field through changes in the plankton communities or through a changed residence time of particulate matter in the deep ocean (cf.Heinze et al., 1991).

The present generation of climate models (cf. Cubasch et al., 1992) use IPCC (Houghton et al., 1990) scenarios for increase of

carbon dioxide as determined from uptake scenarios using steady state ocean models. In this approach, potential variations of the oceanic uptake resulting from a changed oceanic circulation are ignored, since it would imply the computationally ineffective combination of the huge memory requirements of the carbon cycle model with the huge CPU-time requirements of the atmosphere model. In view of the unability to explain the glacial - interglacial variations of atmospheric pCO2 by plausible oceanic processes (Heinze et al. 1991) in stationary states with fixed tracer inventories, it is expected that the feedback of the changing oceanic circulation on the atmospheric concentration of carbon dioxide is small. In a transient state, however, the excursions of atmospheric $pCO_2$ could be more substantial (Archer and Maier-Reimer, 1994).

For an assessment of potential feedbacks during the next century, we performed experiments with the Hamburg model of the oceanic carbon cycle (HAMOCC3, Maier-Reimer, 1993), coupled online to the circulation model (Maier-Reimer et al., 1993) via an empirical linear feedback mechanism between atmospheric $CO_2$ and the atmospheric temperature that acts as a driving force of the ocean circulation. We look for the increase of atmospheric $pCO_2$ as resulting from a high emission scenario in an experiment with prescribed constant seasonally varying ocean circulation (corresponding to traditional $CO_2$ uptake experiments like Maier-Reimer and Hasselmann (1987), or Sarmiento et al., (1992)). In a second experiment, we allow the ocean circulation to change according to the transient thermal boundary condition at the sea surface. The latter experiment can be taken as a first order guess for the feedbacks that could be expected when ocean, atmosphere, and marine carbon cycle models were coupled on-line.

In section 2 we describe the mechanism of coupling the atmospheric boundary condition of the OGCM to the atmospheric $pCO_2$. In section 3 we describe and discuss the results of the experiments.

II.1 The model

We use the 11 layer version ATOS1 of the Hamburg LSG general circulation model as described by Maier-Reimer et al., (1993). This version has been defined from a series of experiments with

different boundary conditions to represent the standard LSG control run. The main criterion for the choice was the realistic reproduction of natural radiocarbon in the Atlantic which is a sensitive variable for the conveyor belt structure (Broecker, 1991) of the circulation model. The thermodynamic variables and the horizontal components of velocity are defined at depth levels 25, 75, 150, 250, 450, 700, 1000, 2000, 3000, 4000, and 5000 m. The layer thickness is normally defined by the computation levels: however, at locations where the assumed topography intersects the levels of computation, a modified layer thickness is introduced in order to obtain smoother variations of depth than could otherwise be achieved for the low number of levels.

The carbon cycle part of the model is based on the assumption of perfect stoichiometric relationships (Redfield ratios) of organic material:

$$\delta P : \delta N : \delta C : \delta O_2 = 1 : 16 : 122 : -172$$

Export production is modeled by simple Michaelis-Menten kinetics, using a basic time constant for production in optimal growth conditions which is locally modified by inhibition factors like lack of light, temperature, convective mixing, and windstirring. The basic time constant was tuned to yield realistic gradients of surface phosphate. Phosphate is assumed to be the one limiting nutrient; the Redfield ratio for nitrogen is used only in the determination of alkalinity changes by new production. The model predicts also production of calcareous shells; the overall rain ratio $C_{calc.}:C_{org}$ (=0.17) was tuned to yield a realistic structure of the lysocline. This configuration of the model coupling has already been applied to study ENSO related fluctuations of the marine carbon cycle (Winguth et al., 1994). In the eastern equatorial Pacific, the production is clearly overestimated. It is partly caused by the mismatch between the equatorial upwelling velocity and the assumed restoring time constant and it leads to "nutrient trapping" (Najjar et al., 1992), i.e. the mechanism that the new production in the equatorial upwelling is not horizontally exported like the upwelled water but delivered back into the region of upwelling. This apparent model flaw is manifested in a large pool of POC in the order of 300 Gt that cannot be remineralized due to lack of oxygen.

III. Experiments.

III.1 Biological vs. Solubility pump

Several pumping mechanisms are identified to transport carbon from surface layers to the deep ocean (cf. Broecker and Peng, 1986). The most obvious process is provided by the high solubility of $CO_2$ at low temperatures. Two other processes are caused by the biological formation of soft tissue matter and of calcium carbonate with opposing effects on the partial pressure of $CO_2$. All three have their counterparts provided by low latitude heating, remineralization , and dissolution, respectively. The deep waters of the ocean are cold, but the products of deep remineralization processes increase the local $pCO_2$ substantially above the values of cold surface waters.

In order to elucidate the danger of premature conclusions from variations in the global export production, we discuss firstly three extreme experiments as unrealistic extrapolations of the expected changes: It is expected that the global thermohaline circulation will be reduced. We performed one experiment in which all currents are set to zero and all convection events are ignored. In this scenario, the surface ocean will soon be completely depleted of nutrients and the biological pump is brought to zero. This experiment is an extrapolation of one of the experiments by Heinze et al. (1991) who found that a halving of the circulation rate lowers the atmospheric $pCO_2$ by 26 ppm.

The increase of incoming UV-B radiation to the earths surface by ozone depletion has stimulated a discussion about the impact of photosynthesis inhibition on the global carbon budget (Häder, 1993). We performed an experiment in which the productivity was set to zero. After some centuries, the nutrients and the high $\Sigma CO_2$ values of the deep ocean are homogeneously (besides the effects caused by the temperature dependent solubility of $CO_2$) mixed in the ocean including the surface layer while the biological pump is zero, too.

In a third experiment we enhanced the effectiveness of the biological pump by the assumption of complete sedimentation of the export production. This is equivalent to a removal of all nutrients together with dissolved carbon and alkalinity according

the Redfield ratio. The resulting drop of the atmospheric $pCO_2$ by only 100 ppm is remarkably small, only 5 % of the appr. 4000 Gt of carbon fixation that would correspond to the existing pool of phosphate. This experiment represents a combination of carbon extraction and homogeneization like in exp.2. When applying the 1:5 partitioning recipe between atmosphere and ocean on the result of exp. 2, the apparent contradiction is removed.

The main results are summarized in figs. 2-4, displaying the time series of global export production (Fig.2) and atmospheric $pCO_2$ (Fig.3). All experiments show a strong reduction of biological activity, but the atmospheric response is quite different.

These experiments indicate, however, also a second order direct effect of marine biology on the uptake of $CO_2$. Since the chemical state of surface water is substantially changed, the uptake capacity as defined by the buffer factor is modified. Fig.4 displays the zonally averaged differential buffer factor (Maier-Reimer and Hasselmann, 1987)

$$B_d = dpCO_2/d\Sigma CO_2 \; (\partial pCO_2/\partial \Sigma CO_2)^{-1}$$

for the three experiments and the control run. The higher the buffer factor, the higher is the airborne fraction of a fixed amount of $CO_2$- injection into the atmosphere. Since the buffer factor increases (at a given alkalinity) with increasing $pCO_2$ (Bacastow, 1981), the experiments with a stronger biological pump imply, indeed, a slightly enhanced uptake of $CO_2$ by the ocean.

II.2 $pCO_2$-SST coupling

Mixed boundary conditions for ocean models have been introduced by Bryan (1986) to study multiple equilibria of the thermohaline circulation. In this technique, the freshwater fluxes as diagnosed from the effect of the restoring surface salinity to a given data set (mostly from Levitus, 1982) are used to drive the model together with a restoring of surface temperatures to a temperature data set. The rationale of this procedure is the fact that a salinity anomaly has almost no direct effect on the atmospheric forcing, whereas a SST anomaly should quickly be extinguished by an adjustment of the heat exchange with the atmosphere. With such boundary conditions the models were able to exhibit a pronounced variability, characterized partly by self sustained oscillations

and transitions into different modes of circulation. The results then could be interpreted as the oceanic contribution to climate variability.

One of the most striking features of coupled ocean-atmosphere models is a transient pronounced cooling of several ocean regions, especially at locations where deep water is formed (Cubasch et al., 1992 and Manabe et al., 1991). These exceptions from the global warming trend are also found analyses of historical data sets from the past 50 years (Jones et al., 1986). Mikolajewicz et al. (1990) discussed an experiment in which the Hamburg LSG model was driven under mixed boundary conditions with the mean warming pattern as predicted from five different AGCM's - coupled to mixed layer ocean, only - for the case of a doubling of $CO_2$. They have shown with an ocean only model that such a cooling may be explained by oceanic processes alone. It is due to the effect that a reduction of deepwater formation weakens the deep flow towards the equator and, thus, reduces the supply of warm water towards the pole. A potential feedback of the SST on the forcing atmospheric temperature was ignored in that experiment. This neglect of a feedback, however, has been identified as a major shortcoming of traditional mixed boundary conditions (Mikolajewicz and Maier-Reimer, 1994). It yields a high sensitivity of the conveyor belt type thermohaline circulation with respect to perturbations, whereas the conveyor "off" type circulation appears to be much more stable than the "on" type circulation. It appears to be more natural to drive the model with the diagnosed heat fluxes from a stationary run with additional weak restoring to an atmospheric temperature which represents a restoring to the zonal mean rather than the local value of atmospheric temperature. With such boundary conditions the relation between the different modes of thermohaline overturning becomes more symmetric and the "conveyor on"mode is much more stable against perturbations (cf. also Tziperman et al., 1994, Rahmstorf 1994).

For the present study we use additionally the atmospheric warming pattern from the Cubasch et al. (1992) simulation. The amplitude of the temperature anomaly was computed according:

$$T' = T^A \quad \alpha \log (pCO_2/278ppm)$$

where $\alpha$ denotes the radiative forcing of increasing $CO_2$ and $T^A$ is

the warming pattern averaged over the last decade of a 100 years integration of a coupled ocean- atmosphere model under the assumption of a "business as usual" increase of $CO_2$ (IPCC scenario A). The experiment is, thus, driven by prescribing the emission of $CO_2$, rather than by prescribing the atmospheric $CO_2$ concentration. Consequently, the historical record of atmospheric $CO_2$ is not exactly matched. Fig. 5a shows the structure of the warming pattern, as obtained from a sudden doubling of $CO_2$. Fig. 5b shows the resulting temperature change in the surface layer of the ocean. Since the warming pattern includes atlready he reaction of the ocean (Cubasch et al., 1992), the difference between figs. 5a and 5b are not so pronounced as in Mikolajewicz et al. (1990).

At the end of the integration, the strength of the oceanic conveyor belt circulation is markedly reduced. The most dramatic change lies in the mass of the Arctic icecap.

III.1 Changes of oceanic uptake.

In this section, we discuss the effect of a varying oceanic circulation on the atmospheric $pCO_2$. We performed two experiments with the same scenario of fossil fuel emissions. Starting in 1800we prescribed the emission according Marland ans Boden (1991) until 1990. For the next century we assumed an annual increase of 1% of the emissions, roughly correspondig to the IPCC scenario A. We compare the uptake by the stationary circulation field of the standard run of the model with the uptake in the warming scenario where the circulation model is driven additionally with the forcing of section II.

The changes in the physical circulation are pretty similar to the changes in the fully coupled ocean atmosphere model (Cubasch et al., 1992). In the Atlantic, the maximum overturning rate is reduced from 24 to 20 Sv, while the outflow of NADW across 30 deg S which indicates the strength of the model conveyor belt, is reduced from 20 to 18 Sv, displayed schematically in Fig.6. Since we did not assume a change of the wind field patterns, the surface circulation including Ekman transport induced upwelling is almost unchanged. Fig. 7 displays time series of the most sensitive physical quantities which are the Arctic ice volume ( panel a) and the loss of potential energy by convection at places of unstable stratification (panel b). The latter can be interpreted as a

measure for the driving force of the conveyor belt circulation. It operates primarily outside the ice covered area and is, thus, relatively less reduced (by 13 %) than the sea ice volume which at the end of the experiment has almost completely vanished. The reduction of overturning also leads to a decrease of the global export production (panel c), but only by appr. 7 %.

The effect of the changing circulation on the atmospheric $pCO_2$ are pretty small. Fig.8 shows the predicted time series for the two experiments. In the given scaling of the fig., they are almost indistinguishable. The difference at the end of the next century is 20 ppm or 42 GtC, representing a change of only 5 % of the total increase.

The corresponding difference in the ocean is quite inhomogeneously distributed (Fig.9). Near the surface, the difference is obviously determined by thermodynamic effects. Since the water in the feedback experiment is warmer by 1-2 K, the solubility is reduced and the concentration of DIC in surface waters is lower despite the higher $pCO_2$ in the atmosphere. Since the difference evolves at an atmospheric level above 700 ppm, the thermodynamic effect acts slightly stronger than in the present ocean, where the sensitivity is appr. 10 ppm/K.

In the deep sea, the structure of the difference between the two experiments is less obvious. The most striking feature near 30 deg N has nothing to do with changes of mediterranean outflow; it is connected rather with the intersect of the section with the deep western boundary current which carries in the undisturbed experiment more dissolved $CO_2$. In the deeper part of the Atlantic, the concentration in the feedback experiment is even higher than in the control experiment. This is clearly linked with the changes in phosphate due to the reduction of deep upwelling.

As a consequence of the reduction in the thermohaline circulation, the annual mean export production reduces from 13 to 12 GtC. This reduction is lagged by appr. 20 years by a corresponding reduction of upwelling of remineralized organic matter. The relative strength of the biological pump is, thus, slightly increased with respect to the solubility pump yielding a slightly stronger fixation of carbon in the deep sea.

On the other hand, the weakening of the thermohaline overturning

reduces the speed of sequestering anthropogenic carbon in the deep ocean. This effect dominates the results yielding an effective reduction of the oceanic carbon uptake. A rigorous discrimination between the two opposing effects could be achieved from a simultaneous analysis of changes in temperature, salinity, phosphate, DIC, and alkalinity, but this goes beyond the scope of this paper.

Conclusions:

We have performed several experiments with a general ocean circulation model coupled online with a model of biogeochemical cycling in the ocean. Since we did not allow for an explicit alteration of the biological activity and since the predicted changes in the ocean circulation are not dramatic, the potential feedbacks during the next century are expected to be small. Our findings may be summarized by the following statements:

Although biology does not transport directly anthropogenic $CO_2$ to the deep sea, it causes changes in the inorganic aspects of marine chemistry which have an effect on the gradual uptake of $CO_2$ by the ocean.

We expect the feedbacks to be so small that the currently used modelling strategy of first using a carbon cycle model for the transformation of anthropogenic emissions into $pCO_2$ and subsequently using the output as forcing for a physical climate model, appears to be justified.

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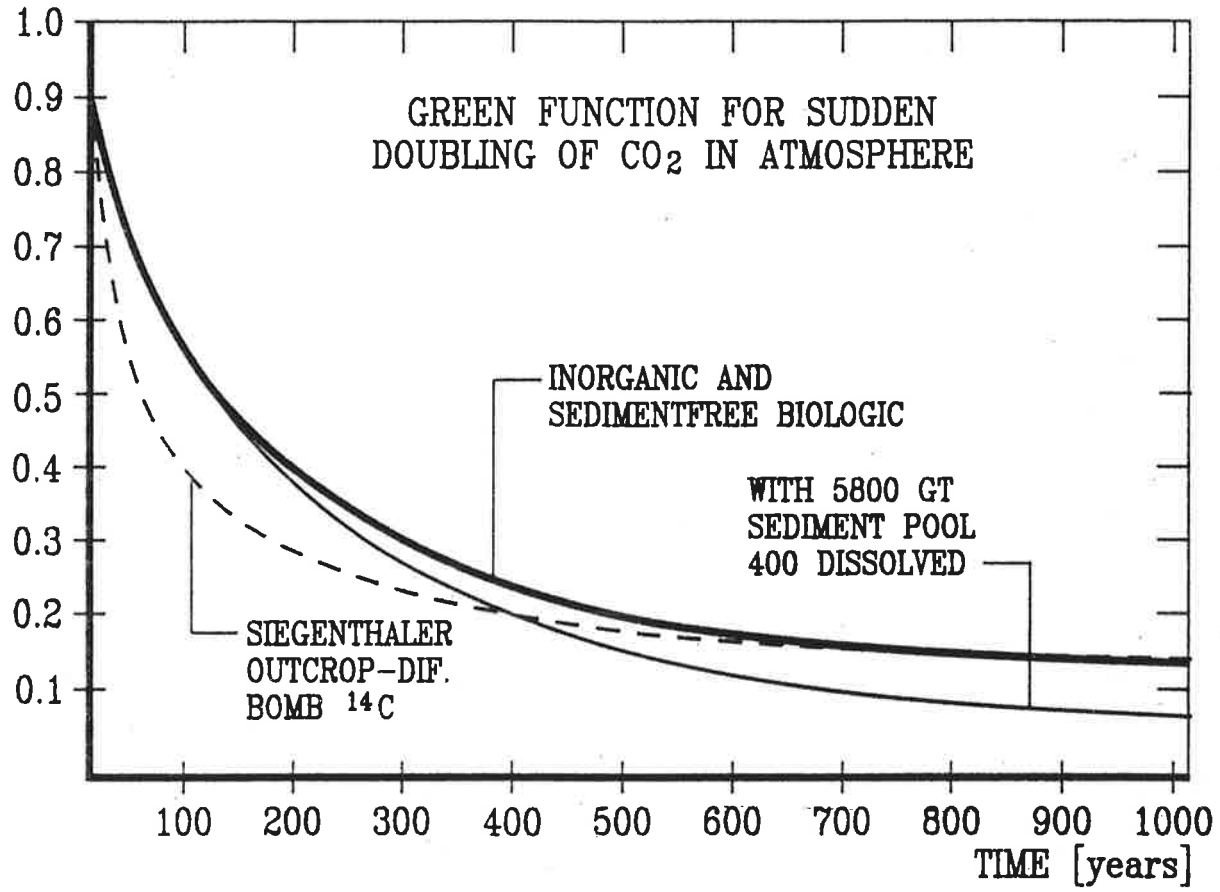


Figure 1: Response of the atmospheric CO2 concentration on a sudden doubling for different versions of the HAMOCC model and for one of Siegenthalers (1983)models ( with the highest uptake rate). Since no explicit fertilization by excess carbon is assumed, the response of the inorganic model and the sedimentfree model are almost identical.

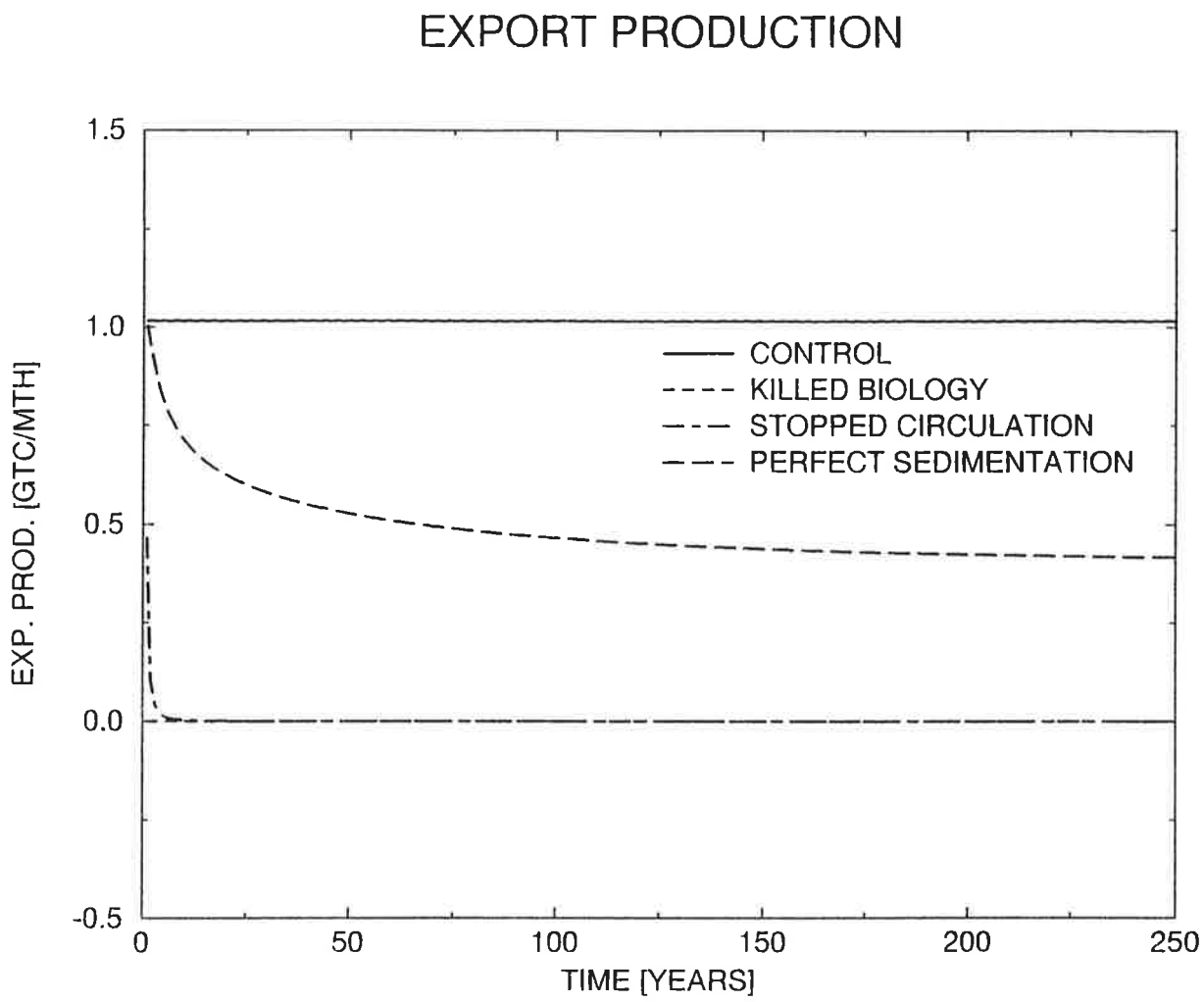


Fig. 2: Timeseries of the global export production for three extreme scenarios of biological activity.

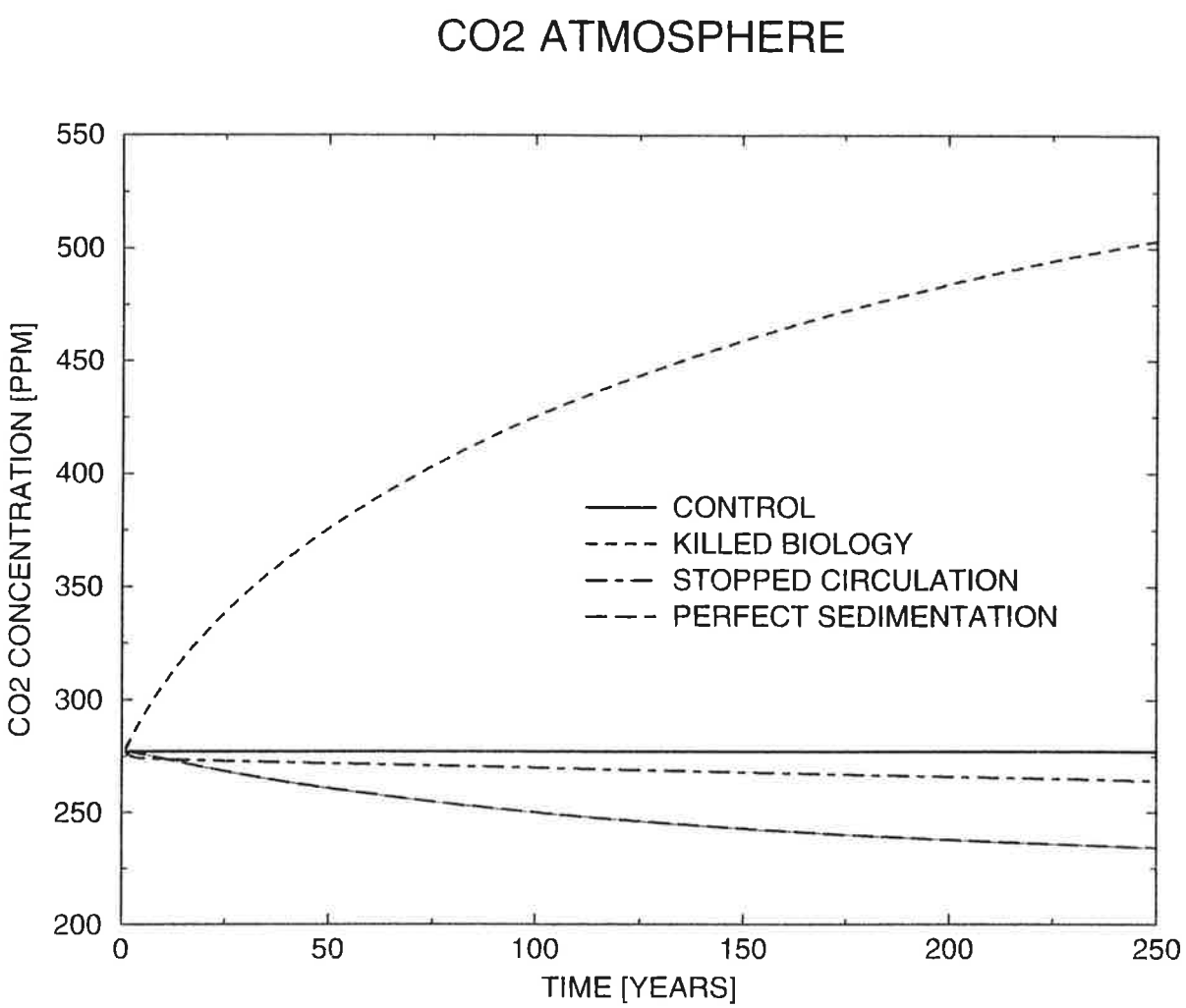


Fig. 3: Timeseries of atmospheric $pCO_2$ for three extreme scenarios of biological production

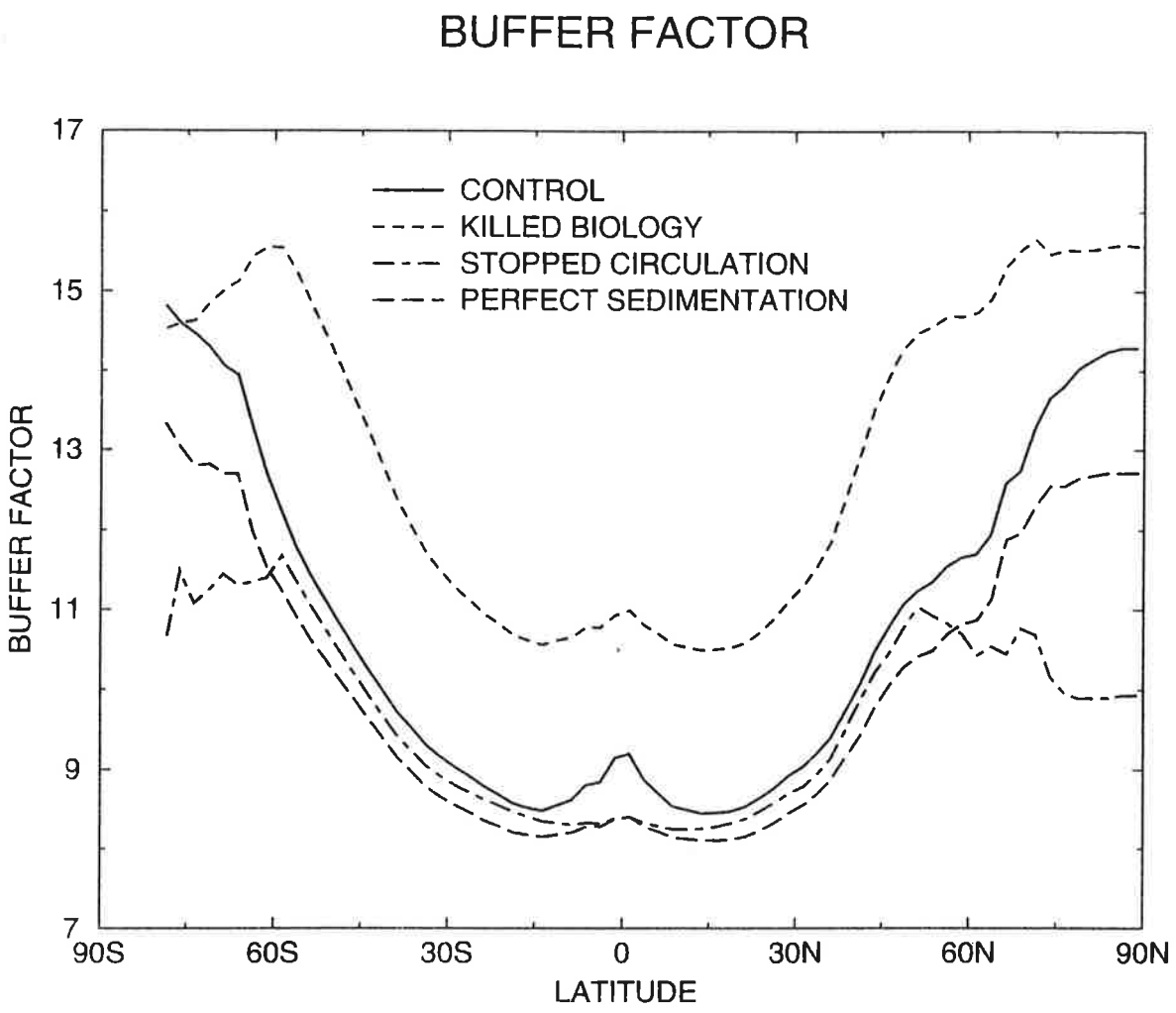


Fig. 4: Zonal average of the buffer factor for three extreme scenarios and the control run..

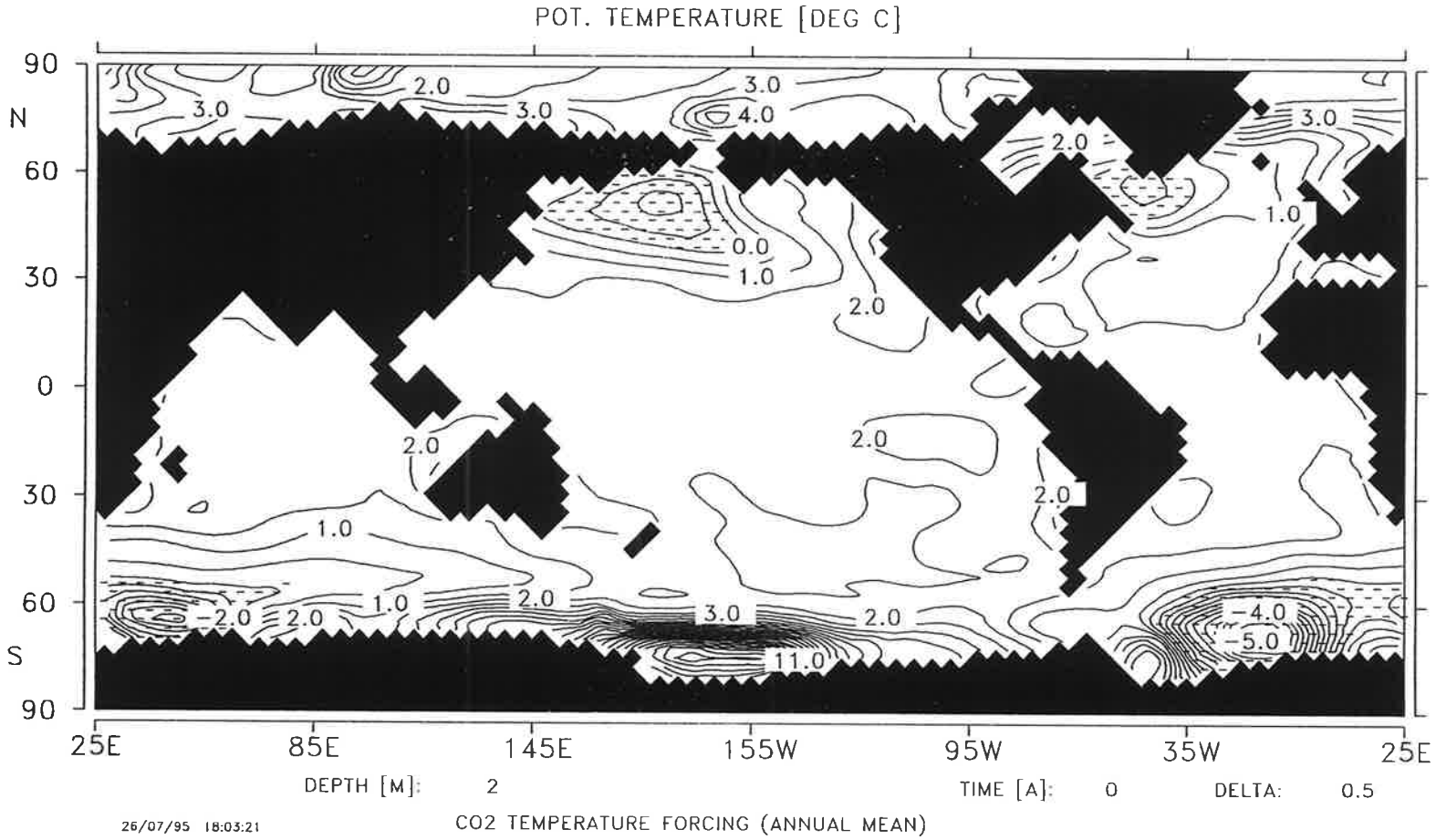


Fig. 5a: Structure of the january atmospheric temperature response as derived from a sudden $CO_2$- doubling in a coupled ocean - atmosphere model (Cubasch et al., 1992).

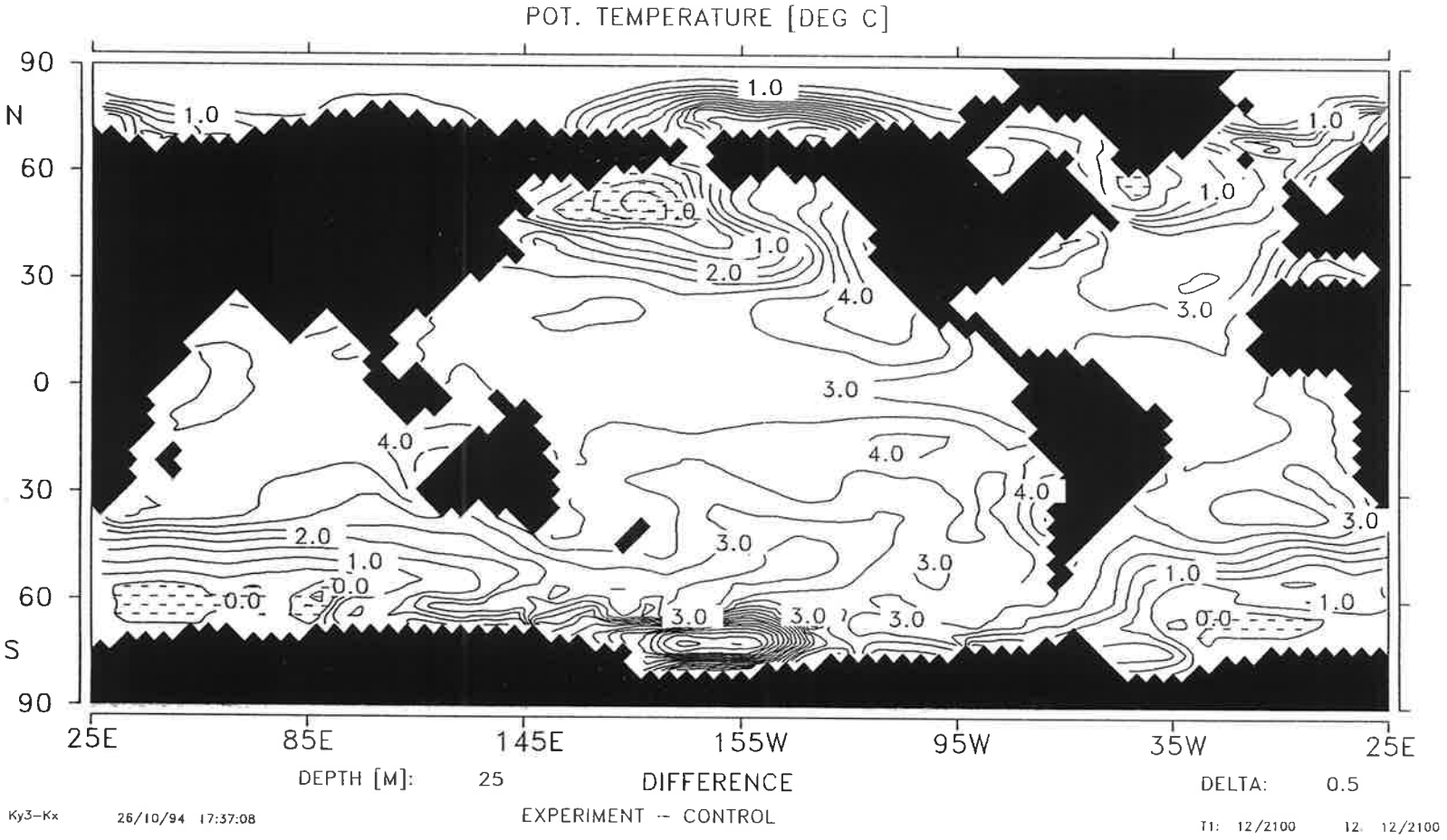


Fig.5b: Change of Sea Surface Temperature at the end of the feedback run.

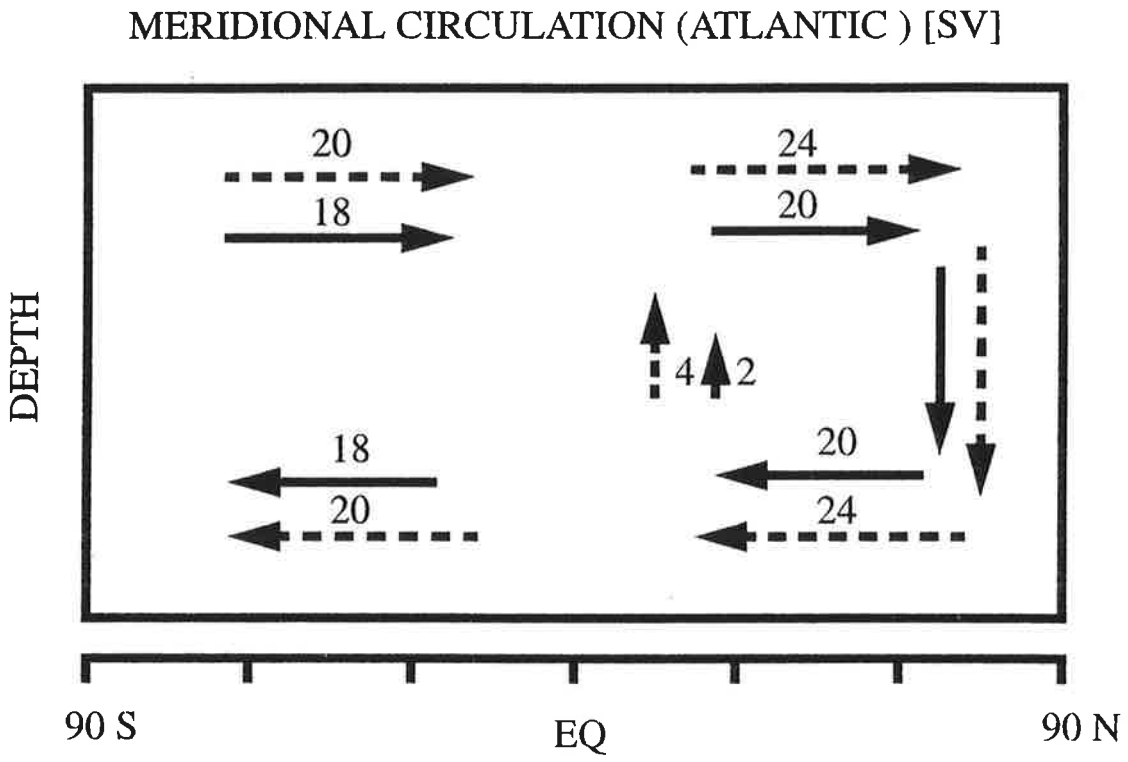


Fig. 6: Change of the Atlantic overturning circulation in the temperature feedback experiment. Dotted arrows for the preindustrial state, solid arrows for 2085.

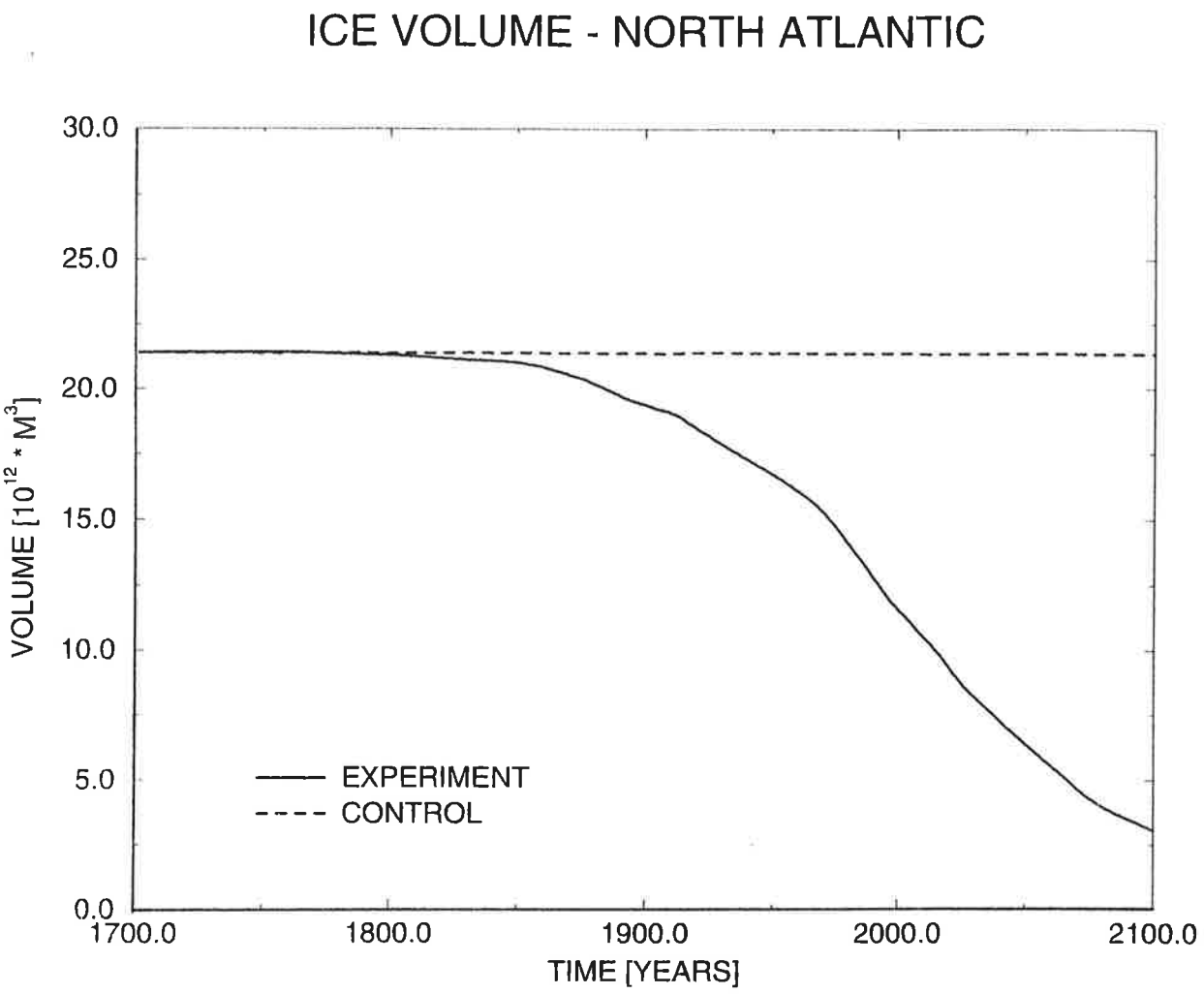


Fig. 7a: Evolution of Arctic Sea ice for the control and the feedback run.

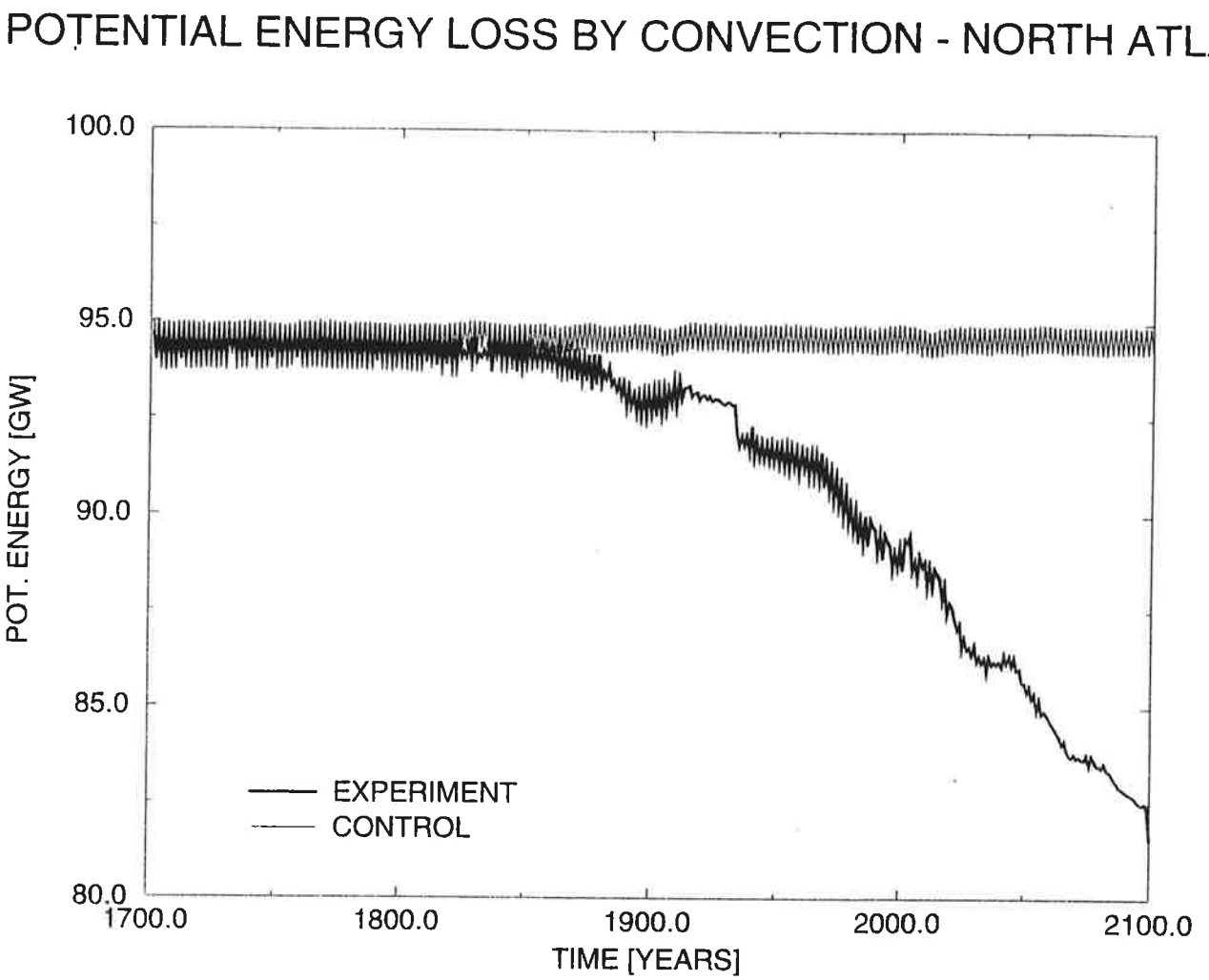


Fig. 7b Convective release of potential energy for the control and the dfeedback run.

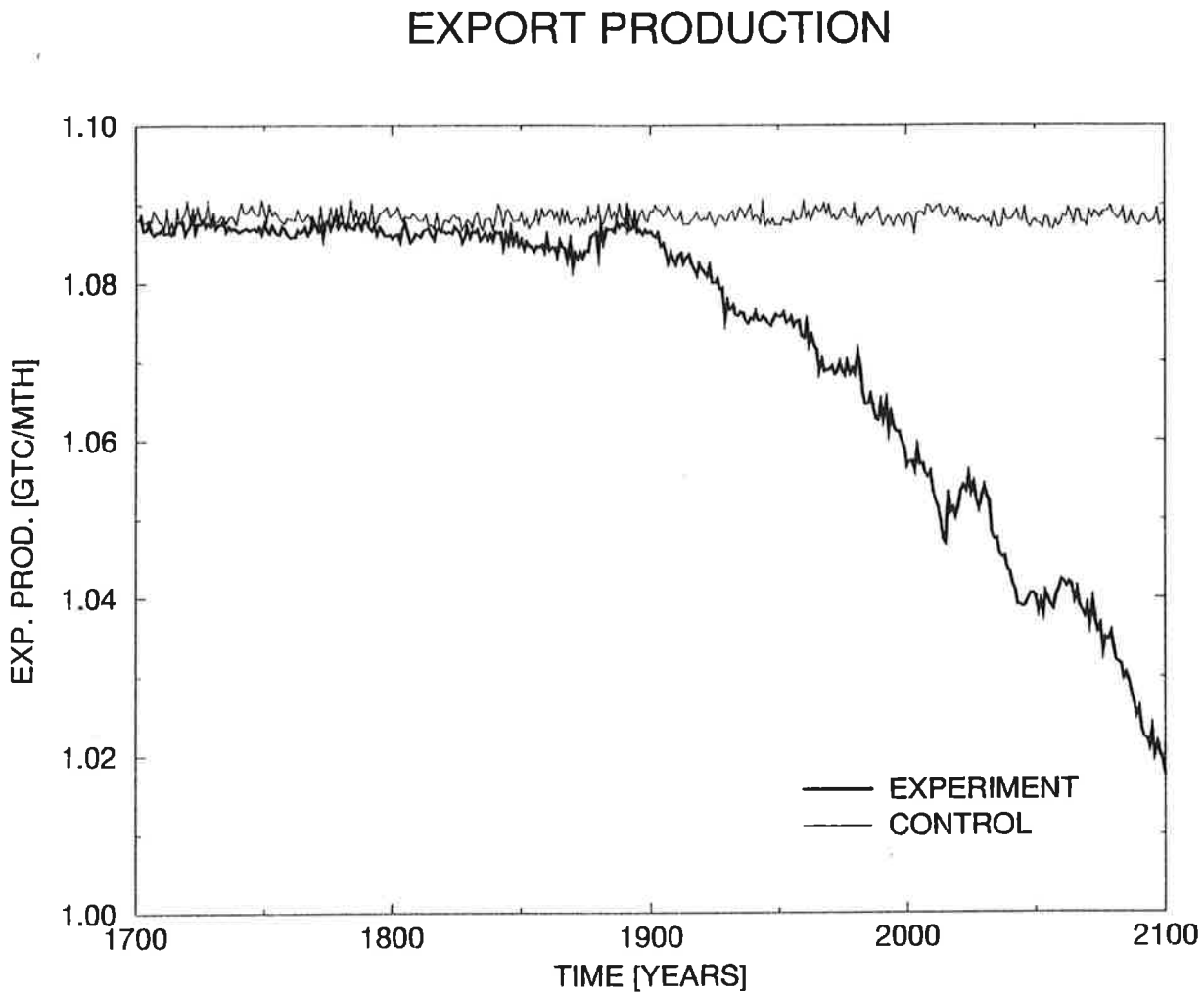


Fig.7c Global export production for the control and the feedback run.

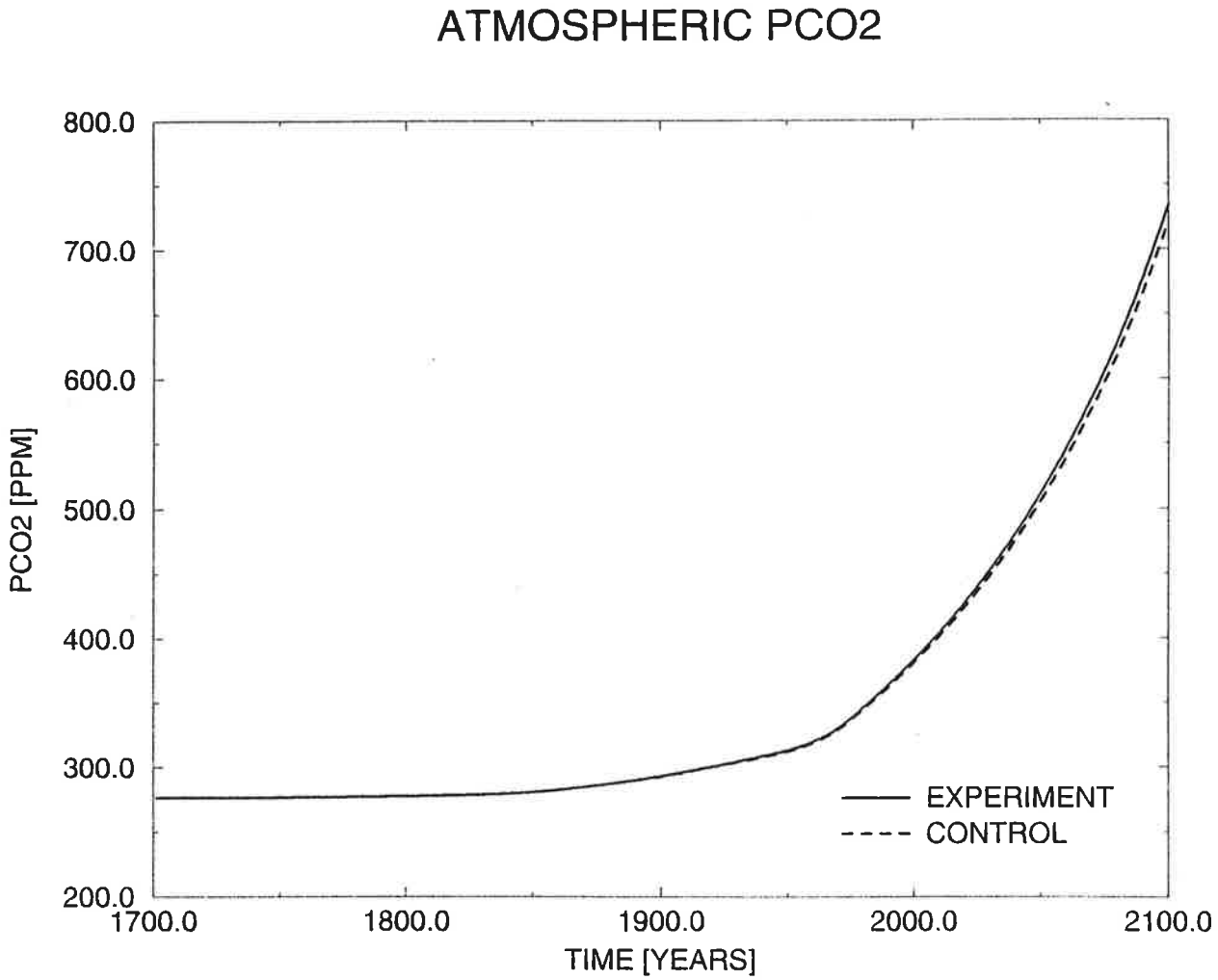


Fig. 8:Evolution of atmospheric $pCO_2$ for the control and the feedback run. The difference at the end of the experiments is 20 ppm.

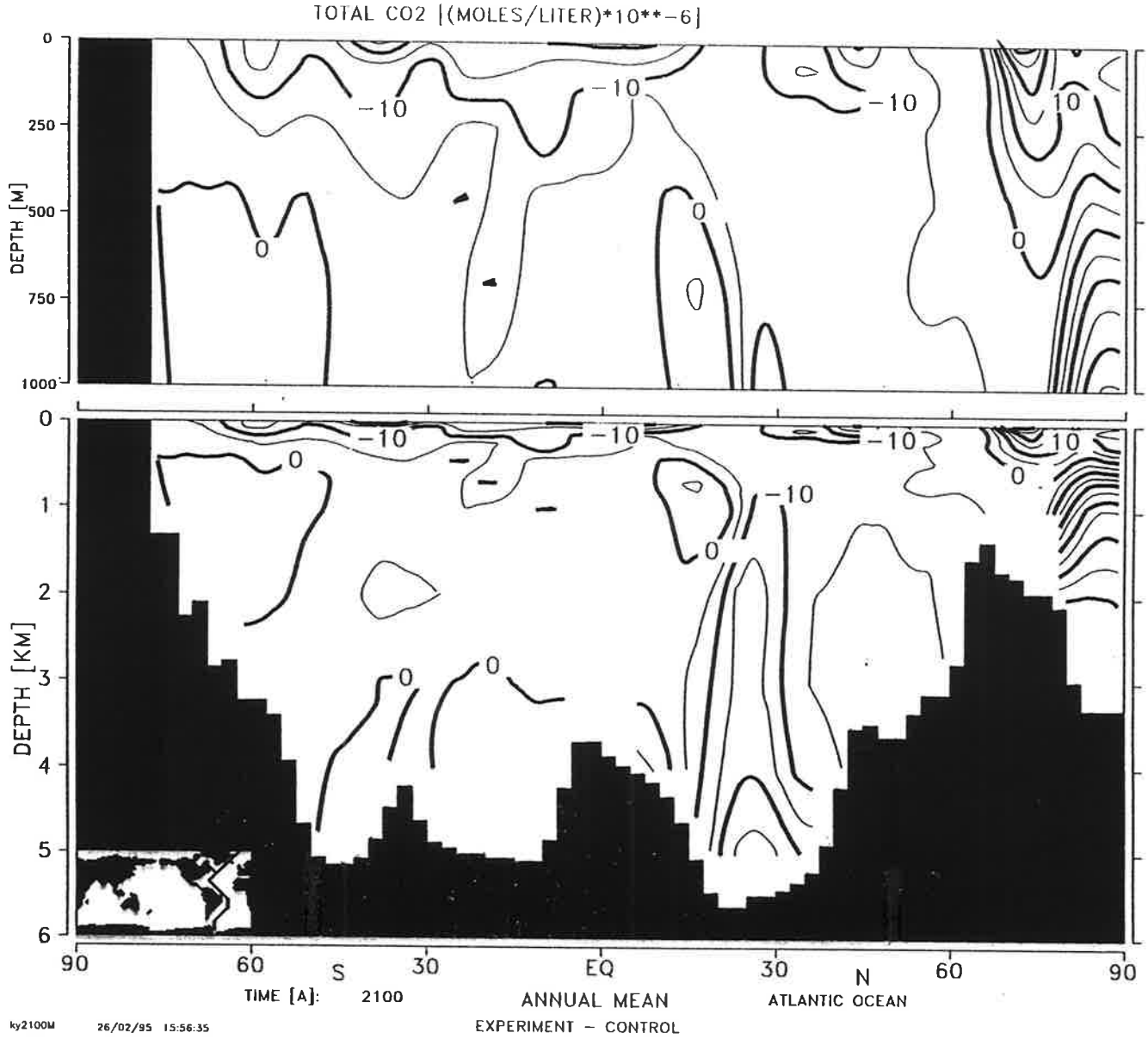


Fig. 9: Difference of $CO_2$ uptake in the Atlantic Ocean.