Winter weather controls net influx of atmospheric CO₂ on the north-west European shelf

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Shelf seas play an important role in the global carbon cycle, absorbing atmospheric carbon dioxide (CO₂) and exporting carbon (C) to the open ocean and sediments. The magnitude of these processes is poorly constrained, because observations are typically interpolated over multiple years. Here, we used 298,500 observations of CO₂ fugacity (fCO₂) from a single year (2015), to estimate the net influx of atmospheric CO₂ as 26.2 ± 4.7 Tg C yr⁻¹ over the open NW European shelf. CO₂ influx from the atmosphere was dominated by influx during winter as a consequence of high winds, despite a smaller, thermally-driven, air-sea fCO₂ gradient compared to the larger, biologically-driven summer gradient. In order to understand this climate regulation service, we constructed a carbon-budget supplemented by data from the literature, where the NW European shelf is treated as a box with carbon entering and leaving the box. This budget showed that net C-burial was a small sink of 1.3 ± 3.1 Tg C yr⁻¹, while CO₂ efflux from estuaries to the atmosphere, removed the majority of river C-inputs. In contrast, the input from the Baltic Sea likely contributes to net export via the continental shelf pump and advection (34.4 ± 6.0 Tg C yr⁻¹).

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Temperate continental shelf seas (<200 m depth) occupy ~7% of the ocean surface area, yet contribute disproportionately to net marine drawdown of atmospheric carbon dioxide \( \text{(CO}_2 \text{)} \), accounting for 10–20% of the global total net uptake\(^1,\text{2–3}. \) Century-scale storage of this \( \text{CO}_2 \) in sediments and seawater results in a substantial climate regulation service, but has also caused ocean acidification \( \text{(pCO}_2 \text{)} \), in which \( \text{pH} \) is projected to continue to decrease as atmospheric \( \text{CO}_2 \) increases\(^4. \) Net atmospheric \( \text{CO}_2 \) drawdown in temperate shelf seas is mediated by physical processes (dissolution of \( \text{CO}_2 \)) and nutrient-limited net community production \( \text{(NCP; photosynthesis minus mineralization).} \) Predictions of the future drawdown of atmospheric \( \text{CO}_2 \) are highly contradictory. On the one hand, increasing \( \text{CO}_2 \) in the atmosphere is expected to enhance the air to sea flux of \( \text{CO}_2 \), simply by driving the respective concentration gradient. However, the inverse dependence of \( \text{CO}_2 \) solubility on temperature\(^5. \) may limit this flux as sea surface temperature \( \text{(SST)} \) increases. In contrast with observations showing a 35-year increase in shelf sea \( \text{CO}_2 \) drawdown over the NW European shelf\(^6. \) models predict a future decrease as a result of reduced nutrient supply and net biological \( \text{CO}_2 \) uptake\(^7,8. \) Other models suggest that saturation of the ‘continental shelf \\
\text{pump}’ \( \text{(CSP; the seasonal C export into intermediate waters of the open ocean)} \) may limit further drawdown from the atmosphere\(^9. \)

Net air to sea \( \text{CO}_2 \) flux may be calculated from ship-board observations of \( \text{CO}_2 \) fugacity \( \text{(fCO}_2 \text{)} \) (the partial pressure of \( \text{CO}_2 \) corrected for its non-ideal-gas behaviour). Here, we evaluate the air to sea \( \text{CO}_2 \) flux for the NW European shelf based on the most geographically-extensive, single-year, \( \text{fCO}_2 \) dataset to date. Compared to previous evaluations (based on regional data extrapolation and combination of multiannual observations), this unique dataset provides a detailed snapshot of the air to sea \( \text{CO}_2 \) flux.

The NW European shelf is hydrologically linked to the surrounding land-mass where temperature, precipitation and river-runoff are projected to increase over the next 50 years\(^10,11. \) Policies regulating the application of fertilisers and wastewater treatment\(^13–16. \) have reduced the nutrient loading on land and river-borne inputs to the shelf and may thereby alter the biological component of the air to sea \( \text{CO}_2 \) flux. Similarly, changes in riverine dissolved organic C (DOC) inputs have the potential to modify shelf C-budgets as DOC is mineralized in shelf seas and returned to the atmosphere as \( \text{CO}_2 \)\(^17. \) The recent trend of decreasing DOC input from major UK rivers (since 2000) has been attributed to implementation of the urban sewage treatment directive\(^18. \) This followed an earlier trend of increasing DOC flux from rivers across the northern hemisphere which was attributed to soil recovery from acid rain after implementation of sulphur emission policies\(^17. \) A comprehensive, observation-based analysis of the relative magnitude of C-fluxes on the NW European shelf which would allow us to constrain the potential impact of these feedbacks is currently missing. In order to address this, we have constructed a balanced budget based on observed C-fluxes on the NW European shelf (including air to sea \( \text{CO}_2 \) flux, riverine and Baltic Sea inputs, inorganic accumulation, C-burial and export via the CSP and advection). This enables an investigation of the factors controlling the air to sea flux, in turn providing guidance on the impact of land management practices.

Results

Air to sea flux in 2015. Monthly average fluxes showed considerable spatial and seasonal variability over 2015 (Fig. 1), reflecting changes in phytoplankton abundance, \( \text{CO}_2 \) solubility and wind speed which generates turbulence at the air-sea interface promoting flux (Fig. 1 and Supplementary Figs. 1–3). The interplay between physical and biological processes can be used to explain the seasonal distribution of air to sea \( \text{CO}_2 \) flux. For example, NCP contributes to the uptake of \( \text{CO}_2 \). Variability in Chlorophyll-a\(^19. \) was used as a proxy for NCP (Supplementary Fig. 1). Chlorophyll-a is weakly correlated with phytoplankton biomass so that monthly changes in the former are only qualitatively related to NCP and by extension biological \( \text{CO}_2 \) uptake or release\(^20. \) A monthly increase in Chlorophyll-a\(^19. \) was 49% higher in winter to early spring (Dec-Mar) compared to summer (Jun-Aug), resulting in 149% greater net-integrated air to sea \( \text{CO}_2 \) flux and Chlorophyll-a are not perfectly anti-correlated, even during this net phytoplankton growth period, likely due to subsurface primary production in summer\(^21. \) hysteresis in \( \text{CO}_2 \) re-equilibration with \( \text{CO}_2 \)\(^22. \) and other factors. The apparently high levels of Chlorophyll-a found in winter (Dec. to Feb.), particularly in shallow coastal areas, are likely due to coloured organic matter and sediment reflectance rather than Chlorophyll-a \( \text{(i.e., Case II waters)} \)\(^23. \) (Supplementary Fig. 1). The solubility of \( \text{CO}_2 \) \( (\alpha_{\text{sea}} \text{ in Eq. 1)} \); a function of seawater temperature and salinity; Supplementary Fig. 2) contributes to the potential in-water-\( \text{CO}_2 \)-concentration \( \text{(fCO}_2 \text{ sea in Eq. 1)} \) and thus the \( \Delta \text{fCO}_2 \) \( (\text{fCO}_2 \text{ sea - fCO}_2 \text{ air}) \). Thereby, lower \( \alpha_{\text{sea}} \) in the southern North Sea in summer as well as remineralization of river-borne organic matter\(^24,25. \) likely contribute to the efflux of \( \text{CO}_2 \) to the atmosphere in this region. Finally, wind speed \( (U_{10}) \) (Supplementary Fig. 3) determines the gas transfer coefficient \( (k \text{ in Eq. 1)} \). Wind speed was 49% higher in winter to early spring (Dec-Mar) compared to summer (Jun-Aug), resulting in 149% greater values in winter and high drawdown of atmospheric \( \text{CO}_2 \).

The net-integrated air to sea flux of \( \text{CO}_2 \) for 2015 within the region defined by the 200 m isobath was 26.2 ± 4.7 Tg \( \text{y}^{-1} \) (Fig. 2c). Including the Norwegian trench, the net influx of \( \text{CO}_2 \) from the atmosphere was 26.7 ± 4.8 Tg \( \text{y}^{-1} \). \( \text{CO}_2 \) flux and air to sea flux were further separated into thermal \( (e.g., \text{due to cooling-enhanced solubility and biological + mixing components following Eqs. 2 and 3}) \) and biological + mixing components following Eqs. 2 and 3. In 2015, the mean biological + mixing component \( (\text{iCO}_2 \text{ bio}) \) closely matched the seasonal pattern of \( \text{fCO}_2 \text{ sea over the domain with lower fCO}_2 \text{ max in spring/summer (Fig. 2a). In contrast the air to sea CO}_2 \text{ flux and its thermal component, were highest in winter, (Fig. 2b,c). A concomitant increase in the gas transfer velocity was evident in winter (Fig. 2b). The annual mean thermal to biological + mixing components ratio was thereby 2.1 ± 1.7 (where the uncertainty given is one standard deviation). The highest values for this ratio, indicating thermal-flux-component dominance, were observed in the western Hebridean Shelf, English Channel and northern Celtic Sea (Fig. 2d). This was largely due to low
biological + mixing flux-component rather than high thermal-flux component in these regions. A notable exception was the Irish Sea where the net biological + mixing component was negative.

**C-budget terms.** Table 1 lists the C-budget terms considered in our study (mean ± standard deviation of corresponding estimates listed in Table 1). The mean multi-year air to sea flux term was calculated as 23.0 ± 4.3 Tg C y⁻¹. Baltic Sea C-inputs contribute 12.6 ± 2.7 Tg C y⁻¹ to the NW European shelf; 89% as DIC, the remainder as
organic-C. Rivers, were a smaller source of C contributing 2.4 ± 0.3 Tg C y⁻¹ of DOC. The sum of pre-estuarine inputs was 21.6 Tg C y⁻¹ while estuarine CO₂ efflux to the atmosphere was calculated as 24.4 ± 10.4 Tg C y⁻¹.

Our C-burial calculations showed that Nret accounted for 63%, 91% and 97% of Ndep in mud, mud-sand and sand sediments respectively (each with an uncertainty of ±22%). Assuming that the conditions in March were representative of the low-productivity (i.e. net respirations) half of the year (winter) and the average of the May and August were representative of the high productivity half (spring and summer), we calculated annual net C-burial as 0.569 mol C m⁻² y⁻¹, 0.055 mol C m⁻² y⁻¹ and 0.008 mol C m⁻² y⁻¹ for mud, mud-sand and sand respectively. The annual net C-burial over the NW European shelf was 1.3 ± 3.1 Tg C y⁻¹. This was therefore a small, but highly uncertain, loss term for shelf-sea C. The DIC accumulation rate, related to OA, was further scaled for the shelf area (1.06 × 10¹² m²) and mean depth (78.3 m) to yield 1.0 ± 0.5 Tg C y⁻¹. We acknowledge that this will vary regionally depending on the buffering capacity of seawater and other factors.

Finally, net exchange of C between the shelf and the open ocean via the CSP and advection was calculated as −34.4 ± 6.0 Tg C y⁻¹ from the sum of the input terms (from the atmosphere, rivers and the Baltic Sea) minus C-burial and DIC accumulation. The uncertainty of our export-estimate is the standard error of the uncertainties in the other terms. Our estimate of the CSP is consistent with a previous estimate of 32.8 Tg C y⁻¹ 26, which was based on the distribution of DIC in the North Sea. A further study estimated a monthly export of 2.2 Tg C via the Norwegian trench in late summer 27, which yields 27.2 Tg C y⁻¹ if extrapolated over the whole year.

Discussion

Over the last two decades, a number of studies have quantified the air to sea flux of CO₂ over the NW European shelf, regions within it or the wider European shelf 25–28. When scaled to their respective areas, these estimates (including our own) fall in the range of 1.3–2.1 mol C m⁻² y⁻¹ with a mean value of 1.8 ± 0.3 mol C m⁻² y⁻¹ (Table 1). This is three-fold higher than the average open ocean influx of 0.60 mol C m⁻² y⁻¹ 20, but lower than some high latitude seas (e.g. the Barents Sea: 4 mol C m⁻² y⁻¹ 31) and upwelling systems (e.g. the South African coastal region: 3.83 mol C m⁻² y⁻¹ 32). Our study provides the first such estimate based on observations collected in a single year. This offers unique advantages for understanding inter-annual variability related to climate and weather. For example, previous work has shown that the North Atlantic Oscillation (NAO; the dominant climate mode over the North Atlantic and a major influence on weather in NW Europe) exerts a strong influence on fCO₂ in the North Sea 33. In the absence of comparable, annual data this could not be explored...
further (for reference, the 2015 winter-NAO index was positive, associated with above average precipitation and temperatures over our domain). We found that the 2015 influx of atmospheric CO₂ was dominated by the winter months (January to March and December) despite the small air-sea concentration gradient during this period. This conclusion is further supported by the flux component analysis (thermal vs. biological mixing) which showed thermal influx dominance over a whole year. Nevertheless, we acknowledge that the flux component analysis creates an artificial situation whereby the expected thermal summer-efflux does not materialize because of net biological drawdown of fCO₂ sea at that time. Peak influx of CO₂ in winter, and/or due to high wind speeds and solubility, has also been observed in other shelf-seas.

The paucity of single-year studies leaves a number of open questions relating to the inter-annual persistence of the pattern observed in 2015 where the winter air to sea CO₂ flux was the dominant component of the net flux. Numerical models may be used to address this, but these are not entirely consistent with observations. Modelling studies for the NW European shelf are within ±50% of observation-based estimates [e.g. 1.5–2.6 mol C m⁻² y⁻¹ for the North Sea, 0 mol C m⁻² y⁻¹ for the English Channel, and 1.0 mol C m⁻² y⁻¹ for the Celtic Sea, Faroes, North Scottish shelf, West Scottish shelf, West Irish shelf, Irish Sea and North Channel – and subsequently corrected for the wind reanalysis by Meyer et al. (2018). Note that the annual accumulation in DIC (DIC accum.) is negative here – otherwise it would count against export. Export via the CSP and advection was calculated as the difference between inputs and C-burial + OA, assuming that 50% of the river-DOC input is mineralized on the shelf (and therefore accounted for in the air to sea flux).

Table 1. C-fluxes into and out of (negative) the NW European Shelf (values in bold represent the mean ± standard deviation of the estimates listed below each flux). Some terms are listed here for information but accounted in other terms and therefore do not contribute to the mean flux (e.g. the river-DIC input is accounted in the estuarine and shelf air to sea flux and therefore not included in the river-flux). The terrestrial input only considers dissolved organic carbon (DOC), as dissolved inorganic carbon (DIC) is implicitly accounted for in the air to sea flux. The air to sea flux estimate by25 was recalculated using 1.4 mol C m⁻² y⁻¹ for the North Sea, 0 mol C m⁻² y⁻¹ for the English Channel, and 1.0 mol C m⁻² y⁻¹ for the Celtic Sea, Faroes, North Scottish shelf, West Scottish shelf, West Irish shelf, Irish Sea and North Channel – and subsequently corrected for the wind reanalysis by Meyer et al. (2018). Note that the annual accumulation in DIC (DIC accum.) is negative here – otherwise it would count against export. Export via the CSP and advection was calculated as the difference between inputs and C-burial + OA, assuming that 50% of the river-DOC input is mineralized on the shelf (and therefore accounted for in the air to sea flux).

| Flux type | Flux (Tg C y⁻¹) | Reference |
|-----------|-----------------|-----------|
| River     | Σ(DOC) 2.4 ± 0.3 |           |
| DOC       | 2.6 post-estuary; scaled from⁴⁴ |           |
|           | 2.1 pre-estuary; scaled from⁴⁰ |           |
| POC       | 0.6 pre-estuary; scaled from⁴⁴ |           |
| DIC       | 2.0 post-estuary; scaled from⁴¹, accounted in air-sea flux |           |
|           | 18.9 pre-estuary; scaled from⁴⁵, accounted in estuarine (24.4 ± 10.4 Tg C y⁻¹) and shelf-air to sea flux |           |
| Baltic Sea| Σ(Σ(DOC + DIC) 12.6 ± 2.7) |           |
| DOC       | 0.8             | ⁵¹        |
|           | 1.9             | ⁵²        |
|           | 1.1             | ⁵⁶        |
|           | 1.9             | ⁵³        |
| DIC       | 9.7             | ⁵²        |
|           | 12.7            | ⁵⁶        |
| Air to Sea| 23.0 ± 4.3      |           |
|           | 26.2            | this study|
|           | 26.9            | Scaled from⁴⁴ |
|           | 19.8            | Scaled from⁴⁵ |
|           | 17.2            | ⁵⁶        |
| DIC accum. | −1.0 ± 0.5      |           |
|           | −0.6            | for ΔpH = −0.0013 units y⁻¹⁰⁶ |
|           | −1.0            | for ΔpH = −0.0022 units y⁻¹⁴⁵ |
|           | −1.5            | for ΔpH = −0.0035 units y⁻¹⁴⁵ |
| Sed. burial| −1.3 ± 3.1      |           |
|           | −1.8            | scaled from⁴⁰ |
|           | −0.9 ± 2.5      | this study |
| Export    | −34.4 ± 6.0     | this study – mass balance assuming that 50% of river-DIC is mineralized and accounted for in the air to sea flux |

Further (for reference, the 2015 winter-NAO index was positive, associated with above average precipitation and temperatures over our domain). We found that the 2015 influx of atmospheric CO₂ was dominated by the winter months (January to March and December) despite the small air-sea concentration gradient during this period. This conclusion is further supported by the flux component analysis (thermal vs. biological mixing) which showed thermal influx dominance over a whole year. Nevertheless, we acknowledge that the flux component analysis creates an artificial situation whereby the expected thermal summer-efflux does not materialize because of net biological drawdown of fCO₂ sea at that time. Peak influx of CO₂ in winter, and/or due to high wind speeds and solubility, has also been observed in other shelf-seas.

The paucity of single-year studies leaves a number of open questions relating to the inter-annual persistence of the pattern observed in 2015 where the winter air to sea CO₂ flux was the dominant component of the net flux. Numerical models may be used to address this, but these are not entirely consistent with observations. Modelling studies for the NW European shelf are within ±50% of observation-based estimates [e.g. 1.5–2.6 mol C m⁻² y⁻¹ for the North Sea⁵⁸,⁵⁹ compared to 1.8 ± 0.3 mol C m⁻² y⁻¹ for observations⁵⁶–⁵⁸]. Wider disagreement is found for long-term climate projections of wind speed (e.g. CMIP5) where different models predict increase or decrease in extreme wind events⁶⁰. If wind speed is critical in determining the overall annual influx of CO₂ from the atmosphere (as shown here), it follows that model projections of this influx are equally contradictory and therefore uncertain.

Our synthesis of C-fluxes showed that the air to sea flux is the largest input term over the open NW European shelf (Fig. 3). River-borne C-inputs are largely lost in estuaries with CO₂ efflux accounting for the majority of...
Reduced annually in shelf-seas is recycled rather than buried in sediments. A previous estimate for North Sea estuaries on the shelf.

The fate of C exported via the CSP and advection is unclear, only providing a climate regulation service if it is entrained in intermediate and deep waters of the open ocean. Climate-regulation service by removing C from the contemporary C-cycle. The fate of C exported via the CSP and advection to the north is unclear, only providing a climate regulation service if it is entrained in intermediate and deep waters of the open ocean.

Our calculation of net C-burial is consistent with previous work showing that most of the organic-C produced annually in shelf-seas is recycled rather than buried in sediments. A previous estimate for North Sea net C-burial was within the uncertainty of our estimate [0.9 Tg C y⁻¹ for the North Sea, scaled here for the larger surface area of our study = 1.8 Tg C y⁻¹]. Nevertheless, the C-burial term carries the largest proportional uncertainty. Although it is a relatively small term in this budget, net C-burial is of great interest as it provides a climate-regulation service by removing C from the contemporary C-cycle. The fate of C exported via the CSP and advection to the north is unclear, only providing a climate regulation service if it is entrained in intermediate and deep waters of the open ocean.

There is significant uncertainty in our estimate of the estuary to atmosphere CO₂ flux. Conservation of mass dictates that if this term changed then export via the CSP and advection should also change (other input/output terms were better constrained in our analysis). Considering lower and upper limits for the estuarine efflux of CO₂ to the atmosphere leads to the conclusion that the corresponding lower and upper limits for the export term would be 24 Tg C y⁻¹ and 44.8 Tg C y⁻¹ respectively. Independent estimates of the export term are consistent with our export estimate of 34.4 Tg C y⁻¹. Our conclusion that nearly all of the supply of DIC to estuaries is lost to the atmosphere within estuaries, is therefore plausible. However there is considerable work to do on refining our estimates of this term. In this respect, it is important to note that our budget only considered estuarine and shelf ‘open water’ and did not include salt-marshes and other ‘blue carbon’ stores, which are important sinks for C, but beyond the scope of this study. The suggestion that the riverine input of DIC and DOC to estuaries is largely lost to the atmosphere with those systems means that the open shelf net air to sea CO₂ flux is at present largely

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**Figure 3.** C-fluxes across the NW European shelf based on the present study and literature. Export is calculated as the difference between input/output terms and further subtracting the DIC accumulation term. (*) The post-estuarine DIC input is not considered in the calculation of export as it contributes to the air to sea flux and is therefore implicitly accounted for in this term. (**) We have arbitrarily assumed that 50% of post-estuarine river-DOC input is mineralized to CO₂ on the shelf – this is implicitly accounted for in the air to sea flux and does not contribute to export.
decoupled from terrestrial influence. Land-management policies which might alter the delivery of organic-C to
the NW European shelf (e.g. restoration of peat-bogs) would only have a modest effect on the open-shelf air to
sea CO₂ flux, burial and export-terms since the river-DOC input is a relatively small term. Instead, the impact of
varying river C-loads would likely be more pronounced in estuaries where some DOC-mineralization takes place
and contributes to efflux of CO₂. Changes in river nutrient concentrations driven by environmental policy would
be expected to affect net primary production and concomitant drawdown of atmospheric CO₂. However, if the
majority of atmospheric CO₂ influx is shown to consistently take place during winter (as shown here for 2015)
then the impact of such policies (designed to reduce eutrophication and improve the health of the coastal envi-
ronment) on the open shelf sink of atmospheric CO₂ will be limited. As was the case for river-C loads, the impact
of varying river nutrient loads is likely more pronounced in estuarine and near-shore environments.

Arguably the largest implications for the air to sea CO₂ flux relate to global CO₂ emission reduction policy. The
most pertinent questions relate to the longevity of the shelf uptake of atmospheric CO₂ when the latter is continu-
ously increasing. Long-term studies have shown that the accumulation of DIC is more rapid on the NW European
shelf than elsewhere, leading to higher rates of OA in the range of −0.0022 to −0.0035 pH units y⁻¹ ⁴²,⁴³, com-
pared to −0.0018 ± 0.0004 y⁻¹ globally ⁴⁴. This may be evidence that export via the CSP and advection is becoming
saturated in agreement with a modelling study which further predicted a negative feedback (decrease) on the air
to sea flux ⁴⁵. Yet, comparing our 2015 air to sea CO₂ flux to previous estimates based on observations in the 1990s
and 2000s does not support a reduction in atmospheric CO₂ influx. Continued international efforts under the
global ocean acidification observing network (http://goa-on.org/), ICOS network ⁴⁵ and others are essential to
help us understand, identify and ultimately predict changes in the annual shelf-sea uptake of atmospheric-CO₂.

Methods

The NW European shelf is hereby defined as the region within the 200 m isobaths north of 47 °N and 8.5 °E. Based
on a dataset of 298500 in situ observations of CO₂ fugacity (Supplementary Table 1) we calculated the air to sea
flux of CO₂ using the open-source FluxEngine toolbox for 2015 ⁴⁶. The dataset comprised three different method
classes: a) continuous-flow equilibrator with partial drying of the headspace gas stream and infra-red detection
following the recommendations of Pierrot et al. ⁴⁷ (equ-IR; 99.9 k observations), b) fCO₂ derived from discrete
measurements of Total Alkalinity and Dissolved Inorganic Carbon (TA/DIC-derived; 0.7 k observations) using
the CO2SYS software package ⁴⁸ and c) fCO₂ measured with a CONTROS HydroC CO₂ flow-through sensor ⁴⁹
(sensor; 198 k observations) (Fig. 4). Further methodological details are given in the Supplementary Information.

Each method for the determination of fCO₂ carries a certain analytical uncertainty defined by its respec-
tive precision and accuracy, which, in turn, are influenced by properties such as instrument-drift, calibration
frequency, equilibration timescale and resolution ⁵⁰,⁵¹. In order to quantify the internal consistency of observa-
tions we examined ‘crossovers’ between different datasets and for the three method classes (see Supplementary
Information for details). A crossover was defined as a maximum distance of 40 km between observations, using
30 km as the equivalent distance for one day (e.g. two data-points, one day apart in time and 10 km apart in space,
yield a nominal equivalent ‘distance’ between data-points of 40 km). The crossover analysis revealed a correlation
between the three different method classes (R² > 0.998) with linear slopes (bias) statistically indistinguishable
from unity. These statistics attest to the validity of the crossover analysis whilst the addition of ‘salinity and tem-
perature agreement’ criteria (<0.2 and <0.5 °C respectively) further minimized the influence of river plumes and
advection. There was therefore no systematic bias by any of the measurement techniques, which enabled us to use
the whole dataset for calculating air-sea fluxes. We calculated a weighted uncertainty of 13.2 μatm for the whole
dataset (note that this is not the accuracy which had a weighted mean value of 4 μatm based on the accuracy of
individual systems; see Supplementary Information). Whilst we stress that this was not a strict inter-comparison

Figure 4. Spatial (left) and temporal (right) data coverage for fCO₂ observations by three methods.
of methods, it nevertheless provides confidence in the dataset consistency as a whole and provides a useful metric to propagate in the calculation of flux uncertainty.

All instantaneous and integrated air–sea gas fluxes were calculated using the freely available FluxEngine toolbox. The air-sea flux of CO$_2$ was calculated from the following equation:

$$F = k \times (\alpha_{\text{sea}} \times f\text{CO}_2\text{ sea} - \alpha_{\text{air}} \times f\text{CO}_2\text{ air}),$$

where $k$ is the gas transfer velocity (units: cm h$^{-1}$), $\alpha_{\text{sea}}$ is the solubility of CO$_2$ at the bottom of the mass boundary layer and within the seawater (a function of salinity and temperature; units: g m$^{-3}$ atm$^{-1}$), $\alpha_{\text{air}}$ is the equivalent solubility at the top of the mass boundary layer (the air-side solubility; units: g m$^{-3}$ atm$^{-1}$), $f\text{CO}_2\text{ sea}$ and $f\text{CO}_2\text{ air}$ are the fugacity of CO$_2$ in seawater and air respectively (units: atm).

$f\text{CO}_2\text{ sea}$ data were re-analyzed to be valid for a consistent sea surface (sub-skin) temperature dataset and depth$^{35}$ using a climate quality (sub-skin) satellite temperature dataset$^{35}$. These $f\text{CO}_2\text{ sea}$ data (valid for the sub-skin and assumed to represent the bottom of the mass boundary layer) were then spatially interpolated (Data Interpolating Variation Analysis, DIVA v 4.7.1)$^{34}$ to a 50 km polar stereographic grid. A cool skin correction of $-0.17^\circ$C was used$^{38}$ to enable the calculation of skin temperature (using the sub-skin dataset as a reference) and thus the air-side solubility $\alpha_{\text{air}}$. Atmospheric $f\text{CO}_2\text{ air}$ was calculated within FluxEngine using the air pressure at the sea surface (European Centre for Medium Range Weather Forecasting re-analysis; https://www.ecmwf.int/en/ research/climate-reanalysis/era-interim) and in-situ CO$_2$ dry air mole fraction from the National Oceanic and Atmospheric Administration Earth System Research Laboratory$^{49}$. The atmospheric CO$_2$ dataset used here offers a well characterized, quality controlled, ‘clean’ and consistent dataset for the calculation of flux. Wind speed ($U_{10}$) data were obtained from the Cross-Calibrated Multi-Platform (CCMP) version 2$^{57}$. Salinity data came from the World Ocean Atlas 2013$^{58}$. We used a quadratic wind driven parameterization of the gas transfer velocity$^{59}$ that was originally developed using shelf sea measurements. All input data were re-gridded to the 50 km polar stereographic grid. Total uncertainties in the air to sea flux calculations solely due to the input data were calculated using an ensemble method. Twenty five sets of calculations were made, with $f\text{CO}_2\text{ sea}$, SST and $U_{10}$ inputs perturbed by random noise representing the known uncertainties of the input parameters; $\sigma(f\text{CO}_2) = 1.32$ μatm; $\sigma(U_{10}) = 0.44$ ms$^{-1}$; $\sigma$(SST skin and sub-skin) = 0.6°C. This ensemble gave an air-sea flux uncertainty of ±2%.

In order to investigate uncertainties arising from the DIVA interpolation of $f\text{CO}_2\text{ sea}$ we compared the interpolated outputs with independent $f\text{CO}_2\text{ sea}$ data from a buoy in the Western English Channel ($f\text{CO}_2\text{ buoy}$ at station L4 operated by PML and not included in the combined $f\text{CO}_2\text{ sea}$ dataset described above; see Supplementary Information for details). The corresponding daily air to sea flux ($F_{\text{buoy}}$) was computed using the same parameters as in FluxEngine (i.e. with the same $k_0$, $\alpha_{\text{sea}}$, $\alpha_{\text{air}}$ and $f\text{CO}_2\text{ air}$), apart from the $f\text{CO}_2\text{ sea}$ field. The daily flux using $f\text{CO}_2\text{ buoy}$ was not significantly different from the corresponding monthly flux using the combined $f\text{CO}_2\text{ sea}$ dataset described above (paired t-test; $t = -9.34, p < 0.001, n = 184$). Nevertheless, the 95% confidence interval 0.003 mol m$^{-2}$ month$^{-1}$ represented an uncertainty of 16% of the annual flux at L4 when extrapolated for the whole year. The maximum total uncertainty from mapping and the ensemble runs described above (perturbations of $f\text{CO}_2\text{ sea}$, $U_{10}$ and SST) was therefore 18%. An additional 5% uncertainty arises from three quadratic parameterizations of $k_0$ for the North Atlantic$^{60}$. However, the ongoing debate regarding the optimum parameterization of $k_0$ is beyond the scope of this paper and not considered further.

We separated the mean $f\text{CO}_2$ for each 50 km grid cell into its biological + mixing (Eq. 2) and thermal components (Eq. 3)$^{60}$.

$$f\text{CO}_2\text{ bio} = f\text{CO}_2\text{ sea} \times e^{0.0423 \times (\text{SST} - \text{SST})},$$

(2)

$$f\text{CO}_2\text{ therm} = -f\text{CO}_2\text{ sea} \times e^{0.0423 \times (\text{SST} - \text{SST})},$$

(3)

where $f\text{CO}_2$ and $f\text{CO}_2\text{ sea}$ are the observed and annual mean $f\text{CO}_2$ for each grid cell, SST and $\text{SST}$ are the observed and annual mean sea surface temperature respectively (sub-skin). $f\text{CO}_2\text{ therm}$ and $f\text{CO}_2\text{ bio}$ were subsequently used to replace $f\text{CO}_2\text{ sea}$ in Eq. 1 in order to derive the respective flux components.

C-burial in shelf sediments was calculated for three sites in the Celtic Sea: a) a muddy sediment (51.2114 °N, 6.1338 °W), b) a sandy sediment (51.0745 °N, 6.5837 °W) and c) a mud-sand sediment (station CCS: 49.4117 °N, 6.1338 °W), b) a sandy sediment (51.0745 °N, 6.5837 °W) and c) a mud-sand sediment (station CCS: 49.4117 °N, 6.1338 °W). The following biogeochemical rates were determined at these sites in March (end of winter), May (during the spring bloom) and August 2015 (late summer): benthic oxygen consumption (Resp), nitrification (Nit), denitrification (Den), anammox (Ax) and sediment-water inorganic-N fluxes (FN; for NO$_3^-$ + NO$_2^-$ + NH$_4^+$) (e.g. Supplementary Table 3)$^{61}$. Our general reasoning for the C-burial calculation relies on the conservation of mass and specified C:N ratios in organic matter, where near-closure of the C-cycle for NO$_3^-$ and annual mean sea surface temperature respectively (sub skin).
[982 \times 10^4 \text{km}^{2.6}] to estimate a post-estuarine input of 4.8 Tg C y\(^{-1}\). This was further separated into DIC (41.1%), DOC (54.3%) and POC (4.4%)\(^{24}\). The latter (0.2 Tg C y\(^{-1}\)) is thought to be trapped in estuaries where it contributes to C-burial and outgassing\(^{24}\) and is therefore not considered further as an input to the NW European shelf. By comparison, the freshwater-endmember (i.e. pre-estuarine) of Grisupos to the North Sea\(^{26}\), scaled to the shelf area in our domain, yielded 18.9 Tg C y\(^{-1}\) and 2.1 Tg C y\(^{-1}\) for DIC and DOC respectively. The estimates of Ciais et al. (2008) and Thomas et al. (2005) are roughly in agreement for DOC, but differ by one order of magnitude for DIC. However, correcting the higher estimate of Thomas et al. (2005) for CO\(_2\) outgassing could easily account for this difference. We estimate the estuarine efflux of CO\(_2\) to the atmosphere as 24.4 ± 10.4 Tg C y\(^{-1}\), based on: a) estuarine sea to air efflux of 19.8–62.0 mol C m\(^{-2}\) yr\(^{-1}\)\(^{52,56,65}\); b) the ratio of estuarine area to coastline length of 2.64 km\(^2\) km\(^{-1}\); and c) the coastline length of the domain considered here (18204 km).

As a result of long-term, net CO\(_2\) influx from the atmosphere and lateral advection from the open ocean, the NW European shelf is gaining C in the form of DIC which causes Ocean Acidification (OA). Whilst this is not an input term, it does not contribute to export via the CSP + advection and must therefore be accounted for as the latter is the remainder of input vs output terms. In our study, this DIC accumulation rate was estimated from annual OA rates in the range of -0.0013 to −0.0035 pH-units\(^{42,43,69}\). We estimated a mean DIC accumulation rate of 1.01 μmol kg\(^{-1}\) yr\(^{-1}\) using the CO2SYS software\(^{48}\) with pH\(_{eq}\) = 8.05 and Total Alkalinity of 2283.1 μmol kg\(^{-1}\) as baseline.

**Data availability**
All ICO\(_2\) data used in this study are available from the SOCAT and Ferrybox databases (www.socat.info and www.ferrybox.org).

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### Author contributions

V.K. designed the study and led the writing. J.D.S., I.A., M.W., P.E.L., D.D., B.G., A.C. contributed to writing this manuscript. J.B., H.F., S.H.E., M.H., C.K., N.G., T.H., D.P., T.Mc.G., B.M.S., P.W., E.Mc.G., Y.B., J.-P.G., Sv.H., M.H., U.S., T.J., A.O., S.K.L., I.S., A.O., T.S., A.K., M.B., N.L., T.G., A.C., W.P., Y.V. contributed fCO₂ data. All authors contributed to Flux Engine work, I.B., H.F., S.H.E., M.H., C.K., N.G., T.H., D.P., T.Mc.G., B.M.S., P.W., E.Mc.G., Y.B., J.-P.G., Sv.H., M.H., U.S., T.J., A.O., S.K.L., I.S., A.O., T.S., A.K., M.B., N.L., T.G., A.C., W.P., Y.V. contributed fCO₂ data. All authors contributed to writing this manuscript.

### Competing interests

The authors declare no competing interests.

### Additional information

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