An increasing CO_2 sink in the Arctic Ocean due to sea-ice loss

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[1] The Arctic Ocean and adjacent continental shelf seas such as the Chukchi and Beaufort Seas are particularly sensitive to long-term change and low-frequency modes of atmosphere-ocean-sea-ice forcing. The cold, low salinity surface waters of the Canada Basin of the Arctic Ocean are undersaturated with respect to CO_2 in the atmosphere and the region has the potential to take up atmospheric CO_2, although presently suppressed by sea-ice cover. Undersaturated seawater CO_2 conditions of the Arctic Ocean are maintained by export of water with low dissolved inorganic carbon content and modified by intense seasonal shelf primary production. Sea-ice extent and volume in the Arctic Ocean has decreased over the last few decades, and we estimate that the Arctic Ocean sink for CO_2 has tripled over the last 3 decades (24 Tg yr^{-1} to 66 Tg yr^{-1}) due to sea-ice retreat with future sea-ice melting enhancing air-to-sea CO_2 flux by ~28% per decade. **Citation:** Bates, N. R., S. B. Moran, D. A. Hansell, and J. T. Mathis (2006), An increasing CO_2 sink in the Arctic Ocean due to sea-ice loss, *Geophys. Res. Lett.*, *33*, L23609, doi:10.1029/2006GL027028.

1. Introduction

[2] The polar continental shelves are an active interface for the terrestrial, oceanic and atmospheric components of the Earth's biogeochemical system. Physical transformations and seasonal sea-ice dynamics play a major role in modifying the water-masses, ecosystems and the carbon cycle on the Chukchi Sea shelf, the western part of the Beaufort Sea shelf, and adjacent Arctic Ocean. Approximately 0.8 Sv (10^6 m^3 s^{-1}) of Pacific Ocean water from the sub-polar Bering Sea enters the Arctic Ocean region through the Bering Strait [*Aagaard et al.*, 1985; *Roach et al.*, 1995; *Woodgate and Aagaard*, 2005; *Woodgate et al.*, 2005], fanning out across the shallow (<50 m deep) and expansive shelf of the Chukchi Sea [*Codispoti et al.*, 2005] before flowing into the Canada Basin of the central Arctic Ocean (Figure 1). Major outflows from the Chukchi Sea shelf of ~0.1–0.3 Sv each occur through Herald Valley, Long Strait and Barrow Canyon [*Woodgate and Aagaard*, 2005]. During transit across the Chukchi Sea shelf, the physical and biogeochemical properties of these water-masses are highly modified due to seasonal light and temperature changes, seasonal sea-ice dynamics, and primary production. The preconditioning and transformations of carbon over the Chukchi Sea continental shelf (and other polar shelves) impacts air-sea CO_2 gas exchange in the Arctic Ocean.

[3] A comprehensive physical and biological oceanography survey of the Chukchi Sea, western Beaufort Sea shelf, and the adjacent Canada Basin of the Arctic Ocean was conducted in 2002 to 2004 under the auspices of Western Arctic Shelf-Basin Interactions (SBI) project [*Grebmeier and Harvey*, 2005; *Codispoti et al.*, 2005; *Bates et al.*, 2005a]. The SBI project provided opportunities to determine dissolved inorganic carbon (DIC), alkalinity (TA) and seawater partial pressure of CO_2 (pCO_2) distributions and CO_2 dynamics on the continental shelf and Arctic Ocean during "*winter/springtime*" sea-ice covered periods and during the "*summertime*" peak period of seasonal sea-ice melt and retraction toward the pole.

2. Data and Methods

[4] As part of the hydrographic and biogeochemical sampling of the SBI project, water-column DIC and alkalinity samples were collected from the icebreaker USCGC *Healy* in 2002 and 2004 at numerous CTD stations across the Chukchi Sea and Canada Basin (Figure 1) of the Arctic Ocean. Two cruises sampled during the sea-ice covered, "*winter/springtime*" period [HLY02-01, 8 May–12 June 2002; HLY04-02; 15 May–23 June 2004]. In addition, there were two "*summertime*" cruises [cruise HLY02-03, 16 July–26 August 2002; HLY04-02; 17 July–26 August 2002] and one late-summer cruise [HLY04-04; 1 September–30 September 2004] during the sea-ice minima period.

2.1. Calculation of Seawater pCO_2

[5] Seawater pCO_2 concentrations were calculated from DIC and TA [*Bates et al.*, 2005a; *Bates*, 2006], and temperature and salinity data [http://catalog.eol.ucar.edu/sbi/]. Here, the carbonic acid dissociation constants of *Mehrbach et al.* [1973], as refit by *Dickson and Millero*, [1987] were used to determine seawater pCO_2 (see auxiliary materials)[1] [*Bates*, 2006]. The majority of samples from the Chukchi Sea and Canada Basin were collected from waters with a temperature range of ~−1.8°C to 0°C, and computed seawater pCO_2 had an average difference of ~2.5 μatm for different carbonic acid dissociation constants [*Bates*, 2006].

2.2. Determination of Air-Sea CO_2 Flux

[6] The net air-sea CO_2 flux (F) was determined by the following formula [$F = k\sigma\ (\Delta pCO_2)$] where k is the transfer velocity, σ is the solubility of CO_2 and, ΔpCO_2 is the difference between atmospheric and oceanic partial pres-

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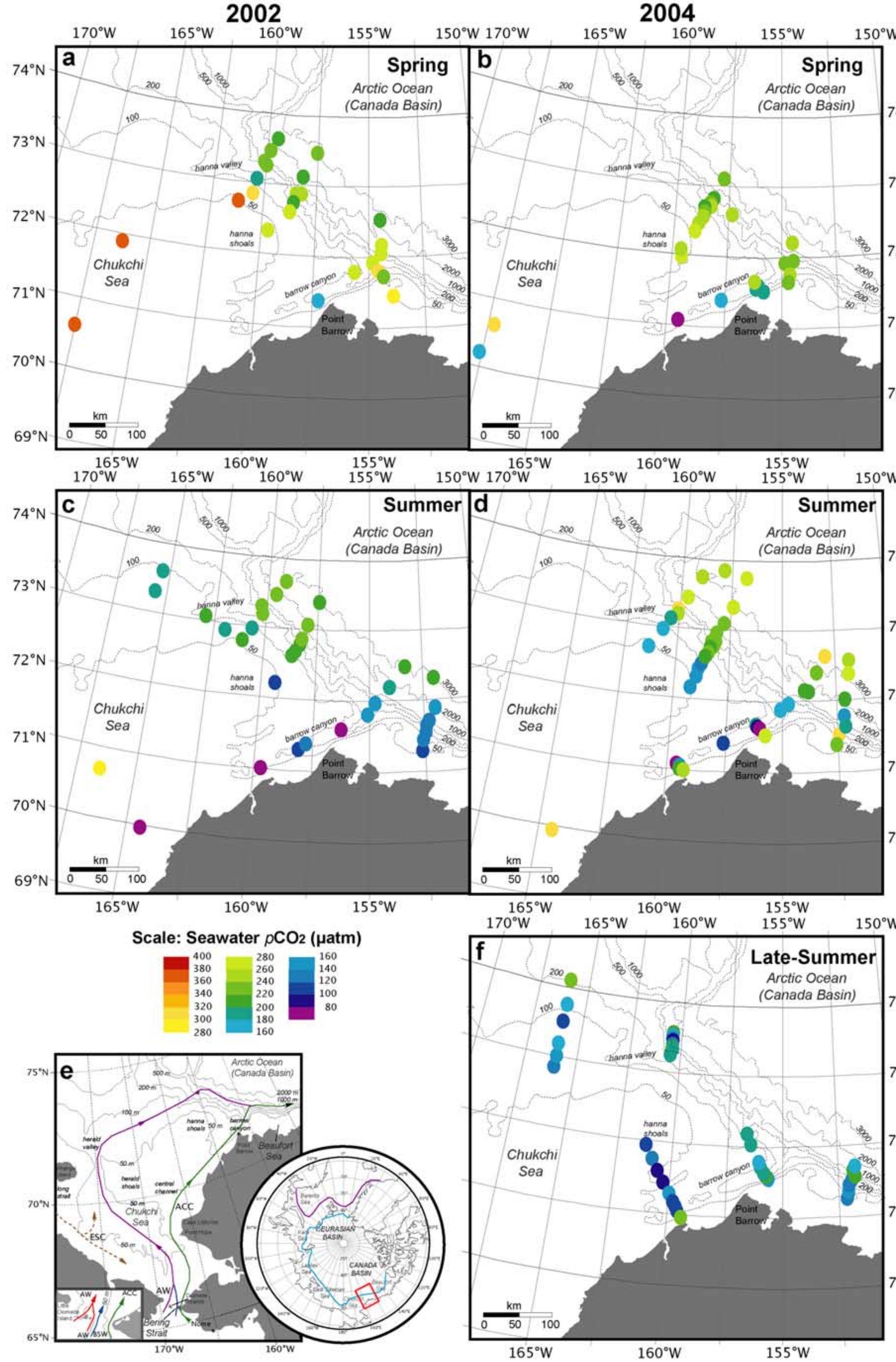
2002
2004
a
Spring
b
Spring
Arctic Ocean (Canada Basin)
Chukchi Sea
hanna valley
hanna shoals
barrow canyon
Point Barrow
c
Summer
d
Summer
f
Late-Summer
Scale: Seawater pCO_2 (µatm)
400
380
360
340
320
300
280
260
240
220
200
180
160
140
120
100
80
e
Bering Strait
Nome
ESC
ACC
AW
EURASIAN BASIN
CANADA BASIN
km
0 50 100

Figure 1

sures of CO_2. The ΔpCO_2, or air-sea CO_2 disequilibrium, sets the direction of CO_2 gas exchange while k determines the rate of air-sea CO_2 transfer. Gas transfer velocity-wind speed relationships for short-term wind conditions based on a quadratic (U^2) dependency [*Wanninkhof*, 1992] between wind speed and k were used to determine air-sea CO_2 fluxes (see auxiliary materials for details). Atmospheric pCO_2 data from Point Barrow in Alaska (http://www.cmdl.noaa.gov) and seawater pCO_2 data from each CTD station were used to compute ΔpCO_2 values. Wind speed data (i.e., daily averages from 6-hourly observations) from the NCEP (National Center for Environmental Prediction)/NCAR (National Center for Atmospheric Research) reanalysis 2 data assimilation model (i.e., NNR data, http://www.cdc.noaa.gov/cdc/data.ncep.html) were used rather than shipboard observations to calculate k values. The spatial resolution of NNR data is 2.5° by 2.5° for the region of interest (i.e., 70°N–75°N; 165°W–150°W).

[7] The sea-ice barrier to gas exchange was accounted for by correcting air-sea CO_2 fluxes to the observed sea-ice conditions (data from SBI shipboard reports and satellite data; http://catalog.eol.ucar.edu/sbi/). For example, with sea-ice coverage conditions of 95%, 50%, and 0%, multipliers of 0.05, 0.5 and 1.00, respectively, were applied to calculation of F. In those regions with 100% sea-ice coverage, air-sea CO_2 gas exchange can occur through leads and fractures in the ice, and also potentially through sea-ice [*Semiletov et al.*, 2004]. In these regions, a multiplier of 0.01 (equivalent to 99% sea-ice cover) was used to allow for minor rates of air-sea CO_2 exchange through leads, fractures and brine channels.

3. Results and Discussion

3.1. Chukchi Sea pCO$_2$ Distributions and Air-Sea CO$_2$ Fluxes

[8] Large seasonal winter/springtime to summertime changes in seawater pCO_2 distributions were observed on the Chukchi Sea shelf adjacent to the Canada Basin. In the spring of 2002 and 2004, during sea-ice covered conditions, surface layer seawater pCO_2 concentrations were undersaturated across the Chukchi Sea shelf (Figures 1a and 1b). At the Chukchi Sea shelf edge (i.e., water depth <200 m), seawater pCO_2 values ranged from ~200–280 μatm (ΔpCO_2 of ~–100–180 μatm) with very low seawater pCO_2 values (<100 μatm; ΔpCO_2 of >280 μatm) observed in the region of the Barrow Canyon polynya. Although mixed layer waters were undersaturated with respect to atmospheric CO_2, the influx of CO_2 to the ocean through air-sea gas exchange was low (<4 mmoles CO_2 m^{-2} d^{-1}), due primarily to suppression by sea-ice cover (~95–100%) [*Bates*, 2006].

[9] By mid-summer, sea-ice had retreated from Bering Strait northwards to the margins of the Canada Basin (Figures 1c and 1d). Across much of the Chukchi Sea, the brief exposure of nutrient laden surface layer (0–~30 m) waters of Pacific origin flowing northward from Bering Strait over the sea-ice free Chukchi Sea shelf led to an intense seasonal bloom of phytoplankton [*Cota et al.*, 1996; *Hill and Cota*, 2005]. In the region of highest primary production, where there was complete exhaustion of inorganic nutrients, removal of DIC [*Bates et al.*, 2005a] and organic carbon [*Bates et al.*, 2005b; *Moran et al.*, 2005] from the surface waters, very low seawater pCO_2 conditions were observed (<120–, 180 μatm) in 2002 and 2004 (Figures 1 and 2a). Extremely low seawater pCO_2 values (<80 μatm; ΔpCO_2 of >300 μatm) were observed in Barrow Canyon (Figures 1c and 1d), through which the northern outflow of Chukchi Sea shelf waters are funneled (Figure 1e). By late summertime in 2004, seawater pCO_2 conditions remained low (<100–160 μatm) prior to seasonal sea-ice advance (Figure 1f).

[10] During the sea-ice free period, surface waters of the Chukchi Sea were highly undersaturated with respect to atmospheric CO_2 (ΔpCO_2 of ~>200–300 μatm), and the region acted seasonally as a very strong sink for atmospheric CO_2. *Air*-to-*sea* CO_2 fluxes (computed with either NNR data or synoptic shipboard observations of wind-speed) had a median value of ~50 mmoles CO_2 m^{-2} d^{-1} (for sea-ice conditions of <5%; see Figure 2b and auxiliary materials Figure S1), with higher values (>120 mmoles CO_2 m^{-2} d^{-1}) in some regions of the Chukchi Sea. The mean *air*-to-*sea* CO_2 fluxes on the Chukchi Sea shelf for summer 2002, summer 2004, and late summer 2004 were 36 ± 25 mmoles CO_2 m^{-2} d^{-1}, 53 ± 27 mmoles CO_2 m^{-2} d^{-1}, and 61 ± 19 mmoles CO_2 m^{-2} d^{-1}, respectively.

[11] Our estimates of mean *air*-to-*sea* CO_2 flux for the Chukchi Sea were three times higher than previous estimates of *Kaltin and Anderson* [2005], who used a simple carbon budget to indirectly estimate air-sea CO_2 fluxes. In their method, *air*-to-*sea* CO_2 flux was computed from the difference in DIC inventories (from very limited data) of waters flowing northward through Bering Strait and at the Chukchi Sea shelf edge, accounting for photosynthetic

Figure 1. Distributions of seawater pCO_2 in the Chukchi Sea and Canada Basin. (a) Spring 2002 surface distributions of seawater pCO_2 (μatm) (HLY-02-01 cruise; 5 May–15 June 2002). (b) Spring 2004 surface distributions of seawater pCO_2 (μatm) (HLY-04-02 cruise; 17 May–21 June 2004). (c) Summer 2002 surface distributions of seawater pCO_2 (μatm) (HLY-02-03 cruise; 17 July–26 August 2002). (d) Summer 2004 surface distributions of seawater pCO_2 (μatm) (HLY-04-03 cruise; 16 July–26 August 2002). (e) Water mass circulation of the Chukchi Sea, western Beaufort Sea, and adjacent Arctic Ocean. Generalized circulation flows are shown. Water transiting through Bering Strait is composed of warmer, fresher Alaskan Coastal Current (ACC; green) waters in the east [*Woodgate and Aagaard*, 2005], Bering Shelf Water (BSW; blue) in central Bering Strait and colder, saltier, more nutrient-rich Anadyr Water (AW; red) to the west. Bering Shelf and Anadyr water merge into Bering Sea water (purple) within the Chukchi Sea [*Roach et al.*, 1995; *Woodgate and Aagaard*, 2005]. East Siberian Current (ESC) water intermittently flows into the Chukchi Sea from the west through Long Strait. (inset) Map of the Arctic Ocean with locations of Chukchi Sea, Beaufort Sea, and Canada Basin. Minimum summertime (cyan) and maximum wintertime (purple) sea-ice extent taken from 1972–2002 SSMI data. (f) Late-summer 2004 surface distributions of seawater pCO_2 (μatm) (HLY-04-04 cruise; 1 September–30 September 2004). Contours (dashed brown line) indicate approximate distributions of sea-ice cover from SBI shipboard reports.

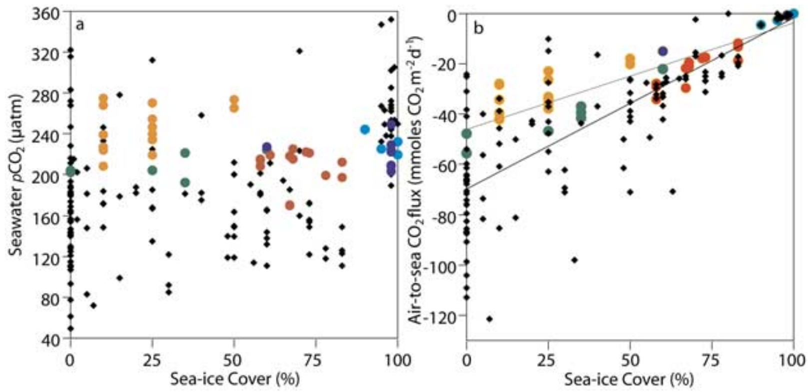

Figure 2. (a) Seawater pCO_2 (μatm) against sea-ice cover (%) from the Chukchi Sea (closed diamond symbol) and Canada Basin. (b) Air-to-sea CO_2 flux (mmoles CO_2 m^{-2} d^{-1}) estimated for the Chukchi Sea (closed diamond symbol) and Canada Basin. Canada Basin data include: "*winter*/*springtime*" cruises (HLY02-01, 8 May–12 June 2002, blue symbol; HLY04-02, 15 May–23 June 2004, cyan symbol), and "*summertime*" cruises (HLY02-01, 8 May–12 June 2002, red symbol; HLY04-02, 15 May–23 June 2002, orange symbol; HLY04-04, 1 September–30 September 2004, green symbol).

uptake of CO_2 as new production (from inorganic nutrient changes). The SBI observations indicate that the DIC content of surface and upper halocline winter/springtime waters across the Chukchi Sea and at Bering Strait (particularly in Alaskan Coastal Current water) were higher by >100 μmoles kg^{-1} [*Bates et al.*, 2005a; *Bates*, 2006] compared to *Kaltin and Anderson* [2005]. In addition, the DIC of surface water outflowing the Chukchi Sea through Barrow Canyon (and Hanna Valley) [*Bates et al.*, 2005a; *Bates*, 2006] was much lower than observed by *Anderson and Kaltin* [2001] near the Chukchi Cap.

3.2. Canada Basin pCO$_2$ Distributions and Air-Sea CO_2 Fluxes

[12] At sampling stations in the Canada Basin of the Arctic Ocean (here defined as deeper than 200 m bottom depth), surface waters of the Polar Mixed Layer (PML; upper 30–50 m) were cold (~−0.5 to −1.7°C) during winter/springtime, with a slightly warmer range (~+2 to −1.7°C) during summertime. Seawater pCO_2 contents had a small range (~200–270 μatm; Figures 1 and 2a) and the average seawater pCO_2 value for the PML was 225 ± 22 μatm. Surface layer seawater pCO_2 showed little seasonal change between periods of high sea-ice cover (winter/springtime) and sea-ice minima conditions during summertime (Figure 2a). Furthermore, the average temperature corrected seawater pCO_2 for PML water was 224 ± 27 μatm atm with little seasonal change (see auxiliary materials Figure S2). In order to account for seasonal temperature change in the PML, seawater pCO_2 was corrected to a constant temperature of 0°C using the thermodynamic equation of 4.21% change in pCO_2 per °C change [*Takahashi et al.*, 1993]. Previous surveys along the sea-ice edge of the Chukchi and Beaufort Seas (~150–160°W, 72–74°N) in September also showed low seawater pCO_2 conditions of ~160–280 μatm [*Pipko et al.*, 2002; *Murata and Takizawa*, 2003]. Taken together, these observations suggest that the sea-ice covered surface waters of the Arctic Ocean (specifically the Canada Basin) remain perennially undersaturated with respect to CO_2 in the atmosphere [*Bates*, 2006].

[13] Our observations indicate that the average ΔpCO_2 of PML waters in the Canada Basin was ~150 ± 25 μatm. Although surface waters of the Canada Basin (and Arctic Ocean) were highly undersaturated and potentially a sink for atmospheric CO_2, the presence of sea-ice suppressed the rates of *air*-to-*sea* CO_2 flux. For example, in the spring of 2002 and 2004, sea-ice cover at stations in the Canada Basin ranged from ~95% to 100%, and *air*-to-*sea* CO_2 fluxes was generally very low (<3 mmoles CO_2 m^{-2} d^{-1}; Figure 2b) compared to the seasonally sea-ice free regions of the adjacent Chukchi and Beaufort Sea shelves (Figure 1). During summertime, sea-ice cover in the PML of the Canada Basin ranged from ~0% to 90% and average *air*-to-*sea* CO_2 fluxes were higher at ~55 mmoles CO_2 m^{-2} d^{-1}.

3.3. What Maintains CO_2 Undersaturation in the Canada Basin?

[14] Our observations of seawater pCO_2 in Canada Basin indicate that the Arctic Ocean has a strong potential to be a sink for atmospheric CO_2, although, at present, suppressed by perennial sea-ice cover. Since sea-ice extent and volume in the Arctic Ocean has decreased over the last few decades [*Cavalieri et al.*, 2003; *Rothrock and Zhang*, 2005; *Stroeve et al.*, 2005; *Wang et al.*, 2005; *Overland and Wang*, 2005], and expected to decrease further over the next few decades, it is important to determine what maintains the seawater undersaturation of CO_2 in the Arctic Ocean and if the CO_2 sink in the Arctic Ocean region will change in the future.

[15] A major determinant of seawater pCO_2 undersaturation in the Arctic Ocean is the physical and biogeochemical preconditioning upstream on the Chukchi and Beaufort Sea shelves. During summertime, the shelf to basin outflow of water from the Chukchi Sea shelf horizontally exports into the Canada Basin warm, biogeochemically modified surface water with a DIC deficit, and low seawater pCO_2 (<200 μatm) content due to intense shelf primary and net community production (NCP) [*Hill and Cota*, 2005; *Bates et al.*, 2005b]. Another contributor is a solubility effect on seawater CO_2, since surface layer waters cool by ~0.5–1.5°C during northward transit

across the Chukchi Sea shelf, decreasing seawater pCO_2 at rate of ~4.21% per °C [*Takahashi et al.*, 1993]. At other times during the year, the continual northward flow of water also exports waters undersaturated with respect to the atmospheric CO_2 (seawater pCO_2 at 300–350 μatm) [*Bates*, 2006]. The Arctic Ocean is also significantly impacted by freshwater inputs from rivers. Recent estimates suggest that approximately 8 m of the upper 40 m of the Canada Basin surface layer has freshwater origins [*Macdonald et al.*, 2002], potentially bringing excess alkalinity into the Arctic Ocean.

[16] These physical and biogeochemical factors combine to enhance the capacity of the Arctic Ocean surface layer to absorb CO_2. The capacity of seawater to absorb CO_2 depends primarily on the buffer capacity or Revelle Factor of seawater [*Takahashi et al.*, 1993; *Sabine et al.*, 2004]. The Revelle Factor quantifies the rate of change of pCO_2 to dissolved inorganic carbon (DIC) in water masses and reflects the underlying seawater charge balance and, importantly, the ratio of DIC to alkalinity. Tropical and subtropical waters tend to have low Revelle Factors (8–10) with a greater capacity to absorb atmospheric CO_2 compared to temperate and sub-polar waters with high Revelle Factors (11–14). Recent studies [*Bates*, 2006] indicate that the surface waters of the Canada Basin have unusually low Revelle Factors (~3.5–6.5) due to the removal of DIC (while alkalinity remains unchanged) on the Chukchi Sea shelf by NCP. The addition of river alkalinity that act to decrease the DIC:TA ratio (shifting waters to lower Revelle Factors) of Arctic Ocean surface waters [*Bates*, 2006].

[17] All these factors have a strong preconditioning influence due to the relatively short residence time of Arctic Ocean surface waters. The residence time of surface layer waters of the Canada Basin has been estimated at ~15 years [*Hansell et al.*, 2004], and shorter (<10 years) in the Eurasian Basin [*Anderson and Kaltin*, 2001] due to export out of the Fram Strait. Given the volume of polar mixed layer waters in the Arctic Ocean (~2.9×10^{14} m^3), the upper 50 m of the Arctic Ocean has a turnover rate of ~10 years due to the northward (~0.8 Sv) flow of Pacific Ocean waters through Bering Strait.

3.4. Future Changes in the Arctic Ocean CO_2 Sink

[18] How will the CO_2 sink in the Arctic Ocean region potentially change in the future? Past studies [*Anderson et al.*, 1998] have estimated the annual net air-to-sea CO_2 flux over the Arctic Ocean and adjacent continental shelves to be ~24 Tg C $year^{-1}$. However, recent estimates suggest that the Chukchi Sea and entire Arctic Ocean (including the Chukchi Sea) is a larger net oceanic sink for atmospheric CO_2 at ~38 and 66 Tg C $year^{-1}$, respectively [*Bates*, 2006]. During the last couple of decades, the rate of loss of sea-ice from the central Arctic Ocean is estimated at 0.036×10^6 km $year^{-1}$ [*Cavalieri et al.*, 2003], approximately 0.6% $year^{-1}$ of the total surface area (5.8×10^6 km^2) of the Arctic Ocean. The loss of sea-ice is expected to , both reduce the % sea-ice cover in the sea-ice marginal , zone and expose undersaturated waters the Arctic Ocean (i.e. Canada and Eurasian Basins). We expect *air*-to-*sea* CO_2 flux to increase relative to decreasing sea-ice % cover generally along the regression line shown in Figure 2b. Where PML waters of the Canada Basin are exposed during seasonally sea-ice free conditions, the mean net air-to-sea CO_2 flux was ~46 ± 8 mmoles CO_2 m^{-2} d^{-1} (i.e., intercept , at 0% sea-ice cover in Figure 2b). Scaling these observations from the Canada Basin to the entire Arctic Ocean region, we compute that the average annual loss of sea-ice enhances *air*-to-*sea* CO_2 flux (i.e., ~0.55 g CO_2 m^{-2} d^{-1}; assuming sea-ice free conditions of ~100 days $year^{-1}$) into the Arctic Ocean by ~2.0 ± 0.3 Tg C $year^{-1}$, with the effect compounding over time. Here, we only consider the effects , of sea-ice loss in the Arctic Ocean and not elsewhere in the marginal sea-ice zones (e.g., Bering Sea).

[19] The extent of Arctic Ocean sea-ice has declined by ~0.9×10^6 km over the last 30 years (1972–2002) [*Cavalieri et al.*, 2003]. At present, the net Arctic Ocean sink for atmospheric CO_2 has been estimated at ~66 Tg C yr^{-1} for 2002 [*Bates*, 2006]. However, using the air-to-sea CO_2 rate change of 2.0 Tg C yr^{-1} due to sea-ice decrease, we compute that the Arctic Ocean sink for atmospheric CO_2 would have been smaller in 1972 at ~20 Tg C yr^{-1}, similar to the carbon budget estimates of *Anderson et al.* [1998]. Thus, from the late 1960's/early 1970's, the Arctic Ocean sink for atmospheric CO_2 may have tripled (from ~24 to ~66 Tg C yr^{-1}) since the onset of sea-ice decline in the Arctic Ocean [e.g., *Cavalieri et al.*, 2003; *Wang et al.*, 2005; *Overland and Wang*, 2005; *Rothrock and Zhang*, 2005; *Stroeve et al.*, 2005]. Extrapolating our results into the future suggests that the net air-to-sea CO_2 flux will increase by ~20 ± 3 Tg C (or ~28 ± 4% by the year 2012 assuming that 2002 is used as a starting point) in the next decade. If all the sea-ice of the Arctic Ocean melted (assuming surface layer pCO_2 of ~225 ± 25 μatm and similar seasonal distribution of wind in the Canada Basin as the adjacent Chukchi Sea slope), there would be a one-time uptake of ~1.2 ± 0.2 Pg C (10^{15} g C). This compares to a previous estimate of 0.28 Pg C [*Anderson and Kaltin*, 2001] who computed how much CO_2 was needed to equilibrate the PML with the atmosphere using data from the Eurasian Basin. If we use a similar approach using vertical profiles of DIC observed in the Canada Basin from 2002 to 2004, we estimate the uptake required would be ~1.86 ± 0.2 Pg C (see auxiliary materials for details). The difference in estimates presumably reflects differences in the vertical distribution of DIC between the Canada Basin and Eurasian Basin.

[20] This future projection of the Arctic Ocean CO_2 sink could be counteracted or enhanced by other mechanisms. Warming associated with melting of sea-ice in the Arctic Ocean basin should conservatively increase surface temperatures (~1.8°C to >0°C) and seawater pCO_2 by ~25–30 μatm, thereby lowering the compounding annual enhancement of net air-to-sea CO_2 flux from ice retreat by ~20% (i.e., 2.0 to 1.6 Tg C $year^{-1}$). We assume that melting of sea-ice dilutes DIC and alkalinity with negligible impact on seawater pCO_2. Freezing and brine formation might also slightly enhance the air-to-sea CO_2 flux [*Anderson et al.*, 2004]. Other studies have suggested that the melting of sea-ice might enhance primary production and the Arctic Ocean CO_2 sink [*Anderson and Kaltin*, 2001]. However, given the very low nitrate and phosphate concentrations observed in the Arctic Ocean [*Codispoti et al.*, 2005], we anticipate a minor enhancement of the vertical C flux (~0.03 Tg C $year^{-1}$) out of the surface

layer (assuming a NO_3 content of 0.2 μmoles kg^{-1} and mixed layer of 50 m).

[21] Processes that can contribute to the budget of CO_2 in the PML include vertical diffusion, primary production, downwelling, gas exchange, and lateral transport. At present, the gain of DIC (and increase in pCO_2) in the PML due to the upward diffusion of CO_2 from the underlying upper halocline layer appears to roughly balance the loss of CO_2 due to downwelling of the PML and export of organic carbon due to phytoplankton primary production (see auxiliary materials[1] for details). The upward vertical diffusional flux of CO_2 was estimated from very low vertical eddy diffusivity reported by *Wallace et al.* [1987] (2×10^{-6} m^2 s^{-1}) or calculated from SBI temperature and salinity profiles [$1.4 \pm 1.0 \times 10^{-6}$ m^2 s^{-1}]. The loss of CO_2 from the PML of the Canada Basin due to downwelling was estimated assuming a downwelling term of 0.5 m yr^{-1} [*Wallace et al.*, 1987]. Finally, the loss of CO_2 from the PML due to primary production was estimated from low estimates of annual primary production in the central Arctic Ocean ($\sim$0.6–1.3 g C m^{-2} year^{-1} [*Anderson et al.*, 2003]; 1.8 g C m^{-2} year^{-1} [*English*, 1961]; 3.6 g C m^{-2} year^{-1} [*Moran et al.*, 1997]) or estimates of net community production from the Canada Basin (e.g., 2.2–6.5 g C m^{-2} year^{-1} [*Bates et al.*, 2005a]). Any significant change in the balance of these and other processes could lead to changes in the seawater pCO_2 undersaturation and potential CO_2 sink observed in the Canada Basin (and Eurasian Basin).

[22] If there were any change in the transport of inorganic nutrients to the surface layer of the Arctic Ocean basin, this may impact the rate of air-sea CO_2 flux. At present, the inorganic nutrients of surface waters transiting northward across the shelf during the summertime ($\sim$100 days duration) are consumed by the time they reach the deep basin [*Codispoti et al.*, 2005]. During the other times of , the year ($\sim$270 days), we estimate that an outflow of surface water (with nitrate concentrations of $\sim$15 μmoles kg^{-1}) of 0.4 Sv from the Chukchi Sea shelf could enhance the rate of primary production in the Arctic Ocean by $\sim$10 Tg C year^{-1}, or $\sim$1.7 g C m^{-2} year^{-1}. This compares to the enhancement of primary production of 17 Tg C year^{-1} for water flowing into the Arctic Ocean from the Barents Sea [*Anderson and Kaltin*, 2001].

[23] Over the next 10 year period (e.g., 2002 to 2012), the surface layer could potentially absorb an extra $\sim$100 Tg C (i.e., $\sim$2.0 Tg C year^{-1} compounded each year) due to continued reduction of Arctic Ocean sea-ice. Since the average polar mixed layer depth is $\sim$50 m, we would expect seawater DIC and pCO_2 contents to increase by $\sim$24 μmoles kg^{-1} and $\sim$30 μatm, respectively over a 10 year period. The increase in seawater pCO_2 would thus reduce the ocean CO_2 sink enhancement by $\sim$20% (this assumes that the rate of sea-ice loss continues at previous rates, and that other potential factors such as Arctic Ocean basin primary production and export, Chukchi Shelf transport of freshwater, nutrients, DIC, organic matter, and alkalinity remain unchanged). But, given the relatively short residence times ($\sim$15 years) of Arctic Ocean surface water, continued inputs of shelf water would dilute the DIC accumulation due to enhanced air-to-sea CO_2 flux. Furthermore, if the freshwater input of alkalinity to the Arctic Ocean by rivers increases in the future, this contribution may sustain the low Revelle Factor and the high capacity of surface water to absorb CO_2. Considering all factors, we anticipate that the oceanic sink of CO_2 in Arctic Ocean will continue to increase for a period related to the future duration of the sea-ice loss from the Arctic Ocean.

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