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TELLUS
ISSN 0280–6509

Climate change feedback on the future oceanic CO_2 uptake

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(Manuscript received 13 August 1998; in final form 15 December 1998)

ABSTRACT

Output from a coupled atmosphere–ocean model forced by the IS92a greenhouse gas scenario was used to investigate the feedback between climate change and the oceanic uptake of CO_2. To improve the climate simulation, we used Gent and co-workers eddy parameterization in the ocean and a prognostic equation for export production from the upper ocean. For the period of 1850 to 2100, the change in the oceanic uptake of CO_2 with climate was separated into 3 feedbacks. (i) Climate change warmed the sea-surface temperature which increased the partial pressure of CO_2 in the surface ocean and reduced the accumulated ocean uptake by 48 Gt C. (ii) Climate change reduced meridional overturning and convective mixing and increased density stratification in high latitudes which slowed the transport of anthropogenic CO_2 into the ocean interior and reduced the cumulative ocean CO_2 uptake by 41 Gt C. (iii) Climate change altered "natural" cycling of carbon in the ocean which increased the cumulative ocean CO_2 uptake by 33 Gt C. The change in natural carbon cycling with climate change was dominated by 2 opposing factors. First, the supply of nutrients to the upper ocean decreased which reduced the export of organic matter (by 15% by year 2100) and produced a net CO_2 flux out of the ocean. However, associated with the reduced nutrient supply was the reduction in the supply of dissolved inorganic carbon to the upper ocean, which produced net CO_2 flux into the ocean. For our model, the latter effect dominated. By the year 2100, the combinations of these 3 climate change feedbacks resulted in a decrease in the cumulative oceanic CO_2 uptake of 56 Gt C or 14% of the 402 Gt C of oceanic CO_2 uptake predicted by a run with no climate change. Our total reduction in oceanic CO_2 uptake with climate change for the 1850 to 2100 period was similar to the 58 Gt C reduction in oceanic CO_2 uptake predicted by Sarmiento and Le Quéré. However, our consistency with this previous estimate is misleading. By including the Gent and co-workers eddy parameterization in the ocean, we reduced the positive feedback between climate change and the oceanic uptake of CO_2 from 169 to 89 Gt C (80 Gt C change). This reduction reflects a decrease in both sea surface warming and anthropogenic forcing feedbacks. By using a prognostic parameterization of export production, we reduced the negative feedback response of the natural carbon cycle to climate change from 111 to 33 Gt C (78 Gt C). These 2 large offsetting changes in the ocean response to climate change produced only a net change of 2 Gt C. This resulted in a net reduction in oceanic uptake of 2 Gt C from the previous study.

1. Introduction

IPCC projections of future atmospheric CO_2 levels assume no feedback between climate change and oceanic uptake of CO_2. The setting of emission targets as part of a global effort to stabilize the atmospheric CO_2 concentration may depend on accounting for the feedbacks between climate change and the oceanic uptake of CO_2. We investigate the potential for climate change to alter the oceanic uptake of CO_2 by using output from a coupled atmosphere–ocean model.

2 previous studies looked at the potential feed-

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back between climate change and ocean uptake of CO_2 (Maier-Reimer et al., 1996; Sarmiento and Le Quéré, 1996) but reached different conclusions. Maier-Reimer et al. (1996) used an ocean only model with a carbon cycle model to investigate this feedback. Their ocean model was forced by the surface temperature generated by the coupled atmosphere–ocean model. By 2100, they predicted climate change would reduce the cumulative oceanic CO_2 uptake by 26 Gt C. By forcing the ocean with only surface temperature, they potentially underestimated the effect of climate change because salinity changes significantly influence the ocean behavior at high latitudes (Manabe and Stouffer, 1993; Hirst et al., 1996; Kattenberg et al., 1996). However, as stated by Maier-Reimer et al. (1996), their simulated ocean pattern with only temperature forcing is consistent with climate model results (Cubasch et al., 1998; Mikolajewicz et al., 1994). The no change in precipitation in the southern ocean reflects a different response in their climate model to greenhouse warming. Sarmiento and Le Quéré (1996) used a coupled atmosphere–ocean model with an ocean carbon cycle to investigate the climate change feedback on oceanic CO_2 uptake. With a solubility model of the ocean carbon cycle, climate change reduced the cumulative oceanic CO_2 uptake by 169 Gt C by 2100. With a biological model, climate change reduced the cumulative oceanic uptake of CO_2 by 58 Gt C by 2100. For their biological model, export production of organic matter was determined from the control run by relaxing the modelled phosphate to the present-day phosphate field; this export production was applied to the climate change run. This approach neglects possible changes in export production with climate change.

Another important factor in the feedback between climate change and oceanic CO_2 uptake is the change in the rate of ventilation of the southern ocean (Sarmiento and Le Quéré, 1996). Many models to date have displayed extensive and very deep convective mixing at high southern latitudes (Sarmiento and Le Quéré, 1996). However, such convection is not observed. Simulations of CFCs (England, 1995; Robitaille and Weaver, 1995) and natural ^{14}C (personal correspondence with Matear, 1998) suggest that models with deep convective mixing over-ventilate the southern ocean. The inclusion of the Gent and McWilliams (GM) eddy advection parameterization (Gent and McWilliams, 1990) in ocean models reduces convective mixing and increases stratification in the southern ocean, thus improving the CFCs (Robitaille and Weaver, 1995; England and Hirst, 1995) and natural ^{14}C simulations ^{14}C (personal correspondence with Matear, 1998).

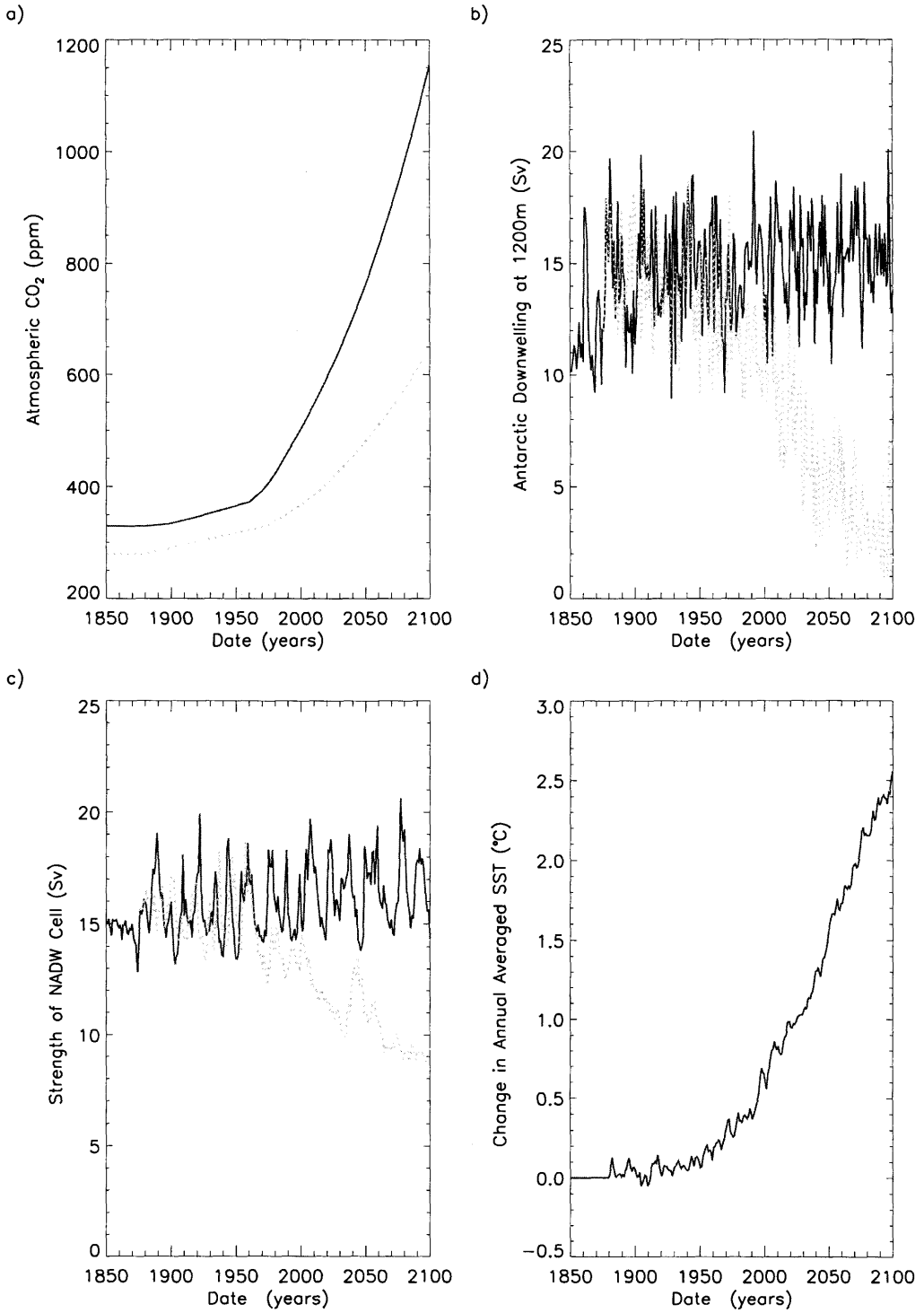
Our work extends the two previous studies (Maier-Reimer et al., 1996; Sarmiento and Le Quéré, 1996) by significantly improving the ocean dynamics in the southern ocean and improving the feedbacks between the biological export production and climate change. A more realistic level of southern ocean convection may alter the feedback between climate change and oceanic CO_2 uptake predicted by previous studies (Maier-Reimer et al., 1996; Sarmiento and Le Quéré, 1996). The improvement of the ocean dynamics in the southern ocean is crucial because the Sarmiento and Le Quéré (1996) model results showed that this region experienced the greatest change in oceanic CO_2 uptake with climate change. Furthermore, we explore the processes controlling the oceanic CO_2 uptake response to climate change.

2. Models

2.1. Climate model

We used the CSIRO climate model (Gordon and O'Farrell, 1997) which is a general circulation model containing atmospheric, oceanic, sea-ice and biospheric sub-models but with the GM scheme (as implemented in Hirst and McDougall (1996)) and horizontal diffusivity set to zero. The spin-up of the separate sub-models is described in Gordon and O'Farrell (1997). The physical response to increased greenhouse gases is discussed in Hirst et al. (1996) and Hirst (1998). The present study utilized 2 experiments with the climate model: (i) a control experiment with a constant atmospheric level of CO_2; (ii) a climate change experiment where the equivalent CO_2 followed the IS92a radiative forcing scenario (Houghton et al., 1995) from 1850 until 2100 (Fig. 1a).

In simulations of greenhouse gas forcing, increased precipitation at high latitudes causes a decline in surface salinity which stabilizes the upper water column (Manabe et al., 1991; Manabe


a)
Atmospheric CO_2 (ppm)
1200
1000
800
600
400
200
1850 1900 1950 2000 2050 2100
Date (years)
b)
Antarctic Downwelling at 1200m (Sv)
25
20
15
10
5
0
1850 1900 1950 2000 2050 2100
Date (years)
c)
Strength of NADW Cell (Sv)
25
20
15
10
5
0
1850 1900 1950 2000 2050 2100
Date (years)
d)
Change in Annual Averaged SST (°C)
3.0
2.5
2.0
1.5
1.0
0.5
0.0
−0.5
1850 1900 1950 2000 2050 2100
Date (years)

and Stouffer, 1993; Murphy and Mitchell, 1995). Further stabilization was produced by increased sea surface temperature and reduced sea ice formation. This stabilization reduces convective mixing and weakens thermohaline overturning, both of which are important for high-latitude ocean ventilation. In the present experiment, Hirst (1998) showed that such a stabilizing response occurs very strongly in the southern ocean. Associated with this stabilization is a reduction in the Antarctic overturning rate. In the control experiment, strong downwelling occurs near Antarctica. The downwelling through 1250 m in the Antarctic region in the control experiment is $13–16 \times 10^6$ $m^3 s^{-1}$ but collapses in the climate change experiment (Fig. 1b). In the north Atlantic, climate change weakens the north Atlantic deep water (NADW) overturning (Fig. 1c). However, with climate change, the reduction in deep water formation is much more dramatic in the southern ocean than in the north Atlantic, an effect not so evident in previous studies (Maier-Reimer et al., 1996; Sarmiento and Le Quéré, 1996). With climate change, the sea surface temperature warms, which reduces the solubility of CO_2 in the surface ocean (Fig. 1d).

2.2. *Ocean carbon cycle model*

A prognostic ocean carbon cycle model was implemented in a global ocean model to determine distributions of dissolved inorganic carbon (DIC), alkalinity, phosphate and dissolved oxygen (RJM4). In the model, the export of organic matter from the euphotic zone (upper 50 m of the ocean) was determined using light, temperature, mixed-layer depth and phosphate-dependent equations. The temperature, light and mixed-layer depth dependency followed Kurz and Maier-Reimer (1993) and phosphate dependency was $cP/(P + k)$, where P is the phosphate concentration, k was 0.1 μM and c a constant scaling factor of export production, set to 1/4 month. The temperature, light, mixed-layer depth and phosphate fields were time-dependent and the constants k and c were determined by comparing the model nutrient fields to observations. The composition of the exported organic matter assumed the Redfield ratio $P:N:C:O_2$ of 1:16:106:138 (Redfield et al., 1963). Following Yamanaka and Tajika (1996), the $CaCO_3$ export was a constant 8% fraction of the organic matter export. The exported material from the euphotic zone was instantaneously remineralized below the euphotic zone. The remineralization profile of organic matter was $(z/100)^{-a}$, where z is depth (m) and $a = 0.9$, and the remineralization profile of $CaCO_3$ was $e^{(-z/3500)}$. The air–sea flux of CO_2 and O_2 were calculated using a wind-speed dependent gas exchange coefficient with chemical enhancement (Wanninkhof, 1992) for CO_2 and O_2 multiplied by air–sea CO_2 and O_2 differences, respectively. The partial pressure of CO_2 in the atmosphere (pCO_2) was specified (Fig. 1), while the surface ocean pCO_2 was determined using the full carbon chemistry equations which required modelled temperature, salinity, alkalinity and dissolved inorganic carbon from the surface ocean. The partial pressure of O_2 in the ocean (pO_2^{ocean}) was determined using modeled surface ocean temperatures and salinities, and the solubility of dissolved oxygen in seawater from Weiss (1970).

The steady-state tracer distributions in the ocean were obtained by running an ocean-only model to steady-state with initial average phosphate and alkalinity set to 2.0 μM and 2431 μM eq, and the atmospheric CO_2 and O_2 set to 280 ppm and 20.85%. The ocean model reproduced the large-scale features of the observed tracer distributions of phosphate, oxygen, DIC and alkalinity.

3. Oceanic CO_2 uptake

Our ocean carbon cycle runs were performed using archived model data following a technique

Fig. 1. (a) The greenhouse gas forcing used for the climate change run given as equivalent CO_2 levels (solid) and the atmospheric CO_2 concentrations used by the ocean carbon cycle model (dotted). The equivalent CO_2 levels include the effects of other greenhouse gases. (b) The transport of Antarctic bottom water (AABW) in Sv ($106\ km^3\ s^{-1}$) through 1250 m from the control run with no anthropogenic greenhouse gas forcing (solid) and with climate change with greenhouse gas forcing (dotted). (c) The strength of the north Atlantic deep water (NADW) for the control run (solid) and the climate change run (dotted). (d) Change in the global averaged sea surface temperature with climate change.

described by Aumont et al. (1998). Archived monthly mean data from a climate model experiment for the years 1850 to 2100 provided the ocean circulation used by the ocean carbon cycle model. The archived data used included the 3-dimensional fields of ocean temperatures, salinities and currents, and the surface fields of windspeed and sea ice cover. With archived data from the control and climate change experiments, we performed a suite of runs to elucidate the feedback between climate change and oceanic CO_2 uptake (Table 1). The runs were chosen to separate the change in CO_2 uptake by the ocean due to changing sea surface temperature from that associated with changes in ocean circulation. In our runs, we considered 2 different scenarios for the evolution of atmospheric CO_2: (i) increasing according to IS92a and (ii) held constant at the pre-industrial level of 280 ppm (Table 1). For runs 1–3, the atmospheric CO_2 levels increased according to IS92a (Fig. 1a). Run 1 used the control experiment's archived data while run 2 used archived data from the climate change experiment. The difference between runs 1 and 2 gave the change in the ocean uptake of CO_2 with climate change. Climate change reduced CO_2 uptake (Fig. 2a, b) and by 2100, the annual oceanic uptake of CO_2 decreased by 20%, and the cumulative ocean CO_2 uptake between 1850 and 2100 decreased by 56 Gt C (Table 2). The decrease in CO_2 uptake is the result of an increase in the surface ocean CO_2 partial pressure (pCO_2) caused by elevated sea surface temperatures and the modification of the anthropogenic CO_2 uptake and ocean carbon cycling caused by changes in oceanic circulation. Run 3 was used to separate the sea surface warming effect from the ocean circulation effect; it was identical to run 2 except that the control experiment sea surface temperatures were used in the pCO_2 calculation. The differences among runs 1–3 showed that the sea surface warming dominated reduction in CO_2 uptake until 2050, but by 2070, the magnitude of the ocean circulation effect was similar to the surface warming effect (Fig. 2a). For the period from 1850 to 2100, changes in CO_2 uptake caused by sea surface warming (−44 Gt C by 2100) exceeded changes due to ocean circulation (−12 Gt C by 2100) (Table 2).

To separate the ocean circulation effect into the impact on the anthropogenic CO_2 uptake and the impact on ocean carbon cycling, we repeated the first 3 runs with the atmospheric CO_2 held at the pre-industrial level (Table 1). The response of

Table 1. *Configuration of the model runs performed for this study*

Run no.	Ocean circulation	Sea-surface temperatures	Atmospheric CO_2	Cumulative CO_2 uptake by 2100 (Gt C)
1	control	control	anthropogenic (IS92a)	376
2	climate change	climate change	anthropogenic (IS92a)	320
3	climate change	fixed	anthropogenic (IS92a)	364
4	control	control	pre-industrial	−29
5	climate change	climate change	pre-industrial	−44
6	climate change	fixed	pre-industrial	4

Control denotes that the fields are from the control experiment (no greenhouse gas forcing) and climate change denotes that the fields are from the climate change experiment (IS92a equivalent greenhouse gas forcing). For fixed sea surface temperatures, we neglected changes in sea-surface temperatures associated with climate change in the calculation of surface ocean pCO_2 and used the monthly-averaged sea surface temperature fields from the 1850–1880 period.

Fig. 2. (a) Air–sea exchange of CO_2 for model 1 (solid), model 2 (dotted) and run 3 (dashed). (b) Changes in the cumulative oceanic CO_2 uptake with climate change for IS92a atmospheric CO_2 levels (solid, run 2–run 1) and for atmospheric CO_2 levels maintained at the pre-industrial level (dashed, run 5–run 4). (c) Climate change feedbacks on the oceanic CO_2 uptake due to changes in SST (solid, run 5–run 6), changes in ocean carbon cycling (dotted, run 6–run 4) and changes in the oceanic uptake of anthropogenic CO_2 (dashed). The sum of the sum of the solid and dotted curves equals the dashed curve in Fig. 2b, while the dashed curve is the difference between the solid and dashed curves in Fig. 2b. (d) Export production of organic matter of run 1 (solid) and run 2 (dashed).

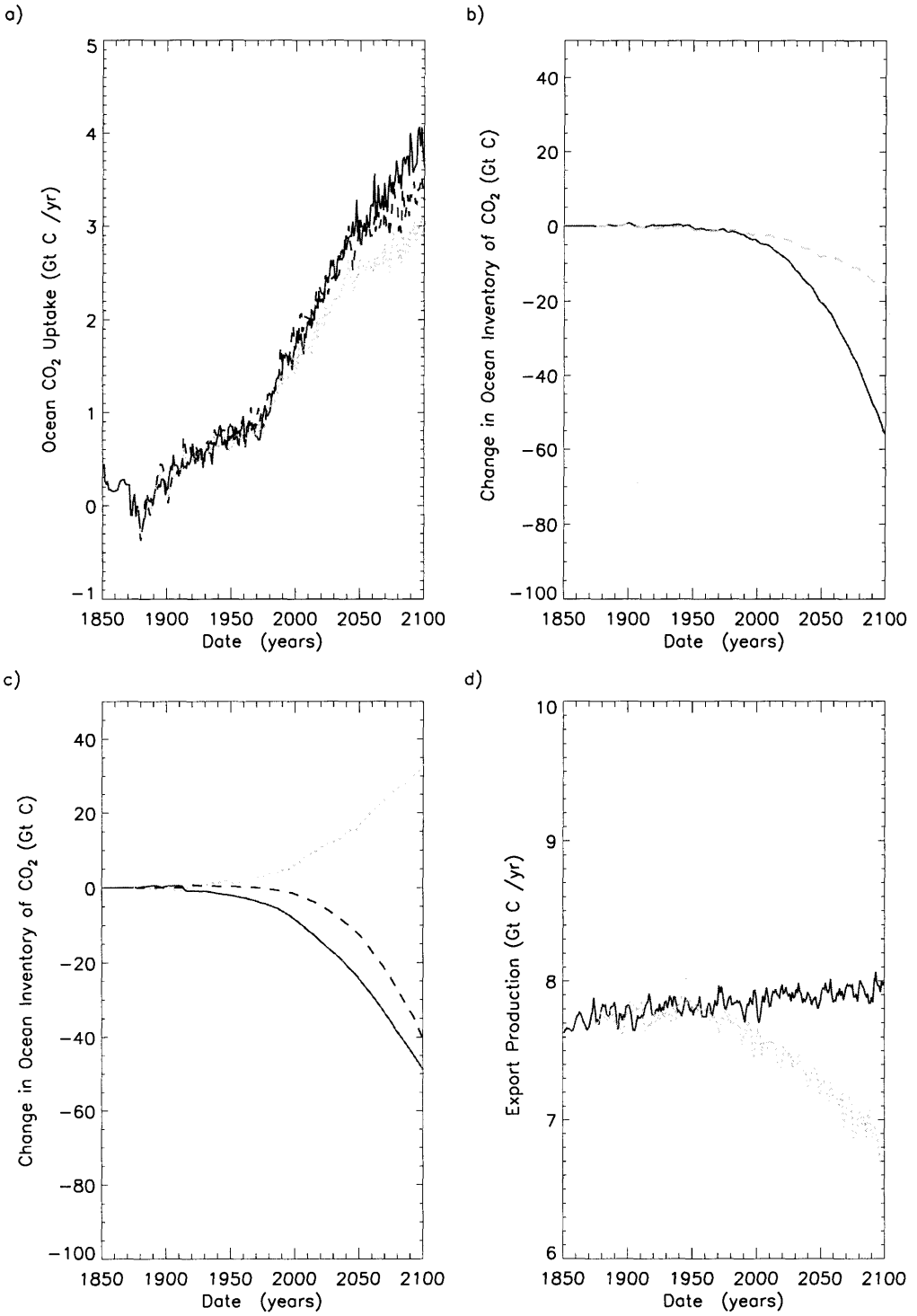

a)
Ocean CO_2 Uptake (Gt C /yr)
5
4
3
2
1
0
−1
1850 1900 1950 2000 2050 2100
Date (years)
b)
Change in Ocean Inventory of CO_2 (Gt C)
40
20
0
−20
−40
−60
−80
−100
1850 1900 1950 2000 2050 2100
Date (years)
c)
Change in Ocean Inventory of CO_2 (Gt C)
40
20
0
−20
−40
−60
−80
−100
1850 1900 1950 2000 2050 2100
Date (years)
d)
Export Production (Gt C /yr)
10
9
8
7
6
1850 1900 1950 2000 2050 2100
Date (years)

Table 2. *Changes in the oceanic CO_2 uptake with climate change due to changes in ocean circulation and changes in sea surface temperatures under anthropogenic and pre-industrial levels of atmospheric CO_2; the difference is interpreted as the change in the uptake of anthropogenic CO_2 with climate change*

	Change in the oceanic CO_2 uptake with climate change (Gt C)		
Atmospheric CO_2 forcing	ocean circulation	sea surface temperature	total
anthropogenic (IS92a)	−12 (runs 3–1)	−44 (runs 2–3)	−56 (runs 2–1)
pre-industrial	+33 (runs 6–4)	−48 (runs 5–6)	−15 (runs 5–4)
difference	−45	4	−41

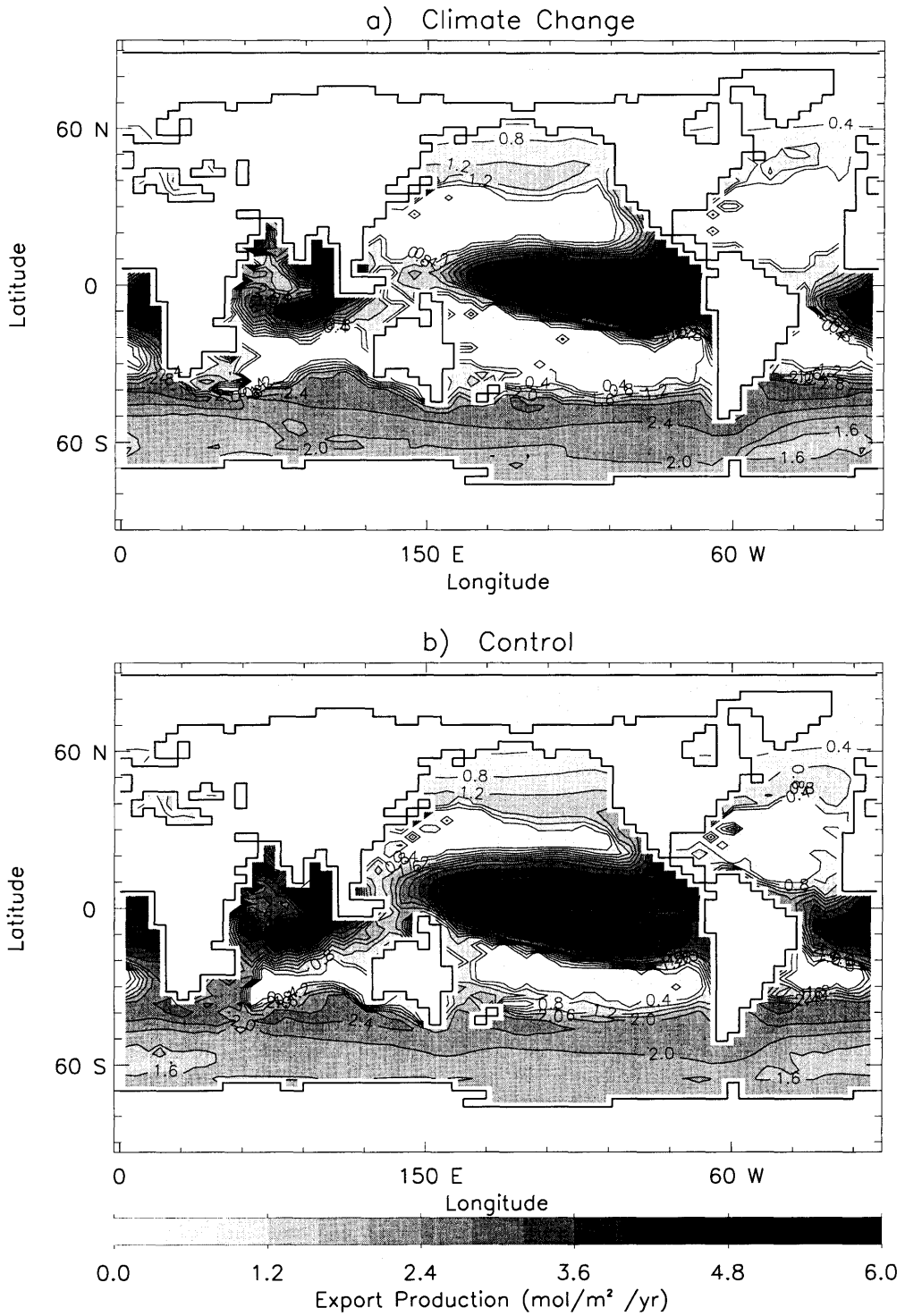
the ocean carbon cycle to changes in circulation was given by the difference between runs 6 and 4; we refer to this as the natural carbon cycle feedback (Table 2, Fig. 2c). The response of anthropogenic CO_2 uptake to circulation changes is given by the difference in the ocean circulation effect under 2 atmospheric CO_2 scenarios (Table 2), and we refer to this as the anthropogenic CO_2 forcing feedback. Due to the non-linearity of ocean carbonate chemistry, our separation of the ocean response into these 3 terms was only an approximation. However, the difference in the sea surface warming effect for the 2 different atmospheric CO_2 scenarios (Table 2) gave an uncertainty of 4 Gt C for the 1850–2100 cumulative carbon uptake. The similarity in the sea-surface warming effect for the 2 atmospheric CO_2 scenarios supports our approximate separation of the ocean uptake responses into sea surface warming, anthropogenic CO_2 forcing and natural carbon cycle feedbacks.

With the control climate (run 4), the ocean experienced outgassing of CO_2 because of the drift in the control experiment. We assume that the same drift was also present in the climate change experiment; the difference between a control run and a climate change run therefore removed the effect of the drift.

With atmospheric CO_2 set at pre-industrial levels, the modifications in ocean circulation with (Broecker, 1998) climate change altered carbon cycling in the ocean by changing the supply of nutrients, DIC and alkalinity to the upper ocean. These changes can produce a net CO_2 exchange between the ocean and atmosphere and alter biological export of organic and inorganic matter from the upper ocean. Climate change reduced the vertical supply of nutrients, DIC and alkalinity to the upper ocean. The reduction in the supply of nutrients and DIC to the upper ocean produced opposing responses on the air–sea exchange of CO_2. The decrease in nutrient supply reduced export production (Fig. 2d), increased the surface ocean pCO_2 and produced a flux of CO_2 out of the ocean. By 2100, climate change runs experienced a 15% reduction in the export production of organic matter from the upper ocean and an expansion of the oligotrophic regions (Fig. 3). The reduced supply of DIC to the upper ocean produced a net uptake of CO_2 by the ocean. Our results show that the reduced supply of DIC dominated the response of the ocean carbon cycle to climate change and produced a natural carbon cycle response that increased the oceanic uptake of CO_2 with climate change (33 Gt C, Table 2).

By comparing the results from runs 1–3 and runs 4–6, we quantify the rôle that rising atmospheric CO_2 levels (anthropogenic CO_2 forcing effect) had on the feedback between climate change and oceanic CO_2 uptake (Table 2). With climate change, the anthropogenic CO_2 forcing decreased the cumulative oceanic CO_2 uptake (−41 Gt C) because the reduction in meridional overturning

Fig. 3. Annual mean export production for 2075 to 2100 for (a) climate change experiment (run 2) and (b) control climate experiment (run 1).


a) Climate Change
60 N
0
60 S
Latitude
0.4
0.8
1.2
1.6
2.0
2.4
0
150 E
60 W
Longitude
b) Control
60 N
0
60 S
Latitude
0.4
0.8
1.2
1.6
2.0
2.4
0
150 E
60 W
Longitude
0.0
1.2
2.4
3.6
4.8
6.0
Export Production (mol/m² /yr)

and convective mixing and increased density stratification in the high latitudes slowed the transfer of anthropogenic CO_2 from the surface into the ocean interior. By 2100, anthropogenic CO_2 forcing was the feedback that was increasing most rapidly (Fig. 2c).

4. Discussion

The IS92a atmospheric CO_2 levels used in our study were similar to the atmospheric CO_2 levels used by Maier-Reimer et al. (1996) and the first 100 years of the 4X CO_2 experiment of Sarmiento and Le Quéré (1996). We will therefore compare our ocean CO_2 uptake response to climate change, with these earlier studies.

By 2100, our reduction in cumulative oceanic CO_2 uptake with climate change was more than double the 26 Gt C predicted by Maier-Reimer et al. (1996). Our climate model predicted 0.5°C less sea surface warming by 2100 than Maier-Reimer et al. (1996), which implies that the sea surface warming feedback cannot explain our increased response of the ocean CO_2 uptake to climate change. The changes in ocean circulation with climate change in Maier-Reimer et al. (1996) may differ significantly from our study. They only needed changes in sea surface temperatures to simulate the effect of climate change on ocean, while in our climate model, the changes in surface freshwater fluxes were of equal importance (Hirst et al., 1996). Our results suggest that the ocean circulation change effect on ocean uptake was small (−12 Gt C) and appear unable to explain the large difference between our value and the Maier-Reimer et al. (1996) study. However, the ocean circulation effect was the combination of 2 opposing effects, the anthropogenic CO_2 forcing feedback (−41 Gt C) and the natural carbon cycle feedback (+33 Gt C), and different changes in the ocean circulation may significantly alter the balance between these 2 feedbacks. We expect the natural carbon cycle feedback between the 2 studies to be significantly different, because by 2100, our predicted reduction of 15% in export production was double the value predicted by Maier-Reimer et al. (1996).

In the Sarmiento and Le Quéré (1996) study, by the end of the first 100 years of the 4X CO_2 experiment, climate change had reduced the oceanic CO_2 uptake by 58 Gt C for their biological carbon cycle model. This result can be compared to our cumulative reduction in ocean uptake with a climate change of 56 Gt C by 2100. Given the differences in ocean dynamics in the southern ocean and differences in the formulation of export production between the 2 studies, it was surprising that this difference was only 2 Gt C. We explored the differences in the 2 studies by utilizing the response of a solubility only ocean carbon cycle model to climate change.

With the separation of our ocean response into the 3 feedbacks, we can compute an upper bound estimate for the response of a solubility-only ocean carbon cycle model to climate change. For our 3 feedbacks, only the natural carbon cycle feedback depended on the formulation of the ocean carbon cycle model. For a solubility-only carbon cycle model, the natural carbon cycle feedback would be approximately zero because there is no biology and the vertical gradients of DIC and alkalinity in the upper ocean are weak. Therefore, the response of a solubility-only carbon cycle model to climate change would be the sum of the sea surface warming and anthropogenic CO_2 forcing feedbacks (sum of solid and dashed curves in Fig. 2c). From our runs, a solubility-only carbon cycle model would experience a 89 Gt C reduction in the cumulative oceanic CO_2 uptake with climate change by 2100. Our solubility model feedback was significantly less than Sarmiento and Le Quéré (1996) value of 169 Gt C over approximately the same period. The 80 Gt C difference reflects a change in the magnitude of the sea surface warming and anthropogenic CO_2 forcing feedbacks of these two studies. Sarmiento and Le Quéré (1996) sea surface warming feedback was 52 Gt C which was only 4 Gt C greater than the 48 Gt C estimated from our runs. Therefore, the remaining 76 Gt C difference in the solubility models was attributed to differences in the anthropogenic CO_2 forcing feedback. A recent study by Sarmiento et al. (1998) also produced a much greater anthropogenic CO_2 forcing feedback (−67 Gt C by 2065) than calculated in our study (−22 Gt C by 2065). Our climate reduced the anthropogenic CO_2 forcing feedback by nearly 1/3 from previous studies. Our reduction in the anthropogenic CO_2 feedback was achieved by dramatic change in the density stratification in the high latitudes of the control run through the inclusion of the Gent et al. (1995)

eddy parameterization. The inclusion of this eddy parameterization reduced the unrealistic deep convective mixing that occurs in the southern ocean present in both the Sarmiento and Le Quéré (1996) and Sarmiento et al.(1998) control runs. This reduction in convective mixing and increased density stratification significantly improved the CFCs (Robitaille and Weaver, 1995; England and Hirst, 1995) simulations in the ocean.

The reduction in deep convective mixing and increased density stratification significantly reduced the oceanic uptake of anthropogenic CO_2.in our the control run. For our control run (run 1), the annual mean oceanic uptake of CO_2 for the 1980–89 period was 1.35 Gt C yr^{-1}. This was less than the 1980–89 uptake of the Sarmiento et al. (1998) control run of 2.0 Gt C yr^{-1} and IPCC estimate of 2.0 ± 0.8 Gt C yr^{-1}. However, recent estimates based on atmospheric oxygen and CO_2 measurements infer less oceanic uptake than the IPCC value. For 1991 to 1994, Keeling et al. (1996) estimated that the annual mean ocean uptake was 1.7 ± 0.9 Gt C yr^{-1}. For the same period, run 1 predicted an uptake of 1.6 Gt C yr^{-1}.

With our control run having a greater density stratification in high latitudes, the stratifying effect of climate change (warming and increased precipitation) produced a less dramatic change in the ocean stratification than the Sarmiento and Le Quéré (1996) and Sarmiento et al. (1998) studies, where their ocean was initially weakly stratified with large regions of deep convective mixing. Reduced changes in the ocean stratification with climate change reduced the anthropogenic CO_2 forcing feedback in our model. These results show that this feedback was sensitive to both the control run estimate of ocean CO_2 uptake and changes in surface forcing that occurred with climate change. Reducing the uncertainty in the present estimate of anthropogenic CO_2 uptake is essential to improving the prediction of the future uptake CO_2 by the ocean.

The parameterization of the biology differed significantly between our study and the work by Sarmiento and Le Quéré (1996) and Sarmiento et al. (1998). To investigate the rôle the biological parameterization has on the response of the ocean CO_2 uptake feedback to climate change, we compared the value of natural carbon cycle feedback from these studies. The natural carbon cycle feedback is the only feedback dependent on the biological parameterization. Our natural carbon cycle feedback was +33 Gt C by 2100. For our runs, changes in the biological export production opposed changes in the re-supply of carbon and alkalinity to the surface, because our runs linked biological export production to circulation through a nutrient-dependent parameterization of export production. This linkage of the export production to the circulation significantly attenuated the ocean circulation feedback. The Sarmiento and Le Quéré (1996) study severed this linkage by maintaining a nearly constant export of organic carbon from the euphotic zone and produced a natural carbon cycle feedback of 111 Gt C for approximately the same period (calculated from the difference between the solubility and biological models). Similarly, Sarmiento et al. (1998) estimated a natural carbon cycle feedback of +108 Gt C by 2065. Our biological parameterization has reduced the natural carbon cycle feedback by 78 Gt C from Sarmiento and Le Quéré (1996).

The uncertainty in the natural carbon cycle feedback depends on the linkage between re-supply of nutrients to the euphotic zone and export production. On millennium time-scales, these 2 processes are linked, but on shorter time-scales, de-coupling may occur. One expects high nutrient-low chlorophyll regions of the surface ocean to exhibit the greatest de-coupling of the export production from the nutrient supply, since biological production in these regions is not limited by macro nutrients. The emergence of iron in controlling biological production in these regions (Coale et al., 1996; La Roche et al., 1996) introduces one mechanism to de-couple upwelling from export production, if the supply of iron to the surface ocean differs from the supply of macro nutrients. Understanding the factors limiting production in high nutrient regions is essential to predicting the feedback between climate change and oceanic CO_2 uptake.

Although our study and the study by Sarmiento and Le Quéré (1996) predicted similar responses for the ocean uptake of CO_2 with climate change, the similarity hides large differences between the models. The sea surface warming feedback of both studies was similar. However, with our modification to the eddy parameterization and the biological formulation, we reduced the positive anthropogenic CO_2 forcing feedback by 80 Gt C

(−95%), and the negative natural carbon cycle feedback by 78 Gt C (−136%). The net result was only a 2 Gt C reduction in ocean CO_2 with climate change.

5. Summary

Our results show a reduction in the future rate of oceanic CO_2 uptake with climate change. Between 1850 and 2100, we predict that climate change will reduce oceanic uptake of anthropogenic CO_2 by 56 Gt C which equates to an increase in atmospheric CO_2 by 28 ppm (8% of the prescribed increase in atmospheric CO_2). This positive feedback is the result of the warming of the sea surface temperature and the changes in ocean circulation. During the 1850–2100 period, our predicted climate change warmed the sea surface temperature and reduced the accumulated oceanic CO_2 uptake by 48 Gt C, and slowed-down the ocean circulation, which reduced the oceanic uptake of anthropogenic CO_2 by 41 Gt C, but increased the storage of carbon in the ocean by 33 Gt C. In comparison, Cao and Woodward (1998) predicted that between 1861 and 2070, the terrestrial biosphere would increase its uptake of carbon by 309 Gt C with climate change.

For the 1995–2010 period, the reduced oceanic CO_2 uptake caused by climate change can be compared to the emission targets of the Kyoto agreement. During this period, the Kyoto agreement aims to reduce CO_2 emissions by ca 8 Gt C (0.5 Gt C yr^{-1}), while we predicted climate change might reduce oceanic CO_2 uptake by 5 Gt C (0.3 Gt C yr^{-1}). Although the annual oceanic CO_2 uptake estimate has large uncertainty, ±40% (±0.8 Gt C yr^{-1}), only the climate change effect on oceanic uptake would produce a systematic reduction in uptake that may significantly impact on the goals of the Kyoto agreement.

6. Acknowledgements

We would like to thank T. Trull, J. Parslow and J. Church for helpful discussions. Funding for this research was provided by the Australian Climate Change Research Program.

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