



Updated climatological mean delta fCO₂ and net sea–air CO₂ flux over the global open ocean regions

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20 Key Points

- An updated surface water CO₂ climatology for 1980-2021 is created using the SOCAT database, following procedures of Takahashi et al. (2009)
- A net air-sea CO₂ flux of -1.79 PgC yr⁻¹ is determined for near-global ocean coverage



25 Abstract

The late Taro Takahashi (LDEO/Columbia University) provided the first near-global monthly air-sea CO₂ flux climatology in Takahashi et al. (1997), based on available surface water partial pressure of CO₂ measurements. This product has been a benchmark for uptake of CO₂ in the ocean. Several versions have been provided since, with improvements in procedures and large increases in observations, culminating in the authoritative assessment in Takahashi et al. (2009). Here we provide and document the last iteration using a greatly increased dataset (SOCATv2022) and determining fluxes using air-sea partial pressure differences as a climatological reference for the period 1980-2021 (Fay et al. 2023). The resulting net flux for the open ocean region is estimated as -1.79 PgC yr⁻¹ which compares well with other global mean flux estimates. While global flux results are consistent, differences in regional means and seasonal amplitudes are discussed. Consistent with other studies, we find the largest differences in the data-sparse southeast Pacific and Southern Ocean.

1. Introduction

40 Atmospheric carbon dioxide (CO₂) levels now exceed 415 ppm on an annual basis and the continued growth of the atmospheric reservoir represents a major societal concern due to the impact on the radiative balance of the atmosphere. Warming and associated environmental changes including sea-level rise and ocean acidification have adverse effects on countless aspects of terrestrial and marine ecosystems which in turn impact air-sea exchange of CO₂ and the trajectory of the atmospheric CO₂ levels. The annually updated Global Carbon Budget (GCB) report estimates current net global ocean carbon uptake has been estimated at nearly 3.0 PgC per year, which corresponds to about a quarter of the total annual emissions (the total anthropogenic CO₂ emission, including the cement carbonation sink, is estimated at 10.9 ± 0.8 PgC yr⁻¹) (Friedlingstein et al. 2022). Given the importance of the ocean as a CO₂ sink, it is essential to continuously monitor changes and improve our understanding of the ocean's role in the global carbon cycle.

55 Over the last several decades, multiple approaches have been developed to measure the impact of the ocean on the global CO₂ cycle. These approaches include atmospheric inversions (Feng et al. 2019), global atmospheric O₂/N₂ (Manning & Keeling 2006), ¹³C measurements (Quay et al. 1992, Tans et al. 1993), ocean inventory approaches (Gruber et al. 2023) and the measurement of surface ocean and atmospheric CO₂ (Takahashi et al. 1993); all methods work towards the goal of elucidating the net flux of CO₂ from the atmosphere into the ocean. These different



approaches have multiple advantages and disadvantages depending on the time and spatial scale of interest. Directly measuring surface ocean and atmospheric CO₂ levels has the advantage, given sufficient measurements, of deriving spatial and temporal
65 variability over the ocean surface on short temporal and spatial scales. This method provides valuable insights into key processes driving the uptake and emissions of carbon when combined with our understanding of ocean physics and biological activity.

The surface ocean and atmospheric CO₂ approach leverages available observations and the dynamic sea-air gradient between the partial pressure of carbon dioxide (pCO₂) in the surface ocean (pCO₂^{oce}) and the overlying atmosphere (pCO₂^{atm}), known as the delta pCO₂ (ΔpCO₂) and typically defined as pCO₂^{oce} - pCO₂^{atm}. This difference is the thermodynamic driving force for the transfer of CO₂ into (negative) and out of (positive) the ocean. On average, the ΔpCO₂ across the global oceans are becoming increasingly
70 negative as atmospheric CO₂ levels steadily rise, leading to an increasing carbon sink. Limited regions around the globe are sources of CO₂ to the atmosphere, including the equatorial Pacific Ocean and other areas of persistent upwelling.
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The late Taro Takahashi was a leader in efforts to characterize air-sea CO₂ flux through the design and deployment of pCO₂ systems throughout the global oceans, and perhaps most importantly, his efforts to assemble, evaluate and construct global ocean climatologies from available pCO₂^{oce} datasets. These versions of the ocean pCO₂ climatologies have been presented in literature (Takahashi et al. 1997, 2002, 2009, and 2014), henceforth referred to as T-1997, T-2002, T-2009, and T-2014; all of which have
80 been highly utilized and cited by carbon cycle researchers from around the world. We present an updated near-global climatological mean distribution and net sea–air CO₂ flux which represent the mean of ocean conditions over the last four decades. This climatology is unique compared to other advanced machine learning approaches (e.g., Rödenbeck et al. 2015) in that it interpolates in time and space using only the available
85 pCO₂ data rather than using proxy variables for gap filling. This difference in methodology provides a valuable alternative approach to the ongoing effort to characterize the global ocean carbon sink. This benchmark is critical for global carbon assessments, notably the Regional Carbon Cycle Assessment and Processes (RECCAP2, DeVries et al. 2023) effort.
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95 Building on previous work of Takahashi and colleagues, we employ the same time-space interpolation method used in the previous versions of the Takahashi climatology (e.g., T-2002, T-2009, T-2014) to create the climatology. However, here we use the Surface Ocean CO₂ Atlas (SOCAT) v2022 database (Bakker et al. 2016, 2022), rather than the LDEO database curated by Taro Takahashi. We use the SOCAT database for
100 this update because it is the most comprehensive database of available observations



105 from international research groups. We have included the climatology produced using
the most recent LDEO database (LDEOv2019, Takahashi et al. 2020) with data
extending to 2019, in supplementary figures and text but our main findings will focus on
the results from the SOCAT database.

2. Data

2.1 SOCAT database

110 The SOCAT database was first released in 2013 (Pfeil et al. 2013) and is updated
annually (Bakker et al. 2016). Observations are reported as values of fugacity of CO₂
(fCO₂) in micro atmospheres (µatm), along with a collection of ancillary data including
concurrent observations such as SST (more accurately, the ship intake temperature),
temperature of equilibration, salinity, and sea level pressure at the time of equilibration.
115 The SOCAT database also includes supplemental variables with values interpolated
from gridded global datasets such as satellite SST and NCEP sea level pressure. The
database restricts the included data to only observations that are measured in near-
continuous operation or in discrete samples with an equilibrator system; specifically, it
does not include fCO₂ values that are calculated from other ocean carbon
measurements including dissolved inorganic carbon, total alkalinity and/or pH. More
120 information on the SOCAT database is available in Bakker et al. (2016); current and
previous releases are available to download at [https://www.socat.info/index.php/data-
access/](https://www.socat.info/index.php/data-access/).

125 The SOCAT database (Bakker et al. 2016) is the largest and most widely used
collection of quality-controlled fCO₂ data with over twice the number of observations
included in the latest LDEO database (LDEOv2019). In this work, we utilize the
SOCATv2022 release which includes over 33.7 million observations spanning the years
1957 through 2021 (DOI:10.25921/1h9f-nb73, accessed on July 15, 2022, Bakker et al.,
2022). We use observations beginning in 1980 due to limited metadata available for
130 earlier observations. This time restriction eliminates only 24,786 observations, or less
than 0.1% of the total number of observations included in the SOCATv2022 release. We
also exclude coastal observations collected within 100km of land similar to past LDEO
climatologies which reduces the total number observations utilized to just over 21.3
million. Unlike past LDEO climatologies, this climatology does not exclude observations
135 collected in the equatorial Pacific during El Niño periods.

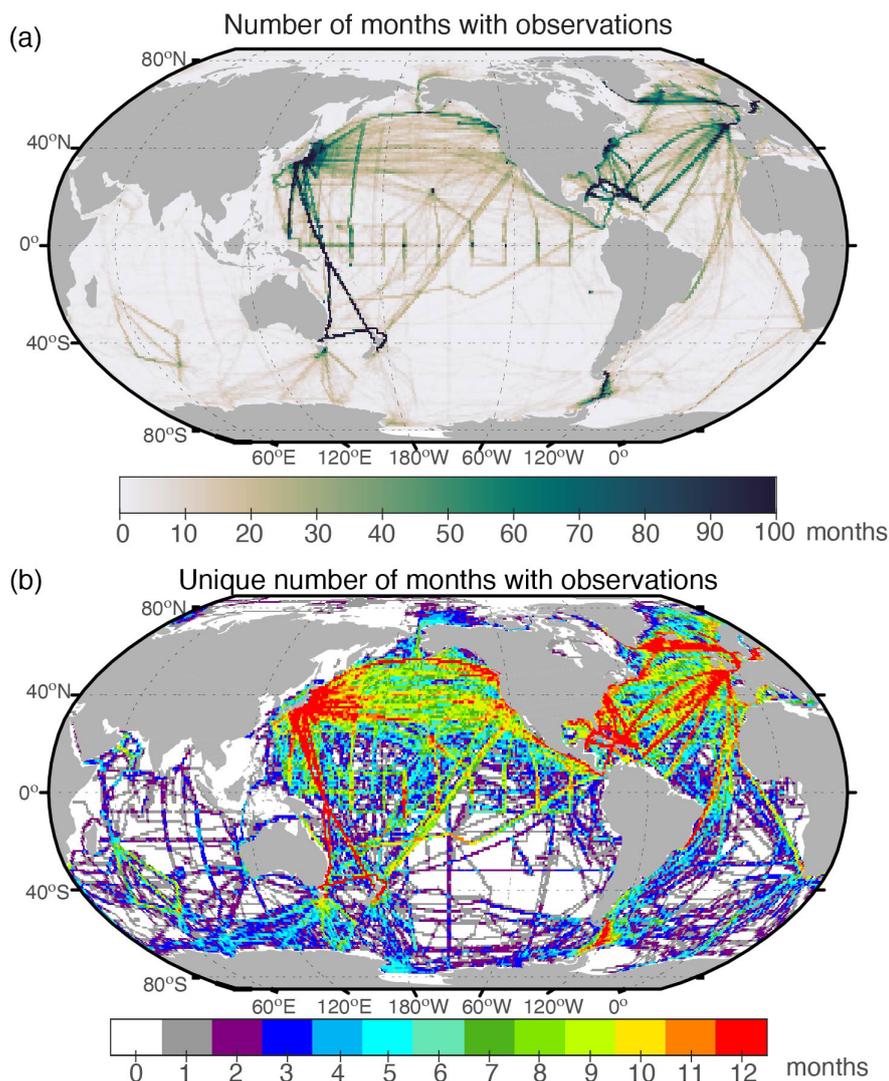


2.2 $f\text{CO}_2$ vs $p\text{CO}_2$

For this climatology, we report values of the fugacity of carbon dioxide ($f\text{CO}_2$) rather than $p\text{CO}_2$. The $f\text{CO}_2$ is equal to the $p\text{CO}_2$ corrected for non-ideality of CO_2 solubility in water using the virial equation of state (Weiss 1974). The fugacity correction for surface water is 0.996 and 0.997 at 0 °C and 30 °C respectively (Dickson et al. 2007), or 0.7 to 1.2 μatm lower than the corresponding $p\text{CO}_2$, and depends primarily on temperature for the conversion although pressure is also included in the conversion equation. It is now common practice in the observational community to present observed values as $f\text{CO}_2$ and this option has been endorsed by the IOCCP (International Ocean Carbon Coordination Project). The correction of $p\text{CO}_2^{\text{oce}}$ and corresponding $p\text{CO}_2^{\text{atm}}$ values to $f\text{CO}_2$ is practically identical such that the resulting $\Delta f\text{CO}_2$ is always within 0.1 μatm compared to the corresponding $\Delta p\text{CO}_2$. As a result, this difference will not have a meaningful impact on air-sea flux calculations. Only at elevated $f\text{CO}_2$ levels, such as those in the subsurface ocean, is the difference between $f\text{CO}_2$ and $p\text{CO}_2$ significant. Therefore, the shift in this climatology from $p\text{CO}_2$ to $f\text{CO}_2$ simply aligns this updated climatology with current community best practices. This choice avoids conversions given that the SOCAT database reports $f\text{CO}_2$ values.

2.3 Distribution of measurements

At present, the SOCAT database relies on voluntary submission of quality controlled data from over 100 scientists. The number of observations has increased significantly over past decades, facilitated by a now-automated data submission process. Even with this increase in observations, there are only a few regions over the global oceans where $f\text{CO}_2^{\text{oce}}$ has been systematically monitored over multiple decades at nearly the same location (Bakker et al. 2016; Bates et al. 2014; Landschützer et al. 2016, Figure 1). Of the observations considered in this analysis, spanning years 1980-2021, only 1.4% of the monthly 1° by 1° global ocean grid cells have measured values. Most of the data (65%) was collected since 2010.



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Figure 1: (a) Total number of months with at least one observation in each 1° grid cell in the SOCATv2022 database, for years 1980-2021 (Bakker et al. 2022). The maximum number possible for a grid cell is 504 (42 years * 12 months). (b) the number of unique calendar months in each grid cell where at least one observation has been made since 1980. Red indicates grid cells where each month (Jan - Dec) has been sampled at least once over the 40+ year time series while white indicates grid cells with no measurements over the length of the time series.

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180 An additional challenge to global monitoring efforts is that observations are not collected consistently throughout the annual cycle in many locations around the globe, thus requiring considerable interpolation to produce a full seasonal climatology. Much of the ocean contains data collected in fewer than three unique months of the year, regardless of how many years of data is available (Figure 1b). Current efforts utilize proxy variables and machine learning to identify relationships between ocean carbon and better-observed variables (often SST, chlorophyll, mixed layer depth, etc.) and then through those relationships extrapolate available ocean $f\text{CO}_2$ to fill the missing months of the seasonal cycle. Unlike those methods (e.g., methods compared by Rödenbeck et al. 185 2015), we do not utilize any proxy variables in this method and only rely on available $f\text{CO}_2^{\text{oce}}$ values in the SOCAT database to estimate the seasonal climatology. Hence, this effort provides a complementary alternative interpolation method to other approaches.

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2.4 Atmospheric $f\text{CO}_2$

For the calculation of atmospheric carbon dioxide, we utilize zonally invariant NOAA marine boundary layer (MBL) $x\text{CO}_2$ values which are reported in units of ppm or $\mu\text{mol/mol}$ (Lan et al. 2023) and provided with each observation in the SOCAT dataset. 195 In order to calculate $f\text{CO}_2^{\text{atm}}$ values from MBL $x\text{CO}_2^{\text{atm}}$ values, we follow standard operating procedures and equations outlined in Dickson et al. (2007) and use the SST, salinity and sea level pressure observations also reported for each value in the SOCAT database (Bakker et al. 2022). The SST (more specifically, ship intake temperature) is measured concurrently with surface ocean CO_2 , sea level pressure is from the National 200 Centers for Environmental Prediction (NCEP), and surface salinity is from the World Ocean Atlas (WOA). Delta $f\text{CO}_2$ ($\Delta f\text{CO}_2$) is calculated by subtracting corresponding atmospheric from ocean values ($f\text{CO}_2^{\text{oce}} - f\text{CO}_2^{\text{atm}}$).

205 3. Methods

3.1 Normalization to a reference year

In previous versions of the LDEO climatology, emphasis was placed on the calculation of trends in surface ocean carbon levels for all regions of the global ocean. These trends were used to normalize all available observations to one reference year by 210 correcting for the estimated change that would be expected between the collection date of the observation and the reference year. In this updated climatology, we use $\Delta f\text{CO}_2$



values as input to the algorithm rather than allowing for the adjustment of $f\text{CO}_2^{\text{oc}}$ to a specific reference year. A similar methodology was used in early versions of the LDEO climatology (i.e., T-1997).

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By utilizing this method to collapse all available data to one year, we make the assumption (as made by T-1997) that the ocean and atmosphere are changing at the same rate and thus the $\Delta f\text{CO}_2$ has been constant over the 40+ years of observations. This assumption allows for a standard method for the normalization of all observations to one calendar year. This method is utilized in contrast to the most recent LDEO climatologies where trends over distinct time periods were investigated and one trend then selected for use in time-normalization throughout much of the global ocean (i.e., T-2002, T-2009, and T-2014).

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Atmospheric $p\text{CO}_2$ change drives rising ocean $p\text{CO}_2$, and surface ocean carbon concentrations follow atmospheric increases on multi-decadal timescales and over large regions and the global scale (Fay & McKinley 2013, McKinley et al. 2020). Large synthesis efforts by those in the global ocean carbon community show that even if ocean trends are larger/smaller than the atmosphere on decadal or multi-year time periods, when considering the longest time periods, the atmosphere and ocean carbon trends are statistically indistinguishable over much of the global ocean (Fay & McKinley 2013, Tjiputra et al. 2014). While the Tjiputra et al (2014) and Fay & McKinley (2013) efforts consider large regional analysis, Bates et al. (2014) provides a synthesis of trends in $p\text{CO}_2$ at long-term observing stations, with most of the stations showing a match to the rise in atmospheric CO_2 concentration (Tanhua et al. 2015).

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Supplementary Table 1 shows biome-scale mean $f\text{CO}_2$ trends computed using all available $1^\circ \times 1^\circ$ grid cells with observations in the gridded product released as part of SOCATv2022 (Sabine et al. 2013). We present seasonal trends for each biome due to seasonal sampling bias over much of the global oceans. Similar to T-2009 (see Tables 1-5 of T-2009), trends for different ocean regions vary significantly (Supplementary Table 1) due in part to differing years with available data across ocean regions (Supplementary Figure 5). Observations within the Indian Ocean, for example, and other regions with blue shading in Supplementary Figure 5 are weighted towards the 1980's and 1990's while regions stretching across the North Atlantic and North Pacific have been heavily sampled over the last two decades with ships of opportunity; These well sampled regions have median years of collected observations later than 2010 (areas with red shading in Supplementary Figure 5).

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Recent studies (Friedlingstein et al. 2022) demonstrate that globally, the oceans lag slightly behind the atmosphere in terms of rates of carbon increase, and thus $\Delta f\text{CO}_2$ has



become increasingly negative as noted above. The central climatological year represented by our method is thus somewhat ambiguous regionally, though globally it is centered at about 2010; the median year of all $f\text{CO}_2$ observations collected in
255 SOCATv2022 greater than 100km from land is 2013; since observations are more densely clustered in the recent period, observations in the early period may have a greater weight in determining climatological values. Given global trends, our approach may estimate a smaller ocean sink in regions where the ocean was sampled more heavily early in the time period (e.g., blue shading in Supplementary Figure 5) and a
260 greater ocean sink in regions with heavy recent sampling (red shading in Supplementary Figure 5). We acknowledge that the assumption of a constant $\Delta f\text{CO}_2$ does not take into account the long-term trend in $\Delta f\text{CO}_2$, however, our simplified approach avoids application of trends determined for well-observed regions and time periods across poorly-observed regions and time periods.

265 We conducted a sensitivity analysis to demonstrate the impact of the $\Delta f\text{CO}_2$ method implemented in this version versus a normalization approach similar to that applied in T-2009. Specifically, we assumed a homogeneous $1.5 \mu\text{atm yr}^{-1}$ trend in $f\text{CO}_2^{\text{oce}}$ for all regions and years, and normalized available observations to a reference year of 2010.
270 Spatial maps of the differences in $f\text{CO}_2$ for the year 2010 for the normalization approach versus $\Delta f\text{CO}_2$ method are shown in the supplemental information (Supplementary Figure 6). Globally, the annual ocean uptake created using the $1.5 \mu\text{atm yr}^{-1}$ normalization method (T-2009) is within 3% of the $\Delta f\text{CO}_2$ method (this analysis); specifically, $-1.85 \text{ PgC yr}^{-1}$ versus $-1.79 \text{ PgC yr}^{-1}$ for the $1.5 \mu\text{atm yr}^{-1}$ normalization
275 method and the presented $\Delta f\text{CO}_2$ approach, respectively, for a reference year of 2010.

3.2 Method for time–space interpolation

The method for spatial interpolation and day of year utilized in the climatology has not changed from the previous versions (e.g., T-2009). As described above, $\Delta f\text{CO}_2$ values
280 are used to compile observations into one reference year in contrast to the time-normalization approach of T-2009. The spatial interpolation scheme requires that all observations are binned into $4^\circ \times 5^\circ$ grids for each day of the year. In some areas of the global ocean, such as the northern and equatorial ocean regions, there are observations in a majority of the pixels. However, vast expanses in the Southern
285 Hemisphere have few observations in each pixel and there are many pixels that contain no observations at all (Figure 1).

We follow the same methodology as T-2009 for binning observations in sparsely sampled grid cells south of 12°S . Here, spatial binning is increased by 4° and 5°
290 longitude and latitude, respectively, extending from the center of each grid cell. This



creates a grid of overlapping $8^{\circ}\times 10^{\circ}$ grid cells. Additionally, bins include the day before and after a given day of year. The mean is computed by weighting measured values inversely proportional to their time-space distance from the pixel center. After the above procedures are applied, more than 50% of the space-time pixels over the global oceans are filled.

To estimate the $\Delta f\text{CO}_2$ values in the remaining cells, an interpolation equation based on the 2-D diffusion–advection transport of surface waters is used, as in T-2009. The equation is discretized onto a $4^{\circ}\times 5^{\circ}$ grid for the global ocean, and solved iteratively using a finite difference algorithm (Takahashi et al. 1995, T-1997). The method avoids singularities at the poles by assigning land to each high latitude region (Antarctica in the south and treating the highest latitudes of the Arctic Ocean as land in the north). With this method, the resulting $\Delta f\text{CO}_2$ values are the solutions obtained after 500 iterations, as previously determined on the basis of interpolation experiments of temperature values (T-2009).

With this interpolation scheme, observed $\Delta f\text{CO}_2$ values where available are preserved, and the continuity equation is used to compute values for grid cells that have no observations. Consistent with previous iterations of this approach, the combined effects of internal sources and sinks of carbon, CO_2 exchange with the atmosphere, as well as upwelling of deep waters are all assumed to be included in the analysis of the observations that feed into the interpolation scheme. Uncertainties persist due to the sparsity of input data, normalization to a reference year, and the space-time interpolation. In part to address these uncertainties, we report only monthly means.

To maintain consistency with similar products and input datasets for flux calculations, we downscale to one degree boxes by assigning all $20\ 1^{\circ}\times 1^{\circ}$ pixels in a $4^{\circ}\times 5^{\circ}$ grid cell the same $\Delta f\text{CO}_2$ value. When calculating sea-air fluxes, because the other inputs to the flux calculation such as wind speed, are varying on a $1^{\circ}\times 1^{\circ}$ degree resolution grid, differences in the gridded flux climatology emerge on this finer spatial scale.

3.3 Flux calculation method: pySeaFlux

To assess the near-global ocean carbon sink associated with these $\Delta f\text{CO}_2$ estimates, air-sea CO_2 exchange must be calculated. The gridded monthly $1^{\circ}\times 1^{\circ}$ $\Delta f\text{CO}_2$ values were used to compute air-sea CO_2 fluxes using the bulk formulation with python package Seaflux.1.3.1 (<https://github.com/lukegre/SeaFlux>, Gregor & Fay, 2021). The net sea–air CO_2 flux (F) is estimated using:

$$\text{Flux} = kw \cdot \text{sol} \cdot (f\text{CO}_2^{\text{oce}} - f\text{CO}_2^{\text{atm}}) \cdot (1 - \text{ice}) \quad \text{Eq. (1)}$$



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where k_w is the gas transfer velocity, sol is the solubility of CO_2 in seawater (in units of $mol\ m^{-3}\ \mu atm^{-1}$), fCO_2^{oce} is the partial pressure carbon in the surface ocean (in μatm), and fCO_2^{atm} (in units of μatm) represents the atmospheric CO_2 levels in the marine boundary layer. Finally, to account for the seasonal ice cover at high latitudes, the

335 fluxes are weighted by one minus the ice fraction (ice), i.e., the open ocean fraction. By utilizing the pySeaFlux package (Fay & Gregor et al. 2021, Gregor & Fay 2021), we are able to include multiple scaled gas transfer velocities for three different wind products and our resulting flux estimate is a mean of the three. Additional inputs to the flux calculation include EN4.2.2 salinity (Good et al. 2013), SST and ice fraction from NOAA
340 Optimum Interpolation Sea Surface Temperature V2 (OISSTv2, Reynolds et al., 2002), European Centre for Medium-Range Weather Forecasts (ECMWF) ERA5 sea level pressure (Hersbach et al. 2020). Finally, surface winds and associated wind scaling factor for the Cross-Calibrated Multi-Platform v2 (CCMP2; Atlas et al. 2011), the Japanese 55-year Reanalysis (JRA- 55; Kobayashi et al. 2015), and the ECMWF ERA5
345 (Hersbach et al. 2020) reanalysis products are used.

Fluxes reported here use inputs from the year 2010 for the k_w , sol , and ice fraction variables. Alternatively, we have calculated fluxes using a mean over a 17-year time period centered on the year 2010. This yields a very similar value with the mean
350 difference being a $0.04\ PgC\ yr^{-1}$ increase in estimated carbon uptake.

4. Results

4.1 Climatological mean distribution of surface water ΔfCO_2

4.1.1 Global

The near-global 12-month climatological mean distribution of ΔfCO_2 (fCO_2^{oce} minus
355 fCO_2^{air}) is reported for the SOCAT database (Figure 2, Fay et al. 2023). Evident in the mapped climatology are the large-scale patterns across the global ocean: the consistent high (positive) ΔfCO_2 values in the equatorial Pacific region where upwelling is a dominant influence, and low (negative) values of ΔfCO_2 in the North Atlantic region where evaporation leads to increase salinity and cooling driving strong uptake of carbon
360 and subduction of surface waters.

A near-global mean climatology curve shows a bimodal shape in ΔfCO_2 , with a smaller peak in boreal spring (March/April) and a much larger peak in late boreal summer (August/September). The curve reaches its minimum in November and begins a
365 recovery throughout the boreal winter before dipping again to a springtime minimum in



June. The near-global annual mean $\Delta f\text{CO}_2$ value is $-4.1\mu\text{atm}$ and it is notable that the mean $\Delta f\text{CO}_2$ value is below zero for every month of the year, suggesting that seasonally the global ocean mean is a perpetual carbon sink with expansive regions of uptake nearly always outweighing smaller regions of efflux (Figure 2, 3).

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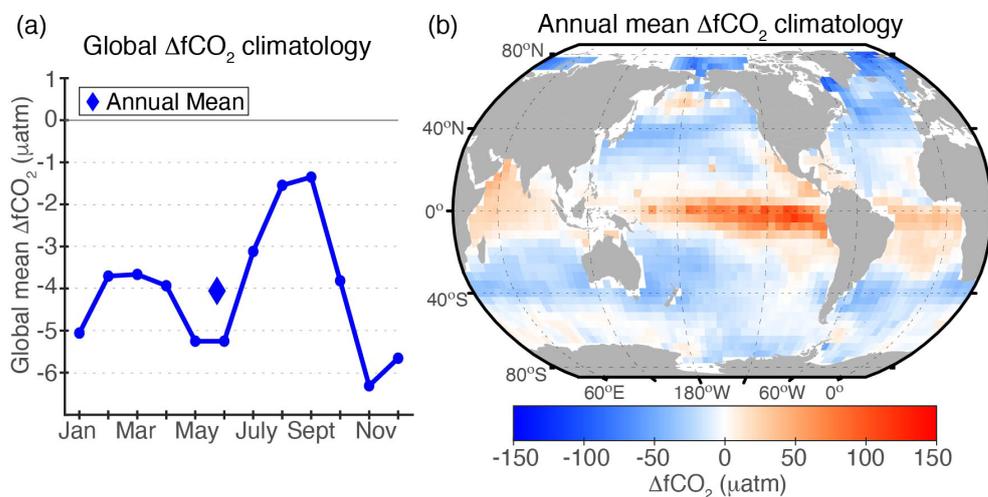


Figure 2: (a) Global mean $\Delta f\text{CO}_2$ seasonal climatology from the SOCAT database; annual mean value is indicated by the diamond ($-4.1\mu\text{atm}$). (b) Map of annual $\Delta f\text{CO}_2$ climatology.

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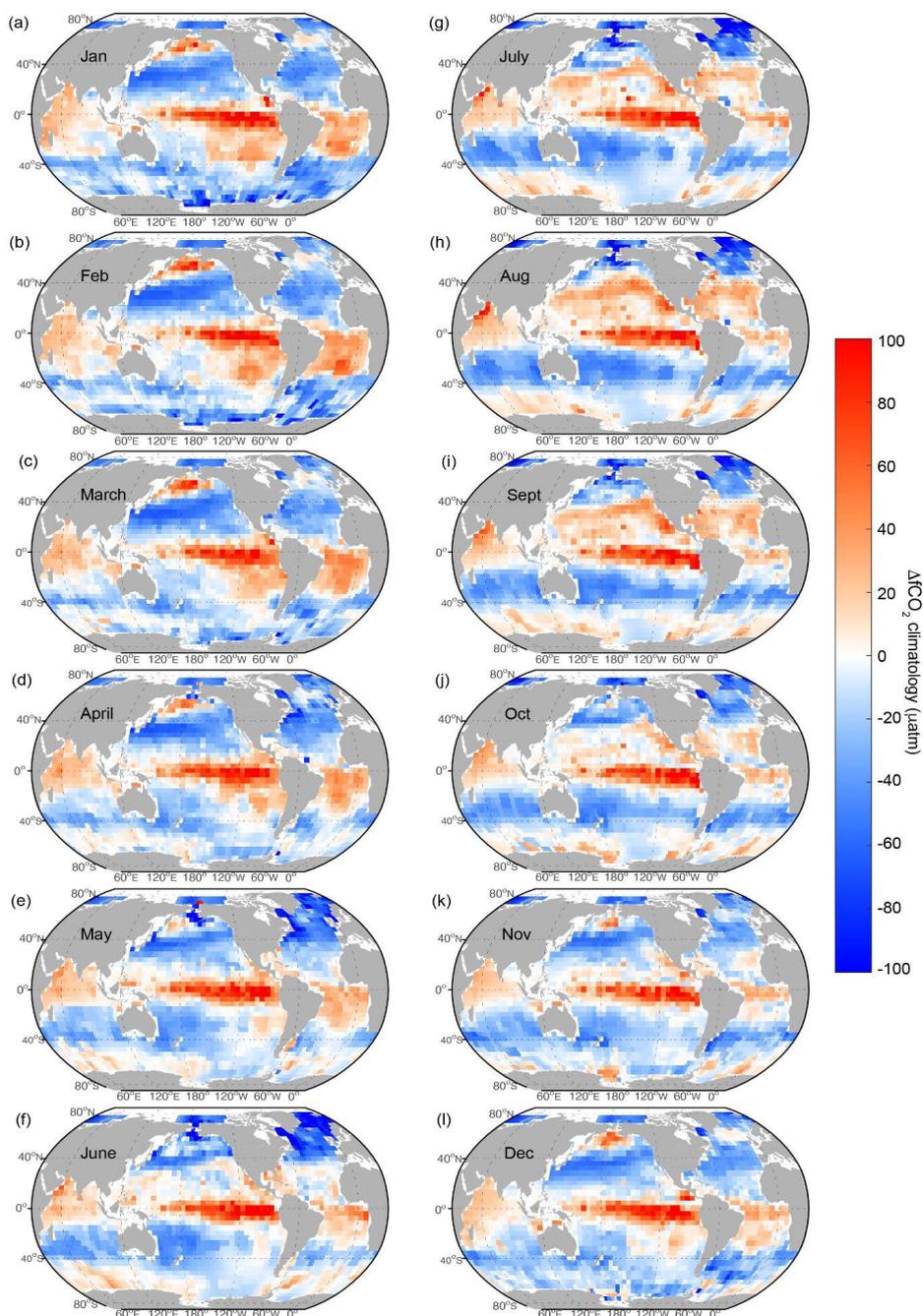


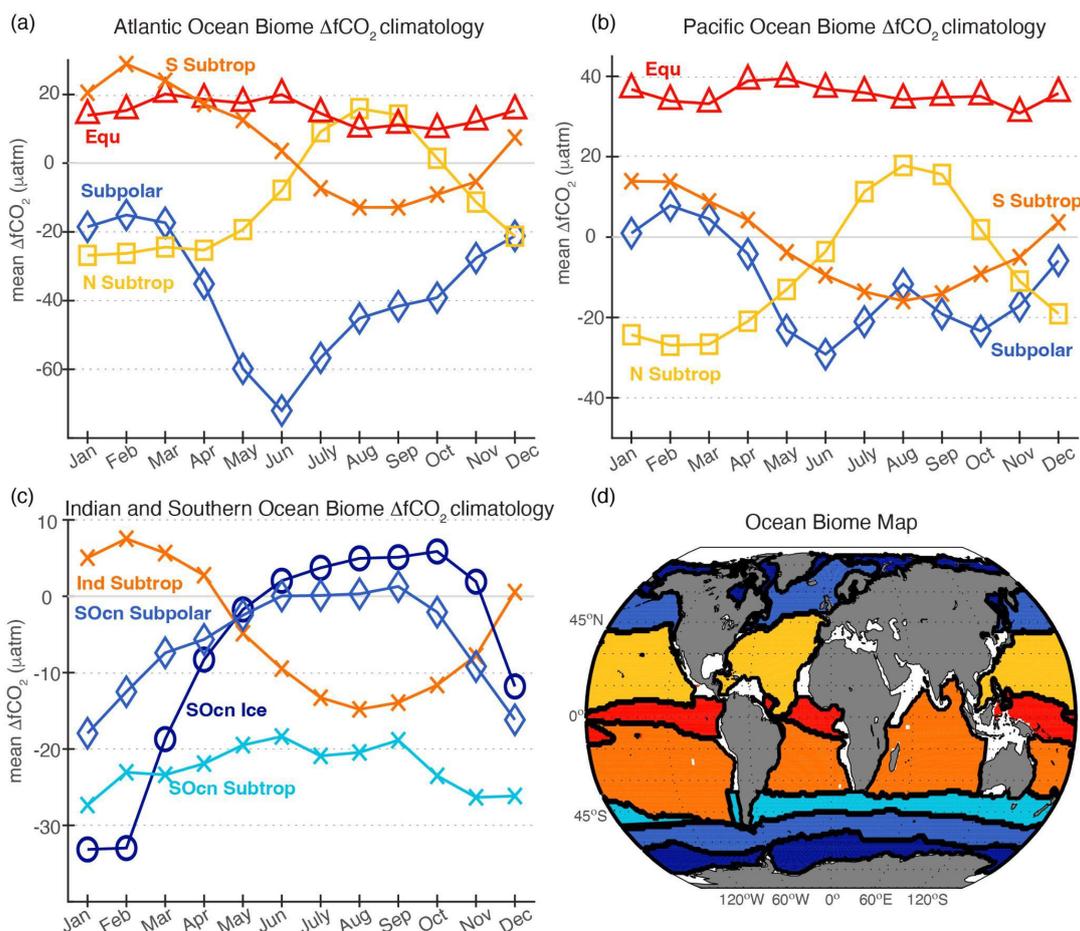
Figure 3: Monthly mean values for sea–air $\Delta f\text{CO}_2$. Warm colors indicate positive $\Delta f\text{CO}_2$ (ocean is greater than atmospheric CO_2), white indicates near zero $\Delta f\text{CO}_2$, and cool colors indicate negative $\Delta f\text{CO}_2$ (ocean CO_2 is lower than the atmosphere).



380 4.1.2 Regional

To show the seasonal changes in $\Delta f\text{CO}_2$ more clearly it is valuable to consider the patterns exhibited over consistent biogeochemical regions around the globe. For this analysis we utilize the biomes of Fay & McKinley (2014), but for simplicity we combine the seasonally stratified and permanently stratified subtropical biomes into one region in the Northern Hemisphere (referred to simply as subtropical in this manuscript). Monthly climatologies for each of the biomes are shown (Figure 4) in addition to the gridscale maps for each climatological month which allows for further regional interpretation (Figure 3).

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Figure 4: Monthly climatologies of $\Delta f\text{CO}_2$ for each regional ocean biome in the (a) Atlantic, (b) Pacific, (c) Indian and Southern Ocean basins. (d) Map of regional biomes. Colors of curves correspond to regions on the map in (d) with labels in matching colored text. Note that the y-axis varies between subplots.



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The equatorial regions of the Pacific and Atlantic oceans have positive $\Delta f\text{CO}_2$ values throughout the annual cycle and little seasonal variability. This indicates that the area is a source of CO_2 to the atmosphere year round. The equatorial Pacific (Figure 4b) has the highest positive $\Delta f\text{CO}_2$ values (annual mean = $35.4 \mu\text{atm}$), followed by the tropical Atlantic (Figure 4a, annual mean = $14.8 \mu\text{atm}$).

The subtropical biomes, representing the temperate North and South Atlantic and Pacific basins exhibit large seasonal $\Delta f\text{CO}_2$ cycles which change sign throughout the year. Here, the $\Delta f\text{CO}_2$ cycle is largely temperature driven; positive $\Delta f\text{CO}_2$ in warm summer months and negative values in colder winter months reflecting the dominance of seasonal temperature changes on the cycles of ocean $f\text{CO}_2$ in these regions. The seasonal amplitude for the subtropical North Pacific is $44.7 \mu\text{atm}$, and is slightly larger than the seasonal amplitude in the subtropical North Atlantic ($42.7 \mu\text{atm}$). Since the mean seasonal amplitudes for SST are quite similar in these two ocean basins, with the Atlantic having a slightly larger seasonal change in surface temperature (4.4°C in Pacific and 5.0°C in Atlantic, not shown), the difference in $\Delta f\text{CO}_2$ amplitudes between the Pacific and Atlantic subtropical regions cannot be attributed solely to SST, and may reflect differences in biogeochemical cycling between these two basins.

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Seasonal changes in the northern subtropical oceans are roughly six months out of phase from the southern subtropical biomes. The South Pacific subtropical biome has a seasonal amplitude of $29.8 \mu\text{atm}$ which is nearly $15 \mu\text{atm}$ smaller than that of the North Pacific subtropical biome. In contrast, the seasonal amplitude of the South Atlantic subtropical basin is just $1 \mu\text{atm}$ smaller than its counterpart in the North Atlantic (the South Atlantic subtropical amplitude is $41.7 \mu\text{atm}$). The Indian Ocean subtropical biome which encompasses most of the Indian Ocean, both above and below the Equator, has a smaller $\Delta f\text{CO}_2$ amplitude ($22.3 \mu\text{atm}$), but the phasing matches well with both the South Pacific and South Atlantic subtropical biomes, with peak (positive) $\Delta f\text{CO}_2$ values in February and the lowest values in August. The smaller $\Delta f\text{CO}_2$ seasonal amplitudes in the Indian and South Pacific subtropical basins are partially attributable to lower SST variability in these regions (SST seasonal cycle amplitudes are 4.0°C and 3.0°C in the South Pacific and Indian subtropics, respectively, compared to 4.6°C in the subtropical South Atlantic). However, it is likely that both differences in spatiotemporal patterns of primary productivity and undersampling in the South Pacific and Indian subtropics (Figure 1) also contribute to differences in $\Delta f\text{CO}_2$ seasonal amplitudes between these basins.

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435 The timing of the $\Delta f\text{CO}_2$ trough in the subpolar regions is opposite that of the subtropical
North Pacific and Atlantic basins. Strongly negative $\Delta f\text{CO}_2$ in the spring and summer
months is due to the effects of intense biological drawdown which quickly and
dramatically lowers carbon levels in the subpolar surface ocean with the onset of the
growing season. Biological productivity and strong spring/summer stratification result in
440 subpolar seasonal cycles that are roughly four to six months out of phase compared to
adjacent subtropical regions. In the Atlantic subpolar biome, $\Delta f\text{CO}_2$ values are
consistently below zero throughout the annual cycle (a maximum of $-15.1 \mu\text{atm}$ occurs
in February). In the subpolar North Pacific basin, positive $\Delta f\text{CO}_2$ values are present over
the boreal winter (Jan-March) before biological drawdown associated with the spring
bloom lowers the $\Delta f\text{CO}_2$ values back below zero for the remainder of the year. The
445 spring drawdown is weaker in the subpolar North Pacific compared to the subpolar
North Atlantic.

Figure 4c displays the seasonal cycle for the Southern Ocean biomes including the
seasonal ice biome, the subpolar region, and the seasonally stratified subtropical region
450 of the Southern Hemisphere. The higher latitude subtropical region has negative $\Delta f\text{CO}_2$
values throughout the year and a relatively small seasonal $\Delta f\text{CO}_2$ amplitude compared
to the more expansive South Atlantic, South Pacific and Indian subtropical basins to the
north. The mean $\Delta f\text{CO}_2$ and seasonal amplitude for the Southern Ocean subtropical
region is $-22.5 \mu\text{atm}$ and $9.0 \mu\text{atm}$, respectively.

455 The Southern Ocean subpolar and ice biomes both have relatively strong seasonal
cycles, reaching maximums of $\Delta f\text{CO}_2$ near and slightly above zero, respectively, during
the late austral winter and early austral spring (Figure 4c). This positive peak during July
through October in the Southern Ocean seasonal ice zone is influenced by under-ice
460 vertical mixing with deep waters that contain excess carbon and nutrients. During the
austral spring and summer months, intense phytoplankton blooms occur near and
around the edges of the retreating sea ice in the seasonal ice zone and within the
subpolar region. These blooms cause dramatic drops in $\Delta f\text{CO}_2$ values at the end of the
calendar year (Oct-Dec). Limited sampling and smoothing from the interpolation method
465 fail to capture the high spatiotemporal variability that characterizes this highly dynamic
region.

4.2 Net air-sea CO_2 flux

The mean climatological global air-sea CO_2 flux estimate using the SOCAT database is
 $-1.79 \text{ PgC yr}^{-1}$, indicating uptake of carbon by the ocean. This is a slightly greater flux
470 into the ocean than the direct estimate from the previous version of the climatology (T-
2009), which reported a direct estimated global mean flux of -1.4 PgC yr^{-1} for the year
2000. For the uncertainty in global ocean-atmosphere CO_2 flux, we use the value



reported by Wanninkhof et al. (2013) who followed the same approach as T-2009. This approach combines uncertainty contributions from the spatial and temporal sampling of
475 $\Delta f\text{CO}_2$ (± 0.18 and ± 0.5 PgC yr^{-1} , respectively) as well as smaller contributions for the uncertainty in the gas exchange parameterization (± 0.2 PgC yr^{-1}), wind (± 0.15 PgC yr^{-1}) and riverine carbon (± 0.2 PgC yr^{-1}).

4.2.1 Mean annual distribution

The near-global mean flux estimate presented here represents 90% of the surface area of the global ocean. Specifically, this estimate excludes the coastal ocean and areas of
480 of the high latitude seas. We have chosen to present values without any adjustment to account for missing areas and acknowledge that this analysis represents a near-global estimate. Suggested methods for filling missing ocean areas in such reconstructions are available in Fay & Gregor et al. (2021) but are not implemented in this climatology to
485 remain consistent with its previous versions. An estimate of the flux from regions missing from this product can be obtained by using the full-coverage $p\text{CO}_2$ climatology combining open and coastal oceans (Landschützer et al. 2020). Considering only grid cells in the Landschützer et al. (2020) product that are missing in this climatology, we estimate an annual average coastal and high latitude flux of -0.38 PgC yr^{-1} . The flux
490 varies through the seasonal cycle, ranging from -0.43 to -0.31 PgC yr^{-1} . This quantity is not included in the climatological estimate presented here.

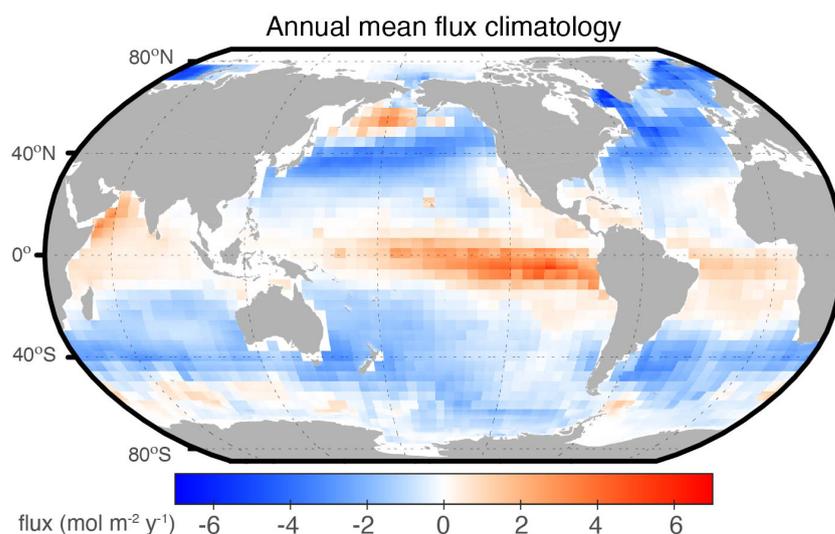


Figure 5: Annual mean CO_2 flux calculated from the SOCAT database. Flux is
495 calculated using the SeaFlux method using the mean of three wind speed reanalysis products. Warm and cool colors indicate regions of carbon efflux and uptake, respectively. The near-global mean flux is -1.79 PgC yr^{-1} .



500 Figure 5 shows the climatological mean annual sea–air CO₂ flux (mol m⁻² yr⁻¹) and maps
of two seasons (DJF and JJA) are displayed in Figure 6. The equatorial Pacific is the
most prominent atmospheric CO₂ source region, with a seasonally persistent sea-to-air
flux of 0.35 PgC yr⁻¹. When combined with the equatorial Atlantic region, the tropical belt
emits an annual mean of 0.39 PgC yr⁻¹ to the atmosphere. Adjacent to this tropical
505 efflux zone, are vast expanses of seasonally-variable flux patterns. The subtropical
basins in both hemispheres act as CO₂ sinks in the cooler months and transition to
regions of neutral or small CO₂ sources during the warmer months. At higher subtropical
latitudes, strong winds and relatively low ocean fCO₂ occur along the subtropical
convergence zone where the cooled subtropical gyre waters with low fCO₂ meet the
510 subpolar waters with biologically-lowered fCO₂.

The Northern Hemisphere mid and high latitude regions represent a smaller sink (-0.63
PgC yr⁻¹) compared to the corresponding regions of the Southern Hemisphere (-0.91
PgC yr⁻¹) largely due to the overall greater surface area of the oceans in the Southern
515 Hemisphere (oceans south of 35S are 25% of total global ocean area while oceans
north of 35N are 15% of total ocean area). The dramatic influence of the expansive
Southern Hemisphere oceans is also demonstrated by the large flux in the Southern
Ocean subtropical region (-0.59 PgC yr⁻¹) that represents 8% of the global ocean
surface area.

520 Moving poleward, a strong sink (-0.27 PgC yr⁻¹) occurs in the North Atlantic subpolar
region which includes the Nordic Seas and the portions of the Arctic Ocean which
contain observations. This strong localized carbon sink is attributed to the import of low
anthropogenic waters at depth in the Gulf Stream that are exposed as mixed layers
525 deepen (Ridge & McKinley 2020), and large phytoplankton blooms in spring followed by
cooling in winter. In the Southern Ocean, annual mean CO₂ flux is heterogeneous and
relatively small in the seasonal ice zone due to the ice cover that reduces sea–air gas
transfer in winter. Additionally, the small annual flux values in the Southern Ocean
subpolar and ice regions (-0.21 PgC yr⁻¹ and -0.08 PgC yr⁻¹, respectively) are a result of
530 a cancellation of the seasonal source (winter) and sink (summer) fluxes.

4.2.2 Seasonal variation of sea–air CO₂ flux

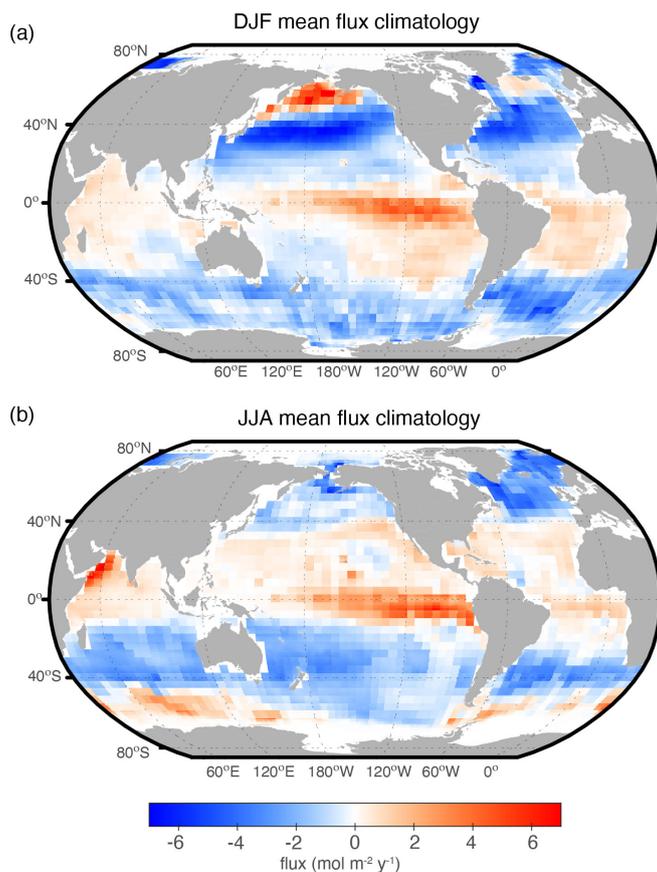
Seasonal variation in air-sea fluxes are clearly seen in the climatology (Figure 6) and
are attributed to a combination of effects including fluctuations in SSTs, biological
535 uptake of carbon dioxide, as well as mixing and wind speeds.



540 The seasonal variability of fluxes in higher latitudes of the subtropics in the Atlantic, Indian and Pacific Oceans cause an oscillation from neutral or weak sources of CO₂ to the atmosphere in the warmer seasons to strong CO₂ sinks in the cooler or winter seasons. Water cools as it is transported poleward by western boundary currents, allowing for carbon uptake (Ayers & Lozier 2012). In spring and summer, the biological drawdown of carbon increases CO₂ uptake by the ocean, partially offset by increases in fCO₂ due to warming.

545 Subtropical gyre regions also transition from weak sinks in the winter seasons to weak sources in the summer seasons, following the seasonal SST cycles and reflecting the dominance of temperature effects in controlling the seasonal variability in the fCO₂ and sea–air fluxes in oligotrophic gyres. In tropical low latitude regions, seasonality is generally smaller, however localized hot spots of high variability and large fluxes do

550 exist, such as in the northwestern Indian Ocean where the strong summer monsoon winds force upwelling of carbon rich subsurface waters and cause high gas transfer rates in this region (Chen et al. 1998, T-2002). The equatorial Pacific and Atlantic show little seasonal variability in CO₂ flux with a persistent efflux throughout the year.



555 Figure 6: Season sea–air CO₂ flux (molC m⁻² year⁻¹) climatology for (a) December, January, February (DJF) and (b) June, July, August (JJA). Positive values (warm colors) indicate sea-to-air fluxes (ocean efflux), and negative values (cool colors) indicate air-to-sea fluxes (ocean uptake).

560 In the Southern Ocean, there is a consistent region of moderate carbon source waters located in the Atlantic and Indian sector south of 45S latitude during the austral winter (Figure 6b). The source in the Southern Ocean region could be influenced by high fCO₂ waters from margins of the Antarctic sea-ice field given that the efflux values occur during the austral winter months (JJA). As the seasons transition to warmer
565 temperatures and the ice edge recedes, this region is impacted by high rates of photosynthesis causing fCO₂ drawdown, and resulting in a transition to moderate carbon sinks during the austral summer (DJF). Uncertainties are higher in the Southern Ocean region due to the limited number of observations, particularly in winter (Figure 1).



5. Discussion

570 It should be noted that there isn't one specific reference year for this release of the
 $\Delta f\text{CO}_2$ climatology as was the case for previous releases (e.g., the year 2000 reference
in T-2009). Instead, this climatology represents a multidecadal time period, beginning in
1980, with the majority of observations feeding into the climatology collected after 2010.
Therefore, while the $\Delta f\text{CO}_2$ climatology is not reported for a specific reference year, it is
575 most representative of the conditions over the past two decades. We note however that
the flux estimates given in this analysis are based on inputs from a single year, the year
2010, as described above in Section 3.3 (comparison of flux estimates using 2010
inputs and averages over several decades yield very similar results, also as described
in Section 3.3).

580

5.1 Comparison with T-2009 climatology

In the previous release of this climatology (T-2009), $p\text{CO}_2$ values were corrected to a
reference year of 2000 using a mean atmospheric CO_2 increase rate of $1.5 \mu\text{atm yr}^{-1}$
585 over the entire ocean area with the exception of the Bering Sea, where the observed
rate of $-1.2 \mu\text{atm yr}^{-1}$ was used. In our updated method described above (Section 3.1),
we eliminate the need to apply a normalization rate for observations and instead
calculate a $\Delta f\text{CO}_2$ value for each observation using a co-located concurrent atmospheric
 $f\text{CO}_2$ value. We note that T-2014 presents a more updated $p\text{CO}_2$ climatology than T-
590 2009; since T-2014 emphasizes climatologies for the other carbonate system variables
(pH, dissolved inorganic carbon, and total alkalinity) and omits estimation of fluxes, we
focus our comparison on values presented in T-2009.

Spatial differences between the climatology created from surface water $\Delta f\text{CO}_2$ values
595 using the methods discussed above and the approach of T-2009 (3 million
observations) are shown in Figure 7 for months February and August; months were
selected for consistency with past comparisons presented in T-2009 and T-2002. The
differences between this updated release and previous versions producing climatologies
for reference years 2000 (T-2009) and 1990 (T-1997) are unlikely to represent real
600 change in the oceans over time, but instead primarily reflect the impact of the greatly
expanded database as well as the use of the $\Delta f\text{CO}_2$ approach as opposed to the time
normalization method of T-2009.

The most significant regional differences between this updated version and that of T-
605 2009 are observed over the subpolar North Atlantic, the subtropical Southeast Pacific
and portions of the Southern Ocean. In the North Atlantic, differences between versions



are largest in the boreal winter (February map, top panel of Figure 7) when the updated climatology exhibits less uptake compared to the T-2009 version (positive values on the map indicate more negative values in the T-2009 version). Differences in the North
610 Atlantic can be at least partially attributed to the much greater availability of observations in this region between the two databases (Figure 1, Supplementary Figure 1). This is discussed in further detail in the supplementary text.

The Southeast Pacific is an area with very limited observations but where a few key
615 datasets have been included in the SOCAT database since 2010. Figure 1 shows that despite these recent additions to SOCAT, there are still only a few observations covering this region. Comparison of Figure 3 in this study with Figure 9 of T-2009 suggests that the additional datasets in SOCATv2022 result in a more defined seasonal cycle for $f\text{CO}_2$ in the subtropical Southeast Pacific in the current release. Specifically,
620 the map of differences shown in Figure 7 shows that this region is a greater source of carbon to the atmosphere in austral summer and a greater sink in austral winter compared to T-2009 (compare also Figure 6 of this study with Figure 15 of T-2009). In contrast, monthly maps included in Figure 3 of T-2009 show little seasonal contrast in $\Delta p\text{CO}_2$ likely due to a lack of observations.

625 In the Southern Ocean region during austral winter (August), $\Delta f\text{CO}_2$ values are more negative in the current version compared to T-2009. The Southern Ocean is also a region of limited data availability particularly in austral winter but one where SOCATv2022 also includes several datasets added in the past decade that have an
630 outsized influence on the resulting climatology.

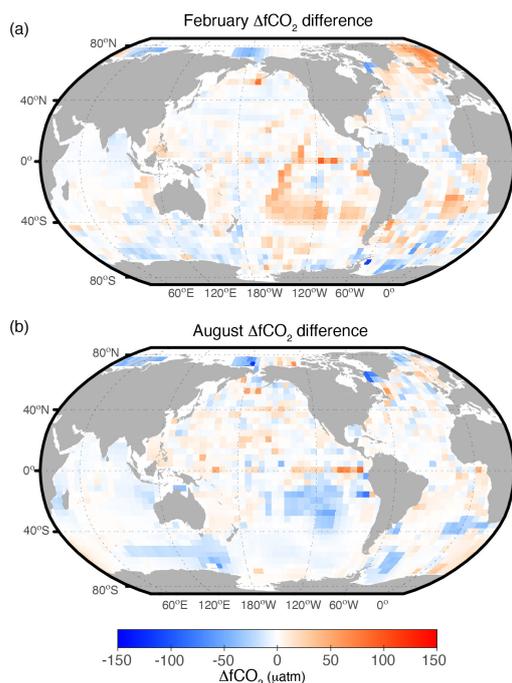


Figure 7: Difference maps for the surface water $\Delta f\text{CO}_2$ climatology produced by this study using SOCAT and that from T-2009 (maps show this study *minus* T-2009) in (a) February and (b) August. In T-2009, the delta CO_2 values are reported as $p\text{CO}_2$ and here we are using $f\text{CO}_2$.

5.2 Comparison to other flux estimates

The estimate presented here for annual global mean carbon flux ($-1.79 \text{ PgC yr}^{-1}$) represents slightly less uptake than reported by other studies, but given uncertainties, as well as differing timeframes, spatial coverage and gap-filling methodologies, our estimate compares closely with current estimates similarly based on observed surface ocean $f\text{CO}_2$.

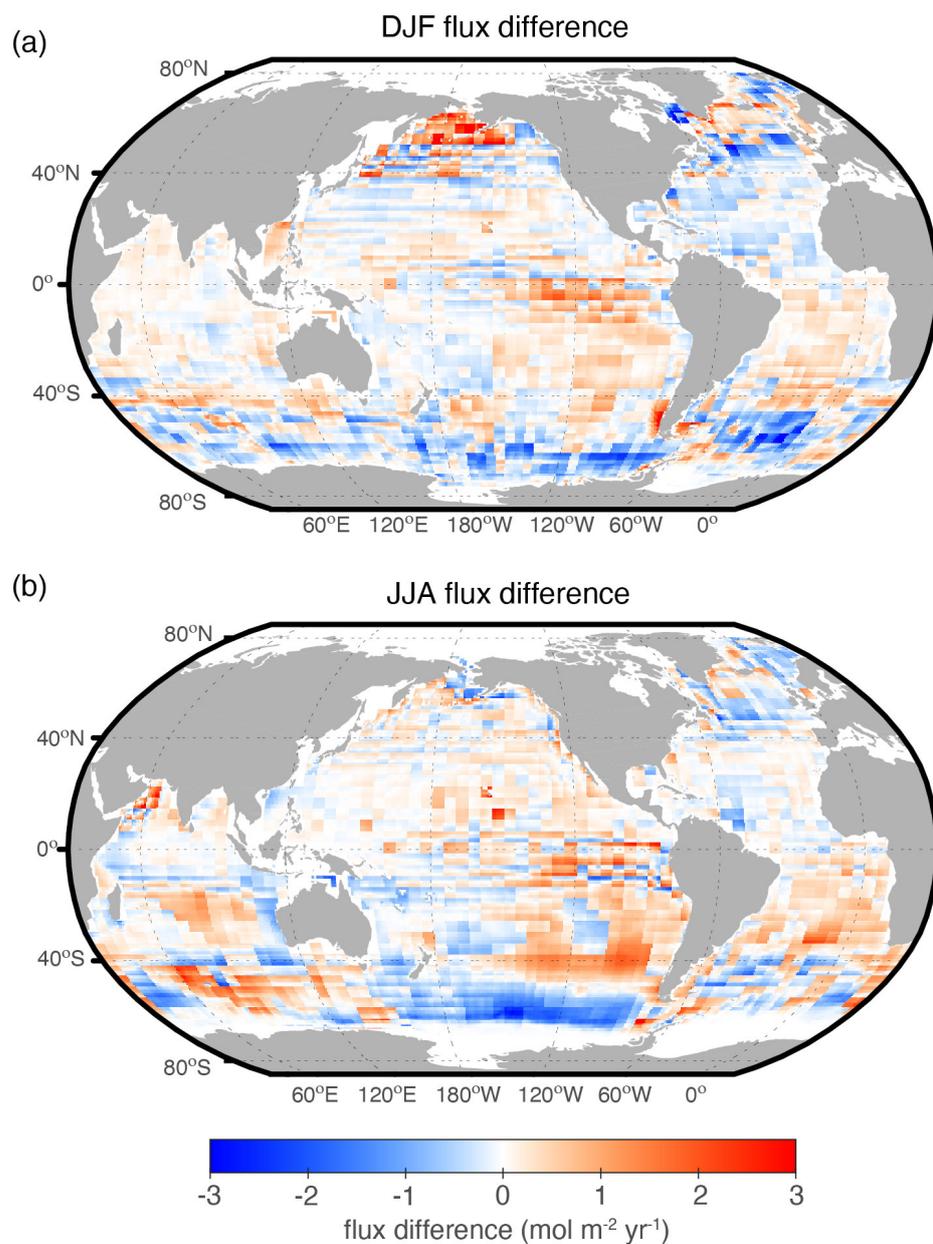
To compare our new climatological estimate of contemporary air–sea net flux from surface ocean $f\text{CO}_2$ with estimates of the anthropogenic carbon flux from interior data (e.g., Gruber et al. 2019) or estimates from global ocean biogeochemical models (e.g., Friedlingstein et al. 2022; Hauck et al. 2020), it is necessary to account for the outgassing of natural carbon supplied to the ocean by rivers. This riverine estimate varies significantly in magnitude between studies and continues to be a research focus for the ocean carbon community. Therefore, we focus on comparisons between our



655 climatological estimate and a mean carbon flux estimate from an ensemble of
observation-based pCO₂ products included in the SeaFlux product (Fay & Gregor et al.
2021).

660 The SeaFlux products span the years 1985-2020, are all similarly based on the SOCAT
database, but employ various methods of machine learning and interpolation to produce
full coverage ocean carbon maps. For this comparison, a climatology of the SeaFlux
product is produced and fluxes are calculated in the same manner as for the climatology
presented here. Following this approach, the SeaFlux climatology ensemble yields a
665 global mean flux of -2.10 PgC yr⁻¹ which represents a slightly larger flux into the ocean
than that produced by the updated climatology (-1.79 PgC yr⁻¹). The differences in
global flux can be attributed to the true global coverage of the SeaFlux product relative
to the 90% global ocean coverage of this study. As mentioned above in Section 4.2, an
estimate of missing coastal and high latitude fluxes increases the ocean carbon uptake
estimate for this climatology by roughly 0.38 PgC yr⁻¹. Adding this additional flux brings
our analysis within 0.1 PgC yr⁻¹ of the SeaFlux ensemble (-2.17 PgC yr⁻¹ versus -2.10
PgC yr⁻¹ for this analysis versus the SeaFlux estimate, respectively).

670 Spatially, comparison of the SeaFlux ensemble of products to our climatology shows
strong agreement in overall patterns but significant differences in the mid and high
latitude Southern Hemisphere oceans (Figure 8). Gloege et al. (2021) analyzed a
machine learning method's ability to reconstruct global carbon fluxes from available
675 observations using a testbed approach and found the highest flux bias in the Southern
Ocean regions as well as an overestimation of decadal variability in this region. Given
the limited availability of year-round observations at high Southern Hemisphere
latitudes, and the resulting reliance on various gap-filling approaches, it is not surprising
that the largest differences between the climatology presented here and the SeaFlux
680 ensemble emerges in this region. Significant differences between these climatologies
are also evident in the high latitude North Pacific and North Atlantic, specifically in the
boreal winter season (Figure 8a). Again, a lack of observations in these regions during
the winter season (Figure 1) likely accounts for much of this disagreement, with more
reliance on the interpolation methods used by each method. Machine learning methods
685 that utilize proxy variables to estimate pCO₂ in unsampled areas, such as those in the
SeaFlux product, often rely on relationships derived from better-observed areas that are
deemed similar in biogeochemical characteristics and it is likely that the mechanisms at
these high latitude locations are not accurately captured by any available interpolation
methods. This is also a current focus of research for the ocean carbon observing
690 community.



695 Figure 8: Difference map for carbon fluxes ($\text{mol m}^{-2} \text{yr}^{-1}$) calculated from this ΔfCO_2 climatology and fluxes reported by an ensemble of observation-based products included in the SeaFlux product for (a) boreal winter, DJF and (b) boreal summer, JJA. Map shows the difference defined as this study *minus* SeaFlux.



6. Conclusion

An updated climatological mean distribution for $\Delta f\text{CO}_2$ (surface water minus atmosphere) using the methods of T-2009 is presented. This climatology is based on approximately seven times more open ocean observations from the SOCATv2022 database (over 21 million values, spanning years 1980-2021) compared to the 3 million data values used in T-2009 (and more than three times the approximately 6.5 million observations used in T-2014). In this updated climatology, observations made during El Niño periods over the equatorial Pacific are included, unlike climatologies presented by T-1997, T-2002, T-2009 and T-2014. In addition to coastal waters, the highest latitudes of the Arctic and the Mediterranean Sea are also excluded as in all previous LDEO climatologies.

To develop a climatology from data collected over multiple decades during which $f\text{CO}_2$ experienced a large secular trend, we calculate $\Delta f\text{CO}_2$ values for each day and grid cell before collapsing all available data onto one climatological year. This method follows the assumption made in previous iterations of this climatology (T-1997) that the ocean surface carbon value follows the rate of increase in the atmospheric $f\text{CO}_2$, such that $\Delta f\text{CO}_2$ is constant over time. Observed $\Delta f\text{CO}_2$ is then interpolated in space-time using a lateral two-dimensional diffusion–advection equation on a $4^\circ \times 5^\circ$ grid (Takahashi et al. 1995, T-1997, T-2002, T-2009, T-2014). Monthly mean $\Delta f\text{CO}_2$ values for each pixel, downscaled to $1^\circ \times 1^\circ$ resolution, are presented here. Net sea-air CO_2 flux is computed using the pySeaFlux package, following the protocol presented in Fay & Gregor et al. (2021).

Regional mean $\Delta f\text{CO}_2$ values vary greatly among the ocean basins (Figures 3 and 4). The high-latitude North Atlantic is the most intense CO_2 sink per unit area as a result of both highly negative $\Delta f\text{CO}_2$ (Figure 3) and strong winds. This is also the region with the largest differences between the climatologies created with previous versions of the LDEO database and this version based on the SOCATv2022 database (Figures 3 and 7). Globally, differences are due to the greater abundance of observations over all regions of the global oceans in the SOCAT database, but particularly the greater seasonal coverage in the Southern Hemisphere oceans and subpolar North Atlantic (Figure 1, Supplementary Figure 1).

The annual mean uptake flux for the global open-ocean region is estimated to be $-1.79 \text{ PgC yr}^{-1}$ for 1980-2021 (Figures 5 and 6). Of the major ocean basins, the Southern Hemisphere oceans are the largest CO_2 sink, taking up 1.19 PgC yr^{-1} , while the Northern Hemisphere oceans (subtropical and subpolar biomes) take up 1.04 PgC yr^{-1} .



The equatorial oceans act as the only year-round region of carbon efflux to the atmosphere with and emit 0.35 PgC yr^{-1} to the atmosphere.

740 While over a million new shipboard $f\text{CO}_2^{\text{oce}}$ observations have been made each year in
the global oceans for the past two decades, there has been a notable decline in the
observations submitted to SOCAT since 2017. This decline is due in part to the
disruption of the COVID19 pandemic, but also reflects a shift away from shipboard
745 $p\text{CO}_2$ measurements. Given the lack of alternative approaches with which to assess
spatial and temporal variability in air-sea CO_2 flux and the need for high accuracy
shipboard measurements (accuracy of $<2 \mu\text{atm}$) to characterize most regions of the
global oceans, this trend to fewer observations is highly detrimental to carbon cycle
research. This is true both in regard to monitoring of ocean carbon uptake and to
750 monitoring of more uncertain fluxes such as that between the atmosphere and terrestrial
biosphere since the high uncertainty of independent terrestrial estimates necessitates
the monitoring of this flux by difference.

Data Availability

The updated climatology is available via The National Center for Environmental
Information (NCEI) at [https://www.ncei.noaa.gov/access/metadata/landing-](https://www.ncei.noaa.gov/access/metadata/landing-page/bin/iso?id=gov.noaa.nodc:028225)
755 [page/bin/iso?id=gov.noaa.nodc:028225](https://www.ncei.noaa.gov/access/metadata/landing-page/bin/iso?id=gov.noaa.nodc:028225), doi.org/10.25921/295g-sn13, (Fay et al. 2023).

Author contribution

ARF and DRM conducted the analysis and prepared the manuscript. GAM, RW, CW,
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760 as contributed to manuscript.

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