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Seasonal response of air-water CO_2 exchange along the land-ocean aquatic continuum of the North East American coast

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Abstract

This regional study quantifies the CO_2 exchange at the air–water interface along the land-ocean aquatic continuum (LOAC) of the North East American coast, from streams to the shelf break. Our analysis explicitly accounts for spatial and seasonal variability in the CO_2 fluxes. The yearly integrated budget reveals the gradual change in the intensity of the CO_2 exchange at the air–water interface, from a strong source towards the atmosphere in streams and rivers ($3.0 \pm 0.5\,\mathrm{Tg\,C\,yr^{-1}}$) and estuaries ($0.8 \pm 0.5\,\mathrm{Tg\,C\,yr^{-1}}$) to a net sink in continental shelf waters ($-1.7 \pm 0.3\,\mathrm{Tg\,C\,yr^{-1}}$). Significant differences in flux intensity and their seasonal response to climate variations is observed between the North and South sections of the study area, both in rivers and coastal waters. Ice cover, snow melt and estuarine surface area are identified as important control factors of the observed spatio-temporal variability in CO_2 exchange along the LOAC.

1 Introduction

Over the past decade, several syntheses have highlighted the significant contribution of the Land–Ocean Aquatic Continuum (LOAC) to the global atmospheric CO_2 budget (Cole et al., 2007; Battin et al., 2009; Mackenzie et al., 2012; Bauer et al., 2013; Ciais et al., 2013; Raymond et al., 2013; Regnier et al., 2013). In a recent review, Regnier et al. (2013) proposed that inland waters (streams, rivers and lakes) and estuaries outgas 1.1 and $0.25\,\mathrm{Pg\,C\,yr^{-1}}$, respectively, while continental shelf seas take up $0.2\,\mathrm{Pg\,C\,yr^{-1}}$. However, CO_2 data are too sparse and unevenly distributed to provide global coverage and large uncertainties remain associated to these estimates. The inland water outgassing could for instance reach $2.1\,\mathrm{Pg\,C\,yr^{-1}}$ with 86 % coming from streams and rivers (Raymond et al., 2013), a value which is about twice that reported in Regnier et al. (2013) and in the 5th assessment report of the IPCC (Ciais et al., 2013). The most recent global budgets for the estuarine CO_2 source and the continental shelf CO_2 sink also reveal significant discrepancies, both falling within the $0.15–0.4\,\mathrm{Pg\,C\,yr^{-1}}$

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range (Laruelle et al., 2010; Cai, 2011; Bauer et al., 2013; Dai et al., 2013; Laruelle et al., 2013). None of these estimates, however, fully resolves the seasonality in CO_2 fluxes because temporal coverage of the global data is insufficient. Complex seasonal dynamics of CO_2 exchanges between the atmosphere and individual components of the LOAC have been reported in previous studies which have highlighted the potential importance of the intra-annual variability for local and regional CO_2 budgets (e.g. Kempe, 1982; Frankignoulle et al., 1998; Jones and Mulholland, 1998; Degrandpré et al., 2002; Thomas and Schneider, 1999; Wallin et al., 2011; Rawlins et al., 2014). Here, we extend the analysis to the sub-continental scale, and present the spatial and seasonal variability of CO_2 fluxes at the air–water interface (FCO_2) for the entire North East American LOAC, from streams to the shelf break. This region of unprecedented data coverage allows us producing, for the first time, empirically-derived monthly maps of CO_2 exchange at 0.25° resolution. Our results allow investigating the seasonal CO_2 dynamics across the inter-connected systems of the LOAC and elucidating their response to contrasting intra-annual changes in climate conditions.

2 Methods

COSCAT 827 is located along the Atlantic coast of the Northern US and Southern Canada and extends from the Albemarie Sound in the South to the Eastern tip of Nova Scotia in the North. COSCATs are homogenous geographical units that divide the global coastline into homogeneous segments according to lithological, morphological, climatic and hydrological properties (Meybeck et al., 2006; Laruelle et al., 2013). The area corresponding to COSCAT 827 comprises 447×10^3 km^2 of watersheds and 357×10^3 km^2 of coastal waters, amongst which 15×10^3 km^2 of estuaries. It is one of the best monitored region in the world with several regularly surveyed rivers (Hudson, Susquehanna, York, Connecticut) and some of the most extensively studied coastal waters (Degrandpré et al., 2002; Chavez et al., 2007; Fennel et al., 2008; Fennel and Wilkin, 2009; Previdi et al., 2009; Fennel, 2010; Shadwick et al., 2010, 2011; Sig-

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norini et al., 2013). For the purpose of this study, the area was divided in a North and a South section (Fig. 1). The boundary is set on land, to delineate the regions subject to seasonal ice freeze and snowfalls from those that are not (Armstrong and Brodzik, 2001). This delineation attributes 96 % of the estuarine surface area to the South section due, for the most part, to the contribution of Chesapeake Bay which accounts for about two thirds of the estuarine area. The delineation extends further into the coastal waters in such a way that the Scotian Shelf and the Gulf of Maine correspond to the North section and the Mid-Atlantic Bight and Georges Banks to the South section. The riverine data are calculated from pH and alkalinity measurements (previously used in Lauerwald et al., 2013) while continental shelf values are calculated from the SOCAT 2.0 database which contains quality controlled direct pCO_2 measurements (http://www.socat.info/, Bakker et al., 2014).

2.1 Rivers

CO_2 evasion from rivers (FCO_2) was calculated monthly per 15s grid cell (resolution of the hydrological routing scheme Hydrosheds 15s, Lehner et al., 2008) from estimates of the effective stream/river surface area A_{eff}, gas exchange velocity k, and water-atmosphere CO_2 concentration gradient $\Delta[CO_2]$:

$$FCO_2 = A_{eff} \times k \times \Delta[CO_2] \quad (1)$$

The calculation of A_{eff} first requires estimation of the total stream/river surface area, A. The latter was calculated from the linear stream network derived from the Hydrosheds 15s routing scheme using a minimum threshold on the catchment area of $10\,km^2$ and estimates of stream width derived from the annual mean discharge Q_{ann} using the equations of Raymond et al. (2012, 2013). A values were not calculated for each individual month, as the discharge-stream width relations only hold true for Q_{ann} (Raymond et al., 2013). Q_{ann} was obtained using Hydrosheds 15s to route the gridded data of average annual runoff from the UNH/GRDC composites (Fekete et al., 2002). For each 15s raster cell covered by lake and reservoir areas as represented in the

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global lake and wetland data base of Lehner and Döll (2004), A was set to 0 km^2. A_{eff} was then derived from A to account for seasonal stream drying and ice cover inhibiting FCO_2. Seasonal stream drying was assumed for each 15s cell and month when the monthly average discharge Q_{month} is 0 $m^3\,s^{-1}$. Values of Q_{month} were calculated similarly to that of Q_{ann} using the gridded data of average monthly runoff from the UNH/GRDC composites (Fekete et al., 2002). Ice cover was assumed for each 15s cell and month when the mean air temperature (T_{air}), derived from the worldclim data set of Hijmans et al. (2005), is below −4.8 °C. In case of ice cover and/or stream drying, A_{eff} is set to 0 m^2. Otherwise A_{eff} equals A.

Values of k were first calculated as standardized values for CO_2 at a water temperature (T_{water}) of 20 °C (k_{600}), from stream channel slope CS and estimates of flowing velocity V. Using the Strahler order (Strahler, 1952) to perform the segmentation of the stream network, CS was calculated for each segment by dividing the change in its altitude by its length. Information on altitude was derived from the Hydrosheds elevation model. V was calculated from Q_{ann} based on the equations of Raymond et al. (2012, 2013). Similarly to the stream width, the V-Q relations only hold true for Q_{ann} (Raymond et al., 2013), and this is why only annually average values for V and k_{600} could be calculated. The k value for each month was calculated from k_{600} and an estimate of the in-situ air temperature $T_{water,}$ based on the mean monthly air temperature derived from the worldclim data set of Hijmans et al. (2005).

Values of $\Delta(CO_2)$ were derived from monitoring data with calculated $pCO_{2\,river}$ (12 300 water samples, from 161 locations, Lauerwald et al., 2013) and assumed $pCO_{2\,atmosphere}$ of 390 µatm. The pCO_2 values were converted into concentrations, $[CO_2]$, using Henry's constant (Henry, 1803) for each sample at its observed temperature T_{water} using the equation of Telmer and Veizer (1999). In order to minimize the influence of extreme values, the results were aggregated to median values per sampling location and month for which at least three values were available. These median values per sampling location and month were then used to calculate maps of $\Delta[CO_2]$ at a 15s resolution using an inverse distance weighted interpolation. To account for downstream decreases in $pCO_{2\,river}$, which are often reported in the literature (Finlay, 2003; Teodoru et al., 2009; Butman and Raymond, 2011), the interpolation was applied separately to three different classes of streams and rivers defined by Q_{ann}, for which sufficiently large subsets of sampling locations could be retained: (1) $Q_{ann} < 10\,m^3\,s^{-1}$ ($n = 76$), (2) $10\,m^3\,s^{-1} \leq Q_{ann} < 100\,m^3\,s^{-1}$ ($n = 47$), and (3) $Q_{ann} \geq 100\,m^3\,s^{-1}$ ($n = 38$). The three maps of $\Delta[CO_2]$ per month were then recombined according to the spatial distribution of Q_{ann} values. The FCO_2 values were first calculated using Eq. (1) at the high spatial resolution of 15s for each month. The results were then aggregated to a 0.25° resolution and three-month period and reported as relative to the terrestrial surface area per raster cell including inland waters. The difference between the FCO_2 s calculated using the equations of Raymond et al. (2012) and Raymond et al. (2013) was used as an estimate of the uncertainty of the mean yearly FCO_2.

2.2 Estuaries

The yearly averaged CO_2 exchange at the air–water interface was obtained from local estimations of emission rates in seven estuaries located within the study area (Raymond et al., 1997, 2000; Raymond and Hopkinson, 2003; Hunt et al., 2010). The limited number of observation does not allow resolving the seasonality in CO_2 emissions. The yearly-average local CO_2 emission rates range from $1.1\,molC\,m^{-2}\,yr^{-1}$ in the Parker River to $9.6\,molC\,m^{-2}\,yr^{-1}$ in the Hudson River estuary, for a mean value of $4.2\,molC\,m^{-2}\,yr^{-1}$ for the seven systems. This value was then multiplied by the estuarine surface areas extracted from the SRTM water body data set (NASA/NGA, 2003), to estimate the bulk outgassing for the North and South sections of COSCAT 827. Similar approaches have been used in the past to produce global estuarine CO_2 budgets (Borges et al., 2005; Laruelle et al., 2010; Cai, 2011; Chen et al., 2013; Laruelle et al., 2013). The standard deviation calculated for the emission rates of all local studies was used as an estimate of the uncertainty of the regional estuarine FCO_2.

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2.3 Continental shelf waters

Monthly CO_2 exchange rates at the air–water interface were calculated in continental shelf waters using 274 291 pCO_2 measurements extracted from the SOCAT 2.0 database (Baker et al., 2014). For each measurement, an instantaneous local CO_2 exchange rate with the atmosphere was calculated using Wanninkhof's equation (Wanninkhof, 1992) which is a function of a transfer coefficient (k), dependent on the square of the wind speed above sea surface, the apparent solubility of CO_2 in water (K_0'), which depends on surface water temperature and salinity, and the gradient of pCO_2 at the air–water interface (ΔpCO_2).

$$FCO_2 = A_s \times k \times K_0' \times \Delta pCO_2 \tag{2}$$

The parameterization used for k is that of Wanninkhof et al. (2013) and all the data necessary for the calculations are available in SOCAT 2.0 except for wind speed, which was extracted from the CCMP database (Altas et al., 2011). The resulting CO_2 exchange rates were then averaged per month for each 0.25° cell in which data were available. Monthly FCO_2 for the North and South sections were then extrapolated using the water surface area and weighted rate for each cell, multiplied by the total surface area A_s of the corresponding section. To refine further the budget, a similar procedure was also applied to 5 depth segments (S1 to S5) corresponding to 0–20 m, 20–50 m, 50–80 m, 80–120 m and 120–150 m, respectively, and their respective surface areas were extracted from a high resolution bathymetric files (Laruelle et al., 2013). The choice of slightly different methodologies for FCO_2 calculations in rivers and continental shelf waters stems from the better data coverage in the continental shelf, which allows capturing the spatial heterogeneity within the region without using interpolation techniques. The standard deviation calculated for all the grid cells of the integration domain was used as an estimate of the uncertainty of the yearly FCO_2s.

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3 Results and discussion

Figure 2 shows the spatial distribution of FCO_2 along the LOAC integrated per season. Throughout the year, river waters are a strong source of CO_2 for the atmosphere. Significant differences in the intensity of the CO_2 exchange at the air–water interface can nevertheless be observed between the North and South sections, both in time and space. During winter, there is nearly no CO_2 evasion from rivers in the North due to ice coverage and stream drying. Over the same period, the CO_2 emissions from the South section range from 0 to $5\,\mathrm{g\,C\,m^{-2}\,season^{-1}}$. During spring, the pattern is reversed and northern rivers exhibit higher outgassing rates than in the South with maximum emissions rates of $> 10\,\mathrm{g\,C\,m^{-2}\,season^{-1}}$. This trend is maintained throughout summer while during fall, the entire COSCAT displays similar emission rates without clear latitudinal signal. Continental shelf waters display a very different spatial and seasonal pattern than that of rivers. During winter, the North section is predominantly a mild CO_2 sink, with rates comprised between +2 and $-5\,\mathrm{g\,C\,m^{-2}\,season^{-1}}$, which intensifies significantly in the South section (-2 to $> -10\,\mathrm{g\,C\,m^{-2}\,season^{-1}}$). During spring, an opposite trend is observed with a quasi-neutral CO_2 uptake in the South and a strong uptake in the North, especially on the Scotian shelves. The entire COSCAT becomes a net CO_2 source in summer with emission rates as high as $5\,\mathrm{g\,C\,m^{-2}\,season^{-1}}$ in the Mid-Atlantic Bight. During fall, the Gulf of Maine and Georges Banks remain CO_2 sources while the Scotian shelves and the Mid-Atlantic Bight become again regions of net CO_2 uptake.

The monthly integrated FCO_2 for the North and South sections provides further evidence of the contrasting seasonal dynamics for the two areas (Fig. 3a and b). In the North section, CO_2 evasion from rivers is almost zero in January and February, rises to a maximum value of $0.26 \pm 0.05\,\mathrm{Tg\,C\,month^{-1}}$ in May, and then progressively decreases until the end of the year. These low winter values are explained by the ice cover inhibiting the gas exchange with the atmosphere. The steep increase and FCO_2 maximum in spring can be related to the flushing of water from the thaw-

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ing top-soils, which is rich in DOC and CO_2, combined to increasing in-stream respiration rates induced by warmer water temperatures (Jones and Mulholland, 1998; Striegl et al., 2012). Compared to rivers, the continental shelf in the North section presents a close mirror behavior from winter through spring, with a mild carbon uptake rate in January and February (-0.04 ± 0.25 Tg C month^{-1}) followed by a maximum uptake rate in April (-0.50 ± 0.20 Tg C month^{-1}). This CO_2 uptake in spring has been attributed to photosynthesis associated to the seasonal phytoplankton bloom (Shadwick et al., 2010). Continental shelf waters behave quasi neutral during summer ($< 0.05 \pm 0.09$ Tg C month^{-1}) and emit CO_2 at a high rate in November and December ($> 0.15 \pm 0.21$ Tg C month^{-1}). Overall, the rivers of the North section emit 1.31 ± 0.24 Tg C yr^{-1} while the continental shelf waters take up 0.47 ± 0.17 Tg C yr^{-1}. The very limited estuarine surface area (0.5 10^3 km^2) only yields an annual outgasing of 0.03 ± 0.02 Tg C yr^{-1}. The shelf sink calculated for the region differs from that of Shadwick et al. (2011) which reports a source for the Scotian Shelves, in contrast to the current estimate. Our seasonally resolved budget is however in line with the -0.6 Tg C yr^{-1} sink calculated by Signorini et al. (2013) using a 8 years dataset as well as with the simulations of Fennel and Wilkin (2009) which also predict sinks of -0.7 Tg C yr^{-1} and -0.6 Tg C yr^{-1} for 2004 and 2005, respectively. No similar analysis was so far performed for inland waters.

In the South section of the COSCAT, the warmer winter temperature leads to the absence of ice cover (Armstrong and Brodzik, 2001). Our calculations predict that the riverine surface area remains stable over time, favoring a relatively constant outgassing comprised between 0.1 and 0.2 Tg C month^{-1} throughout the year, adding up to a yearly source of 1.69 ± 0.31 Tg C yr^{-1}. Estuaries emit 0.73 ± 0.45 Tg C yr^{-1}, because of their large surface area of 14.5 10^3 km^2, one order of magnitude larger than that of rivers. The continental shelf CO_2 sink is strongest in January (-0.47 ± 0.30 Tg C month^{-1}) and decreases until June, when a period of moderate CO_2 emission begins (max of 0.13 ± 0.08 Tg C month^{-1} in August) and lasts until October. Finally, November and December are characterized by mild CO_2 sinks. Such seasonal signal, following that of water temperature, is consistent with the hypothesis of a CO_2 exchange in the South section regulated by variations in gas solubility, as suggested by Degrandpré et al. (2002) for the Mid-Atlantic Bight.

The analysis of the intensity of the river CO_2 outgassing reveals that the smallest streams ($Q < 1\,m^3\,s^{-1}$, Q1 in Table 1) display the highest emission rates per unit surface area, with values ranging from 1961 g C $m^{-2}\,yr^{-1}$ in the South section to 2893 g C $m^{-2}\,yr^{-1}$ in the North section. These values gradually decrease with increasing river discharge to 729 g C $m^{-2}\,yr^{-1}$ in the South section and 891 g C $m^{-2}\,yr^{-1}$ in the North section for $Q > 100\,m^3\,s^{-1}$ (Q4, Table 1). The emission rates for this latter class of rivers are consistent with the median emission rate of 720 g C $m^{-2}\,yr^{-1}$ proposed by Aufdenkampe et al. (2011) for temperate rivers with widths larger than 60–100 m. Aufdenkampe et al. (2011) also report a median emission rates of 2600 g C $m^{-2}\,yr^{-1}$ for the smaller streams and rivers, which falls on the high end of the range calculated for Q1 in the present study. The surface area of the river network is relatively evenly distributed amongst the four discharges classes of rivers (Table 1). Yet, river sections for which $Q < 10\,m^3\,s^{-1}$ (Q1 + Q2) contribute to 65 % of the total CO_2 outgassing although they only represent 51 % of the surface area. This result therefore highlights that streams and small rivers are characterized by the highest surface-area specific emission rates. On the continental shelf, the shallowest depth interval is a CO_2 source while all other depth intervals are CO_2 sinks (Table 1). The magnitude of the air–sea exchange for each segment is comprised between the values calculated for estuaries (50 g C $m^{-2}\,yr^{-1}$) and the nearby open ocean (~ 20 g C $m^{-2}\,yr^{-1}$, according to Takahashi et al., 2009). This trend along a depth transect, suggesting a more pronounced continental influence on near-shore waters was already discussed in the regional analysis of Chavez et al. (2007) and by Jiang et al. (2013) specifically for the South Atlantic Bight. Modeling studies over a larger domain including the upper slope of the continental shelf also suggest that the coastal waters of the North East US are not a more intense CO_2 sink than the neighboring open ocean (Fennel and Wilkin, 2009; Fennel, 2010). Our analysis further suggests that the continental influence is more pronounced in the North section. Here, the shallowest waters (S1) are strong net sources of CO_2 while the intensity of the CO_2 sink for the other depth intervals gradually decreases, but only to a maximum value of $-4\,g\,C\,m^{-2}\,yr^{-1}$ for S5. This value is about 3 times smaller than in the South section ($-12\,g\,C\,m^{-2}\,yr^{-1}$).

Annually, river and estuarine waters of the entire COSCAT 827 outgas $3.0 \pm 0.5\,Tg\,C\,yr^{-1}$ and $0.8 \pm 0.5\,Tg\,C\,yr^{-1}$, respectively, while continental shelf waters take up $1.7 \pm 0.3\,Tg\,C\,yr^{-1}$ (Fig. 3c). The total riverine carbon load exported from rivers to estuaries for the same area has been estimated to $4.65\,Tg\,C\,yr^{-1}$, 45 % as dissolved and particulate organic carbon ($2.10\,Tg\,C\,yr^{-1}$, Mayorga et al., 2010) and 55 % as dissolved inorganic carbon ($2.55\,Tg\,C\,yr^{-1}$, Hartmann et al., 2009). Estimates of the total amount of terrestrial carbon transferred to the riverine network are not available but the sum of the river export and the outgassing, which ignores the contribution of carbon burial and lateral exchange with wetlands, provides a lower bound estimate of $7.65\,Tg\,C\,yr^{-1}$. Under this hypothesis, ~ 40 % of the terrestrial carbon exported to rivers is emitted to the atmosphere before reaching estuaries. In spite of higher emission rates per unit surface area in the North (Table 1), the overall efficiency of the riverine carbon filter is essentially the same in the two sections (40 % and 38 % outgassing for the North and the South, respectively). On the shelf, however, the South section exhibit a significantly more intense CO_2 sink ($-1.25 \pm 0.2\,Tg\,C\,yr^{-1}$) than in the North ($-0.47 \pm 0.2\,Tg\,C\,yr^{-1}$). A possible reason for this difference can be found in the contribution of the estuarine carbon filter. In the South, where 96 % of the estuarine surface area is located, these systems contribute to an outgassing of $0.73\,Tg\,C\,yr^{-1}$ while in the North, their influence is negligible. Cole and Caraco (2001) estimated that 28 % of the DOC entering the relatively short Hudson River estuary is respired in-situ before reaching the continental shelf and it is thus likely that the estuarine outgassing in the South section is fueled by the respiration of the organic carbon loads from rivers. In contrast, the absence of estuaries in the North favors the direct export of terrestrial organic carbon onto continental shelf waters where it can be buried and decomposed. The respiration of terrestrial organic carbon could therefore explain why the strength of the shelf CO_2

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sink is weaker in this portion of the domain. This view is further substantiated by the similar cumulated estuarine and continental shelf FCO_2 fluxes in both sections (Fig. 3a and b). Naturally, other environmental and physical factors, such as, for example, local coastal currents (Wang et al., 2013), temperature changes and phytoplankton growth could also contribute to the difference in CO_2 uptake intensity between both sections. Additionally, modeling studies evidenced the potential influence of sediment denitrification on water $p CO_2$ through the removal of fixed nitrogen in the water column and consequent inhibition of primary production (Fennel et al., 2008; Fennel, 2010). This removal was estimated to be of similar magnitude as the lateral nitrogen loads, except for estuaries of the MAB region (Fennel, 2010). It can nonetheless be suggested that the estuarine carbon filter in the South section of COSCAT 827 is an important control factor of the CO_2 sink in the Mid-Atlantic Bight, which is stronger than in any other area along the entire Atlantic coast of the US (Signorini et al., 2013).

4 Conclusions

Our spatially and seasonally resolved budget analysis captures the main characteristics of the air-water CO_2 exchange along the LOAC of COSCAT 827. It evidences the contrasting dynamics of the North and South section of the study area and an overall gradual shift from a strong source in small streams oversaturated in CO_2 towards a net sink in continental shelf waters. Our study also reveals the role of ice and snow cover as an important controlling factor of the seasonal dynamics of CO_2 outgassing in streams and rivers. Additionally, the incorporation of the LOAC as a whole supports an integrated analysis that highlights the contribution of estuaries as filters of the terrestrial carbon inputs and their influence on the continental shelf carbon uptake.

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Table 1. Surface areas, CO_2 exchange rate with the atmosphere and surface integrated FCO_2 for the North and South sections of COSCAT 827, subdivided by river discharge classes and continental shelf water depth intervals.

	North Surface Area 10^3 km^2	Rate g C m^{-2} yr^{-1}	FCO_2 10^9 g C yr^{-1}	South Surface Area 10^3 km^2	Rate g C m^{-2} yr^{-1}	FCO_2 10^9 g C yr^{-1}	Total Surface Area 10^3 km^2	Rate g C m^{-2} yr^{-1}	FCO_2 10^9 g C yr^{-1}
Rivers									
Q1 ($Q < 1$ m s^{-1})	0.14	2893 ± 521	391 ± 70	0.27	1961 ± 353	532 ± 96	0.41	2271 ± 409	924 ± 166
Q2 (1 m s^{-1} $< Q <$ 10 m s^{-1})	0.21	2538 ± 457	525 ± 95	0.32	1570 ± 283	506 ± 91	0.53	1948 ± 351	1032 ± 186
Q3 (10 m s^{-1} $< Q <$ 100 m s^{-1})	0.16	1476 ± 267	237 ± 43	0.30	1307 ± 235	392 ± 71	0.46	1366 ± 246	629 ± 113
Q4 (100 m s^{-1} $< Q$)	0.17	891 ± 160	152 ± 27	0.36	729 ± 131	261 ± 47	0.52	781 ± 141	412 ± 74
Sub-total	0.67	1939 ± 349	1305 ± 235	1.25	1351 ± 243	1692 ± 305	1.92	1557 ± 280	2997 ± 539
Estuaries	0.53	50 ± 31	27 ± 19	14.51	50 ± 31	731 ± 453	15.04	50 ± 31	758 ± 469
Shelf									
S1 (depth < 20 m)	11.21	5 ± 1	53 ± 19	24.28	−3 ± 1	−79 ± 11	35.49	−1 ± 1	−27 ± 5
S2 (20 m < depth < 50 m)	26.25	−1 ± 1	−35 ± 12	63.88	−8 ± 1	−521 ± 70	90.13	−6 ± 1	−556 ± 108
S3 (50 m < depth < 80 m)	39.28	−3 ± 1	−128 ± 45	48.63	−7 ± 1	−359 ± 126	87.91	−6 ± 1	−488 ± 95
S4 (80 m < depth < 120 m)	60.69	−3 ± 1	−209 ± 73	25.18	−8 ± 1	−199 ± 27	85.87	−5 ± 1	−409 ± 80
S5 (120 m < depth < 150 m)	34.73	−4 ± 1	−151 ± 18	7.63	−12 ± 1	−91 ± 12	42.36	−6 ± 1	−242 ± 47
Sub-total	172.17	−3 ± 1	−472 ± 166	169.59	−7 ± 1	−1250 ± 169	341.77	−5 ± 1	−1722 ± 335

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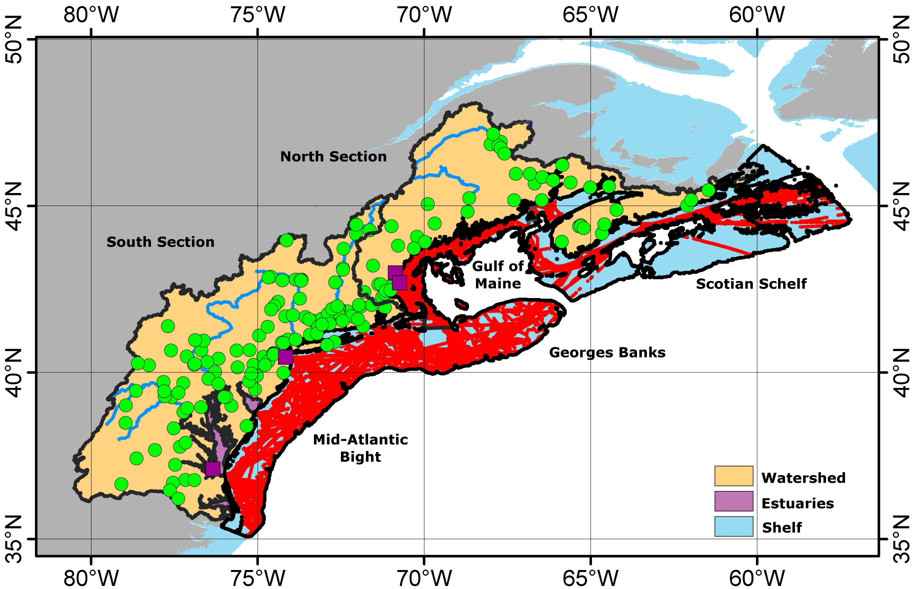

Figure 1. Geographic limits of the study area with the location of the riverine (Glorich database, in green; Lauerwald et al., 2013) and continental shelf waters data used for our calculations (SOCAT 2.0 database, in red; Bakker et al., 2014). The location of the estuarine studies used is indicated by purple squares.

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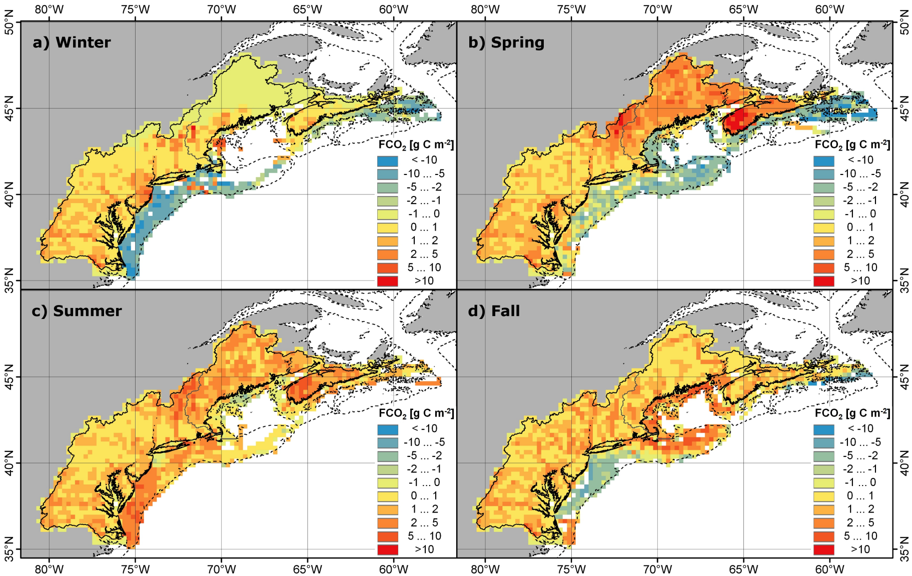

Figure 2. Spatial distribution of the CO_2 exchange with the atmosphere in rivers and continental shelf waters aggregated by seasons. The fluxes are net FCO_2 rates averaged over the surface area of each 0.25° cells and a period of 3 months. Positive values correspond to fluxes towards the atmosphere. Winter is defined as January, February and March, Spring as April, May and June and so forth.

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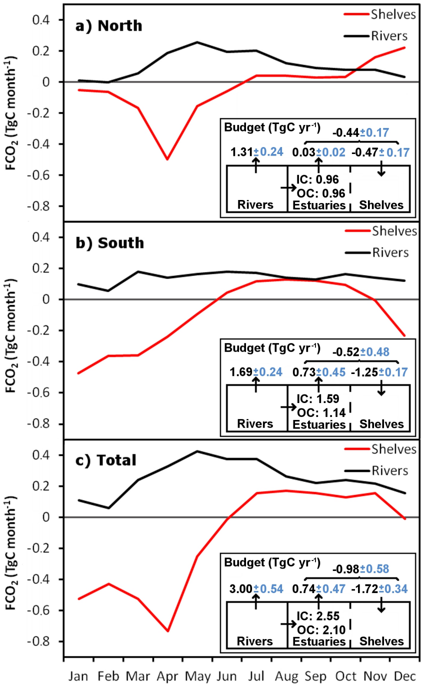

Figure 3. Areal-integrated monthly air-water CO_2 flux for rivers and the continental shelf waters in the North section **(a)**, South section **(b)**, and entire study area **(c)**. Positive values correspond to fluxes towards the atmosphere. The boxes inside each panel correspond to the annual carbon budgets for the region including the lateral carbon fluxes at the river–estuary interface, as inorganic (IC) and organic carbon (OC). The values in blue represent the uncertainties of the annual fluxes.

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