Carbon fluxes in the Arctic Ocean—potential impact by climate change

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Because of its ice cover the central Arctic Ocean has not been considered as a sink of atmospheric carbon dioxide. With recent observations of decreasing ice cover there is the potential for an increased air–sea carbon dioxide flux. Though the sensitivity of the carbon fluxes to a climate change can at present only be speculated, we know the responses to some of the forcing, including: melting of the sea ice cover make the air–sea flux operate towards equilibrium; increased temperature of the surface water will decrease the solubility and thus the air–sea flux; and an open ocean might increase primary production through better utilization of the nutrients.

The potential change in air–sea CO_2 fluxes caused by different forcing as a result of climate change is quantified based on measured data. If the sea ice melts, the top 100 m water column of the Eurasian Basin has, with the present conditions, a potential to take up close to 50 g C m^{-2}. The freshening of the surface water caused by a sea ice melt will increase the CO_2 solubility corresponding to an uptake of ~3 g C m^{-2}, while a temperature increase of 1 °C in the same waters will out-gas 8 g C m^{-2}, and a utilization of all phosphate will increase primary production by 75 g C m^{-2}.

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Most climate scientists do not question whether climate change might occur, but when and how this change will develop. The primary tools for prediction are coupled atmosphere–ocean Global Circulation Models (GCMs) run under different scenarios of atmospheric greenhouse gas increase. Ongoing research is being performed to improve the physics in the various components of these models, but they can never have reliable predictive capability unless they include the relevant processes and feedbacks in an appropriate way. One aspect of these processes and feedbacks relates to the atmosphere–ocean exchange of carbon dioxide. How will the present-day driving forces be affected by climate change?

The Barents Sea is a region where changes in the driving forces can have a significant impact on the thermohaline circulation, and therefore on climate. Much of the warm, high salinity Atlantic Water that flows into the Arctic Ocean enters the Barents Sea. During transit through the Barents Sea, the Atlantic water masses lose heat and thereby gain a sufficient density increase to produce intermediate and deep waters of the Arctic Ocean (e.g. Schauer et al. 1997; Anderson et al. 1999). The significant cooling of the surface waters in the Barents Sea also drives a flux of CO_2 from the atmosphere into the sea, a flux that is amplified by extensive biological primary production (Walsh 1989; Sakshaug et al. 1994). The formation of subsurface waters in the Arctic Mediterranean seas transport carbon, which partly include CO_2 of atmospheric origin (Anderson et al. 1998a).

The increased transport by the Gulf Stream during high NAO index (Hurrell 1995) conditions

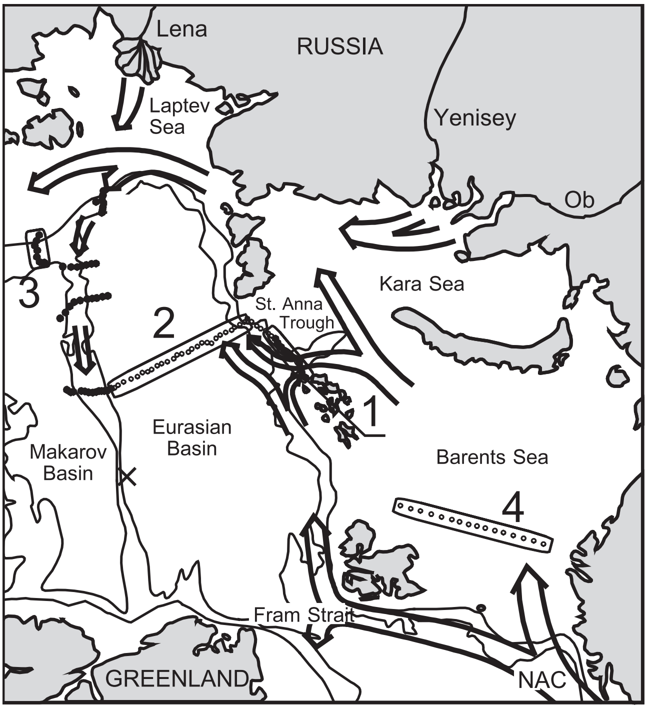

Fig. 1. Map of the eastern Arctic Ocean with the stations of the different cruises noted. The data used in this evaluation were from the regions indicated by boxes and marked 1 to 4. The Norwegian Atlantic Current is abbreviated as NAC.

contributes to increased temperatures along the Norwegian coast during winter. The increased temperature in the Atlantic Water (core of the West Spitsbergen Current) that has been observed in Fram Strait during the 1990s (Swift et al. 1997) has been considered one cause of the sea ice thinning observed in the Arctic Ocean (Rothrock et al. 1999). Reduced sea ice cover will promote increased atmospheric warming via albedo feedback (Morales Maqueda et al. 1999). A decrease in the sea ice thickness can have effects on the headwaters (regions of intermediate and deep water formation) of the thermohaline circulation, most notably through the increased freshwater export into the Greenland and Iceland seas (Steele et al. 1996). However, during recent years the modification of Atlantic Water in the Barents Sea has been emphasized as a significant contribution to the Scotland–Greenland overflow into the North Atlantic Ocean (Mauritzen 1996). The Barents Sea region is very little affected by an increased freshwater outflow from the Arctic Ocean, and therefore the contribution to the Scotland–Greenland overflow will only be marginally altered.

To predict future climate it is of great importance to improve our understanding of the processes controlling deep water formation and their sensitivity to possible climate change. The strength of the deep water formation will have a direct effect on the carbon cycle, and the strength of the deep water formation might be affected by the carbon cycle as the increasing atmospheric CO_2 content is the major cause of the potential climate change. To achieve an overall understanding of this coupling a description of the present functioning of the carbon cycle is required. It is also critical to assess how changes in the Earth's climate would alter cycling and partitioning of carbon between the ocean and atmosphere. Specifically, changes in ocean dynamics and ocean thermodynamics may cause significant changes in air–sea CO_2 cycling. Such feedbacks are not included in most coupled ocean–atmosphere climate models. Instead, the future atmospheric concentration of CO_2 is prescribed in these models.

Because of its ice cover, the Arctic Ocean has not been considered as a sink of carbon dioxide. The biological activity in the central Arctic Ocean is small and the air–sea flux must be negligible compared to the shelf seas. In this presentation we evaluate the potential change of air–sea carbon dioxide flux in the Arctic Ocean caused by climate change.

Present carbon fluxes

The air–sea carbon flux have been estimated from budget computations to be 110×10^{12} g C yr^{-1} in the Arctic Mediterranean seas (Arctic Ocean and Nordic seas) (Lundberg & Haugan 1996) and 24×10^{12} g C yr^{-1} in the Arctic Ocean, including the shelf seas (Anderson et al. 1998b). If these estimates are comparable, uptake by the Nordic seas equals 86×10^{12} g C yr^{-1}. These estimates have significant uncertainties, but they point to the importance of the Nordic seas for the uptake of CO_2 from the atmosphere. For the Arctic Ocean the shelf seas are the most important, with the uptake from the atmosphere being 9×10^{12} g C yr^{-1} in the Barents Sea (Fransson et al. 2001). The other very important region is likely the Chukchi Sea. On a global perspective these fluxes are not impressive, but the potential for an increase in a climate change scenario is considerable.

Data and computation models

The data used in this evaluation were collected during the ACSYS 1996 cruise to the eastern Arctic Ocean on R/V *Polarstern* (Augstein 1997) and a cruise to the Barents Sea in July 1999 on board the Norwegian R/V *Jan Mayen* (with Dr. P. Wassmann, University of Tromsø, as chief scientist). Figure 1 shows the positions of the stations occupied during these cruises. Standard analytical methods were applied for the determination of total dissolved inorganic carbon (C_T), total alkalinity (A_T) and pH. Total dissolved inorganic carbon was determined by gas extraction of an acidified water sample followed by coulometric titration (Johnson et al. 1985, 1987). Typically the precision was ± 4 µmol kg^{-1} and the accuracy was set by running a certified reference material supplied by A. Dickson (Scripps Institution of Oceanography) at each change of cell solution. Total alkalinity was determined on board the ship by titrating the samples with 0.1 M HCl and measuring the change in pH with a potentiometric method (Haraldsson et al. 1997). The precision

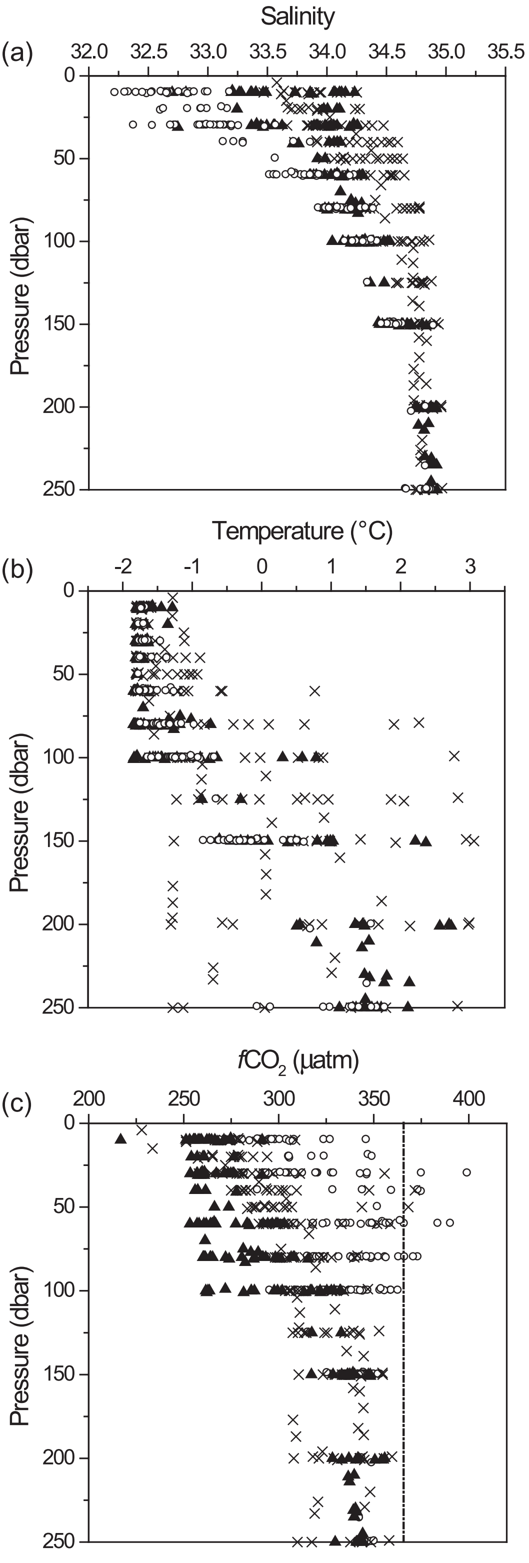

Fig. 2. Depth profiles of (a) salinity, (b) temperature and (c) fCO_2 for all stations occupied during the ACSYS-96 cruise. Crosses are from stations in the St. Anna Trough, filled triangles in the deep Arctic Ocean, and open circles along the shelf slope north of the Laptev Sea.

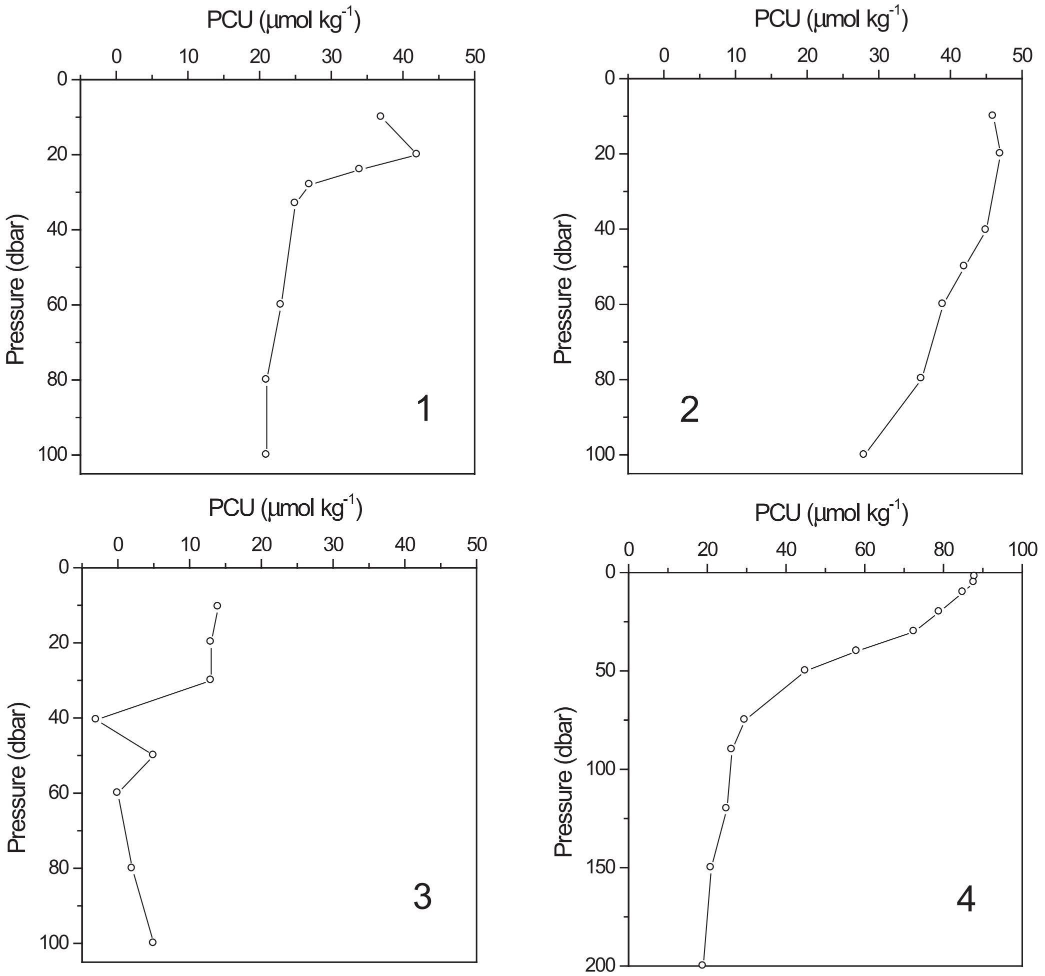

Fig. 3. The Potential Carbon Uptake (PCU) computed as the difference between that computed from the observed A_T and a fCO_2 of 365 µatm and the measured C_T, for four regions of the eastern Arctic Ocean (see Fig. 1 for positions). Note the different scales for the Barents Sea profile.

of the data was ± 2 µmol kg^{-1} and the accuracy was determined in the same way as for C_T. The determination of pH was performed spectrophotometrically using m-cresol purple as indicator (Clayton & Byrne 1993; Lee & Millero 1995). The average standard deviation was 0.0006 ± 0.0006 and the accuracy was set by the accuracy of the temperature measurements and the accuracy in the determination of the stability constant of the dye, being approximately ± 0.002 (Dickson 1993). The fugacity of carbon dioxide (fCO_2) was calculated with the CO_2 program developed by Lewis & Wallace (1998) as were all other computations of the carbonate system. The total hydrogen ion scale and the carbon dioxide constants from Roy et al. (1993, 1994) were used.

Results and discussion

First we investigate the present status with regard to fCO_2 saturation in the central Arctic Ocean, using the ACSYS-96 data (Fig. 2). All of the top 250 m has an Atlantic origin, where the upper ~100 m has a contribution of melted sea ice, resulting in lower salinity than the underlying more or less unmodified Atlantic Water (Rudels et al. 1996). At the shelf slope north of the Laptev

Sea river runoff has been added, decreasing the salinity even further. The temperature signal is very variable below 100 m in the St. Anna Trough region, a result of the variable composition of the waters flowing off the shelf. The fCO_2 level computed from C_T and pH show under-saturation all through the top 250 m, indicating that the surface water of the central Arctic Ocean could take up atmospheric CO_2 if sea ice were not hampering this. The fCO_2 is quite variable in the top 100 m, reflecting the different conditions in the regions, but more constant around 340 µatm in the deeper layers. The under-saturation of ~25 atm in the warm water at depths below 150 m indicates that, during its flow north in the Norwegian Atlantic Current, the Atlantic Water does not have time to equilibrate with the atmosphere. This is a result of the heat loss to the atmosphere being a faster process than the air–sea flux of CO_2.

To calculate the magnitude of the potential uptake by the surface water we used the measured A_T and a fCO_2 of 365 µatm (close to that in the atmosphere in 1996) to compute C_T corresponding to that of a water in saturation with respect to the CO_2 (C_T^{sat}). The computation has been made for three regions occupied during the ACSYS-96 cruise (numbered 1 to 3 in Fig. 1). Furthermore, we made a corresponding computation for the data of the 1999 Barents Sea cruise. The mean profiles of the Potential Carbon Uptake (PCU) equal to the difference between C_T^{sat} and the measured C_T (C_T^{meas})—

$$\mathrm{PCU} = C_T^{sat} - C_T^{meas}$$

—are presented in Fig. 3.

As mentioned above, the water is under-saturated well into the Atlantic Layer, corresponding to a PCU of about 20 µmol kg^{-1} at ~200 m depth in the Barents Sea. The PCU is larger in the Eurasian Basin relative to the St. Anna Trough, which could be explained by the larger fraction of the Fram Strait Branch of Atlantic Water found in this region of the Arctic Ocean. The lowest PCU is found in the data over the Makarov Shelf slope, which might be due to a longer residence time over the shelves with more time to take up atmospheric CO_2. The largest PCU is found in the Barents Sea and this is caused by the seasonal consumption of CO_2 by photosynthesis. However, this water will spend a significant time in open water in the Barents Sea before entering the ice covered central Arctic Ocean. Hence, it is not relevant to discuss this profile in relation to future climate caused changes in the ice cover. Biological activity is much less in the central Arctic Ocean than in the shelf seas, making the data less seasonal dependent. Furthermore, the seasonal signals that are transferred from the shelves into the central Arctic Ocean are smoothed out by the residence time of the surface water—on the order of 10 years. However, some of the variability seen in the top ~100 m close to the shelf seas can be a result of this effect.

To make quantitative estimates for the central Arctic Ocean we assume that the top 100 m has a potential to take up atmospheric CO_2. Integrating down to that depth gives potential uptakes for the St. Anna Trough region of 35 g C m^{-2}, for the Eurasian Basin 48 g C m^{-2} and for the Makarov Shelf slope 7 g C m^{-2}. These uptakes are only to balance the under-saturation of today and therefore only a one-time event. The annual uptake potential will be discussed below.

It is only possible to increase the CO_2 uptake of ice-covered areas from the atmosphere by reducing ice cover, to get closer to equilibrium. However, with less ice cover other factors will also change, like salinity and possibly also temperature. If 2 to 3 m of ice (typical of multi-year ice in the eastern part of the Arctic Ocean) melts and the meltwater mixes into the top 100 m, the salinity decreases with 2 to 3 %. This freshening increases the solubility of CO_2 to a degree that it might take up an additional ~3 g C m^{-2}. If the climate warms to the extent that the sea ice melts it might also cause a temperature increase of the surface water, where each degree results in a potential out-gassing of 8 g C m^{-2}.

Another factor that will accompany a disappearance of the sea ice (at least during summer) is better light conditions for biological primary production. At present about 1/3 of the phosphate supplied to the central Arctic Ocean is consumed (Fig. 4), most likely largely over the shelves, while the rest is available for photosynthesis within the central Arctic Ocean. If all the surplus phosphate were to be utilized it would correspond to a carbon consumption of 75 g C m^{-2}, if the classical RKR phosphate–carbon ratio of 1:106 is used. It is not likely that all phosphate is used on an annual basis as other factors play an important role in the draw-down of carbon from the surface layer. These include the effect of phytoplankton and zooplankton species on sedimentation as well as remineralization in the surface layer at the end of the productive season. To address how the eco-

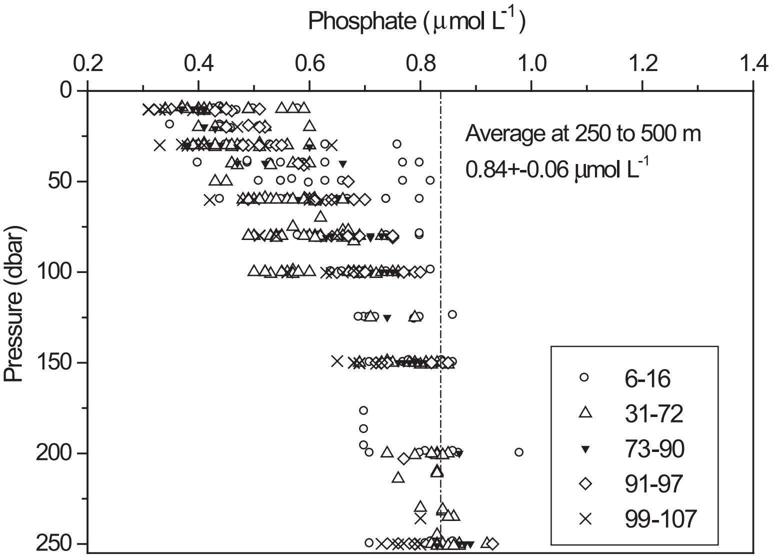

Fig. 4. The phosphate distribution in the top 250 m of the eastern Arctic Ocean during the ACSYS-96 cruise. Stations 6–16 are from the St. Anna Trough, 31–72 from the Eurasian Basin section, 73–90 from over the Lomonosov Ridge, 91–97 from the Makarov shelf slope and 99–107 from the Laptev Sea shelf slope. The mean concentration in the Atlantic Layer water (250 to 500 m depth) equals 0.84 ± 0.06 µmol L^{-1}.

logical system effects the draw-down of carbon and sedimentation is very complicated and far beyond the scope of this relatively simple computation of potential carbon flux changes with climate change.

Using the area of the whole deep central Arctic Ocean (5.8×10^{12} m^2), we get potential uptake of atmospheric CO_2 from the different processes discussed above (Table 1). The different potential uptakes cannot be summed up, but the relative importance of the different processes is illustrated. The observed PCU and increased export production are quite significant relative to the global uptake of anthropogenic CO_2 (~2000 × $10^{12\ g}$ C yr^{-1}).

To elucidate the annual potential uptake, we have to consider the volume input to the top 100 m of the central Arctic Ocean. Inflow to the surface mixed layer and halocline (contributing to the top 100 m) is from the Pacific and the Atlantic (both over the Barents Sea and through Fram Strait) as well as from rivers. The question is what volume fluxes to apply for the different inflows. Anderson et al. (1998b) used the literature values of the total exchange with the surrounding oceans and, by applying salt and mass balance, computed the contributions to the different water masses of the Arctic Ocean. In short, these add up to 0.9×10^6 m^3 s^{-1} from the Atlantic, 0.4×10^6 m^3 s^{-1} from the Pacific, and 0.1×10^6 m^3 s^{-1} from runoff. If we take the sum of these inflows and multiply the estimated potential uptake of atmospheric carbon dioxide (divided by 100 m) by the different processes, we get the uptake from the atmosphere as shown in Table 2. The potential uptakes used for

Table 1. The potential uptake of atmospheric CO_2 in the top 100 m of the total deep central Arctic Ocean by different processes.

Process	Potential uptake (10^{12} g C)
PCU[a]	278
Salinity decrease of 2 %	14
Temperature increase of 1 °C	-46
Export production corresponding to 50 % of the available phosphate	218

[a] The PCU value is taken from the Eurasian Basin profile.

Table 2. The potential annual uptake of atmospheric CO_2 by different processes in the deep central Arctic Ocean. The values are achieved by multiplying the average computed uptake of CO_2 by each process with the volume flux to the top 100 m (a total of 1.4×10^6 m^3 s^{-1} according to Anderson et al. 1998b).

Process	Potential uptake (10^{12} g C yr^{-1})
PCU[a]	21
Salinity decrease of 2 %	1
Temperature increase of 1 °C	-4
Export production corresponding to 50 % of the phosphate supply through St. Anna Trough	17

[a] The PCU value is taken from the St. Anna Trough profile.

these calculations are: 0.48 g C m^{-2} by PCU, 0.03 g C m^{-2} by freshening, -0.08 g C m^{-2} by 1 °C warming, and 37 g C m^{-2} by utilization of half the phosphate available. These estimates are based on the assumption that our estimated uptake potential is representative for the whole Arctic Ocean.

The annual fluxes of Table 2 are not large in a global view, but the observed PCU and increased export production is of the same order as the air–sea fluxes within the Arctic Ocean, including the shelf seas (Anderson et al. 1998b). From the results presented in Table 2 it is easy to evaluate the impact that changes in the strength of the oceanic circulation have on the air–sea CO_2 flux, as they are directly proportional.

Another effect of a disappearance of the sea ice cover is that it might increase vertical mixing and thus bring up more nutrients to the photic zone. This could be the situation if an open ocean persist under a time long enough for the resulting freshening to be exported to the adjacent seas. Wallace et al. (1987) evaluated the vertical mixing coefficient (K_z) to 2×10^{-6} m^2 s^{-1}. If this were doubled, the increased supply of phosphate (F^{PO_4}) would according to Fick´s Law be:

$$F^{PO_4} = K_z \frac{d[PO_4]}{dz},$$

where, according to Fig. 4,

$$\frac{d[PO_4]}{dz} = \frac{0.3\cdot10^{-3}}{100}\left[\frac{\mathrm{mol\cdot m^{-3}}}{\mathrm{m}}\right]$$

Making this computation and converting to carbon units we get:

> 0.3×10^{-3} (molP m^{-3}) × 0.01 (m^{-1}) × 2 × 10^{-6} ($m^2 s^{-1}$) × 60 × 60 × 24 × 365 (s yr^{-1}) × 106 (molC $molP^{-1}$) × 12 (g C $molC^{-1}$) = 0.24 g C m^{-2} yr^{-1}.

This flux has to be multiplied by the area of the central Arctic Ocean to be compared to the values of Table 2. This gives an increased annual export production of 1.4×10^{12} g C yr^{-1}. This is about 10 % of the increased export production potentially caused by the horizontal supply of phosphate and is well within the uncertainties of that calculation. The vertical mixing of nutrients will also drive a flux of carbon to the surface layer, lowering the potential effect of this process in driving an air–sea flux of CO_2.

All the above calculations have large uncertainties, but the objective of this contribution is to compute the potential changes for certain scenarios. Hence, these numbers should be seen relative to each other, and as indicators of the potential magnitude of change in air–sea flux of CO_2 that might be caused by a climate change. For instance if the Arctic Ocean sea ice melts there is a potential effect to take up ~280×10^{12} g C yr^{-1} from the atmosphere just to reach CO_2 equilibrium and another ~220×10^{12} g C yr^{-1} by increased export production consuming half of the available phosphate. Smaller alteration in the uptake might also be caused by changes in salinity and temperature coupled to this climatic effect.

Summary and conclusions

We have shown that climate change involving significant decreases in the ice cover of the central Arctic Ocean could cause a one-time uptake of atmospheric CO_2 on the order of 500×10^{12} g C, which is substantial relative to the global oceanic uptake of anthropogenic CO_2. This estimate is based on two processes: getting the surface water to equilibrium with the atmospheric CO_2 and an increased export production, consuming half of the available phosphate. On the other hand, the change in the annual uptake is more modest and in a scenario of an ice-free central Arctic Ocean (at least during the productive summer), the potential uptake of CO_2 from the atmosphere reaches ~40×10^{12} g C yr^{-1}. This is the contribution by an export production corresponding to half the influxing phosphate and by reaching CO_2 saturation in the water flowing into the central Arctic Ocean. If the oceanic circulation strengthens or weakens this air–sea flux will change accordingly. It is not possible to foresee the magnitude of changes of the processes discussed, but at least it is possible to evaluate which processes have the largest impact and in what direction.

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