Biological carbon pump affected by CO₂ uptake in the Benguela Upwelling System

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Abstract

Eastern Boundary Upwelling Systems (EBUS) are well-known for their high productivity and fishery yields. However, being scarcely sampled and poorly represented in global models, their role as CO₂ sources and sinks to the atmosphere remains elusive. Here, we present a compilation of shipboard measurements over the past two decades, showing how the Benguela Upwelling System (BUS) in the southeast Atlantic Ocean acts as a CO₂ source in the north and CO₂ sink in the south. Surface warming of upwelled waters increases the partial pressure of CO₂ (pCO₂) and outgassing in both regions, but in the south, the biologically-mediated drawdown of CO₂ exceeds this warming effect. Here, the biological carbon pump owes its stronger impact on pCO₂ to higher shares of upwelling source waters carrying preformed nutrients supplied from the Southern Ocean. Their formation increases pCO₂ in surface waters and counteracts human-induced invasion of CO₂ in the Southern Ocean. However, their utilization in the BUS compensates for over 20% of the CO₂ loss occurring in the Atlantic sector of the Southern Ocean. This emphasizes the role of the BUS as key to improve our understanding of the ocean's response to climate change and the future evolution of CO₂ in the atmosphere.

Main

Eastern Boundary Upwelling Systems (EBUS) are among the most productive regions in the ocean and contribute 11% to global new production, which refers to biomass largely produced on the basis of nutrients introduced via upwelling and vertical mixing from the deep dark ocean into surface waters¹⁻³. The associated assimilation of CO₂ through the generation of biomass and its transfer to the deep sea is an integral part of the biological carbon pump^{4,5} which reduces atmospheric CO₂ concentrations through the storage of CO₂ as dissolved inorganic carbon (DIC) in the deep ocean⁵. Even though it is widely assumed that the biological carbon pump responds to climate change, the magnitude and even the direction of change is still unknown^{6,7}. However, the amount of DIC kept by the biological carbon pump is assumed to be linearly related to the inventory of regenerated nutrients, which are released during the remineralization of biomass in the deep ocean⁸. They stand in contrast to biologically unused so-called

preformed nutrients, whose formation represents a leakage through which the biological carbon pump
 losses CO_2 and turned e.g. the Southern Ocean south of 44°S into a natural CO_2 source to the
 atmosphere^{9,10}. This leakage evolves when nutrients that upwell along with DIC at the Antarctic
 Divergence¹¹ are not fully utilized by biological production due to light and iron limitation^{12,13}.
 Nowadays, rising CO_2 concentrations in the atmosphere revert the air sea fluxes of CO_2 and converted the
 Southern Ocean into a key site for anthropogenic CO_2 uptake with a current estimate of approximately
 $740 \pm 290 \text{ Tg C year}^{-1}$ ^{14,15}. Due to the counteracting effect of natural CO_2 released from the biological
 carbon pump¹² of about $400 \pm 180 \text{ Tg C year}^{-1}$ ^{14,15}, this results in a CO_2 invasion of $380 \pm 115 \text{ Tg C year}^{-1}$
 between 1990 and 2009¹⁵⁻¹⁷. However, approximately 34% ($\sim 110 \text{ Tg C year}^{-1}$) of the natural CO_2 release
 occurs within the Atlantic sector of the Southern Ocean between 44° and 58°S ^{14,15} and the resulting
 preformed nutrients are transported northwards and subducted beneath warmer and lighter subtropical
 water masses¹⁸⁻²⁰. Loaded with preformed nutrients, these mode waters support primary productivity in
 upwelling systems at lower latitudes and thus could potentially restore the CO_2 uptake efficiency of the
 biological carbon pump. Although EBUS could act as regional CO_2 sinks through the utilization of
 preformed nutrients and low sea water temperatures which increase the solubility of CO_2 ²¹⁻²³, global
 models suggest that upwelling systems (in particular ones at lower latitudes) act as net CO_2 sources to the
 atmosphere^{24,25,26}. This was also found to be the case in a modelling study of the Benguela Upwelling
 System (BUS), which is located in the southeast Atlantic Ocean (Fig. 1) and considered the most
 productive of all EBUS²⁷. The study however suffered from a model that poorly represented the BUS, and
 an unclearly defined upwelling region²⁶. Other studies within the BUS hardly improve these conditions as
 indicated by estimated CO_2 fluxes ranging from -5.1 ^{25,24} Tg C year^{-1} to $1.54 \text{ Tg C year}^{-1}$ ²⁸. It thus remains
 elusive whether the BUS is a net CO_2 sink or CO_2 source to the atmosphere. Similar to the Southern
 Ocean, the BUS also suffers from a sparsity of data and model outcomes^{25,26} were constrained by pCO_2
 data (partial pressure of CO_2) from the Surface Ocean CO_2 Atlas (SOCAT)²⁹. So far this data product
 misses coverage particularly over the northern BUS shelf region, thus curtailing estimates of air-sea gas
 exchange for the BUS (Fig. 1).

Here, we address this problem by presenting a compilation of shipboard pCO₂ data for the BUS from the extended SOCAT v2020²⁹ data base that includes data gathered from 14 cruises throughout the BUS from 2008 to 2019 (see Supplementary Table 1). This revised data set allows for a seasonal-based examination of regional air-sea fluxes of CO₂ across the BUS and underlying mechanisms affecting these fluxes, which indicate that the biological carbon pump in the BUS serves as a globally significant CO₂ sink.

Sea surface pCO₂ characteristics

In order to overcome the problem of unclearly defined upwelling regions that led to strong discrepancies in the definition of the BUSs offshore boundary^{25,30,31}, we specified the upwelling zone's spatial extent by considering the average cross-shelf distribution of sea surface pCO₂ (Fig. 2a,b) for both the northern and southern subsystems (NBUS, SBUS). In general, upwelling systems exhibit highest pCO₂ in the nearshore region due to the upwelling of carbon-rich waters, and a decreasing trend offshore due to degassing and the biologically-mediated carbon uptake within the offshore-advecting upwelled water^{30,32,33}. By assuming that a decrease in pCO₂ variability marks a decreasing influence of upwelling on pCO₂, we determined the upwelling zone's boundary as the distance from shore when the standard deviation (s.d.) decreases persistently by ~85% to below $\pm 35.5 \mu\text{atm}$ in the NBUS and $\pm 29.5 \mu\text{atm}$ in the SBUS.

In the NBUS, this results in an upwelling boundary at 340 km offshore, which lies within the range of other studies that considered an offshore extension from 300 to 800 km^{26,27,34,35}. In the SBUS, the decreasing variability in pCO₂ suggests a closer boundary at 200 km distance to shore. In comparison to the NBUS, the much lower intensity of upwelling observed in the SBUS³⁶ may explain the discrepancy observed in the offshore extent of elevated pCO₂ values between the subsystems. Considering additionally the latitudinal extent of the SBUS (26°S to 34°S) and NBUS (16°S to 26°S) results in upwelling areas of 177.600 km² and 377.400 km², respectively.

To estimate the seasonal variability, pCO₂ data were averaged across the mean offshore extent and latitudes of the upwelling areas for the austral spring and summer (September-March), and austral autumn and winter seasons (April-August) (Fig. 2c). The resulting plot of pCO₂ variability with latitude indicates

a seasonal influence in the northern region between $\sim 17^{\circ}\text{S}$ and $\sim 21^{\circ}\text{S}$ which diminishes towards the south. The seasonality observed in the north corresponds to seasonal variations in upwelling intensities, which for this region are strongest during the winter³⁷. However, when averaged across the whole NBUS, the $\sim 15\%$ difference in seasonal means and high s.d. in pCO_2 between spring and summer ($485.7 \pm 113 \mu\text{atm}$), and autumn and winter ($556.03 \pm 126 \mu\text{atm}$), imply that the upwelling-related seasonality is only weakly pronounced. In the SBUS, seasonal mean pCO_2 s of $378.15 \pm 65 \mu\text{atm}$ (spring, summer) and $381.91 \pm 36 \mu\text{atm}$ (autumn, winter) do not reflect the seasonality of upwelling intensities, which are strongest during the summer³⁸. This could be attributed to a lower data coverage in SBUS as compared to NBUS (Supplementary Fig. 1) or to the function of the SBUS as CO_2 source and sink as indicated by pCO_2 concentrations above and below those in the atmosphere at the coast and further offshore (Fig. 2b). Relative to the mean atmospheric pCO_2 of our reference year 2020 ($\sim 414 \mu\text{atm}$), the seasonal as well as annual mean pCO_2 s of $487.6 \pm 2.9 \mu\text{atm}$ for the NBUS and $379.2 \pm 1.74 \mu\text{atm}$ for the SBUS reveal the opposing character of the two subsystems as a CO_2 source and sink, respectively.

Air-sea CO_2 flux estimates

In contrast to pCO_2 , the air-sea CO_2 fluxes reveal a pronounced seasonality in both systems as being more than twice as high during upwelling than during non-upwelling seasons. The intensification of wind during the upwelling season is considered the primary driver of this difference which, according to the opposing signs, strengthens the CO_2 source and sink functions in the NBUS and SBUS, respectively (Table 1).

Integrating the annual mean fluxes over the upwelling area results in a CO_2 emission of $14.98 \text{ Tg C year}^{-1}$ in the NBUS and a CO_2 uptake of $-3.37 \text{ Tg C year}^{-1}$ in the SBUS. These area-integrated CO_2 fluxes correspond with those given in ref.³¹ for the NBUS ($11.5 \text{ Tg C year}^{-1}$) and SBUS ($-1.4 \text{ Tg C year}^{-1}$), respectively, but are about 30% and 140% higher, mainly due to the use of a smaller area in the ref.³¹ study. Similar to these data-based estimates, modelled carbon fluxes also indicate a CO_2 sink region in the

south and source region in the north²⁶, while modelled CO₂ fluxes from the entire BUS between 18°S to 28°S amount to 2.25 mol C m⁻² year⁻¹ on average over the time period between 1982 and 2015. Comparing our measurements with modelled data by recalculating CO₂ fluxes for the same region with an offshore boundary at 800 km, results in a weaker annual CO₂ source into the atmosphere of 1.19 mol C m⁻² year⁻¹. This discrepancy seems to be caused mainly by a poleward misplacement of the outgassing cell in the applied model framework²⁶, which underestimates the CO₂ sink behavior of the SBUS.

Preformed nutrients as a driver of regional variability in sea surface pCO₂

As seen for the Oregon upwelling system, the utilization of preformed nitrate (N_{pref}), with concentrations of up to 12 μmol kg⁻¹, turned the upwelling area into a CO₂ sink^{21,23}, while the impact of N_{pref} with concentrations of around zero μmol kg⁻¹ were too low to support the sink functionality of the Peruvian upwelling system²². In addition, the inaccessibility of phosphate (P) through lack of fixed nitrogen and the effect of surface warming on the CO₂ solubility denote further contributing factors turning the Peruvian upwelling system into a CO₂ source²². Based on the biogeochemical analysis of water samples during our cruise to the BUS (see Methods section), South Atlantic Central Water (SACW), being the upwelling source water mass of the NBUS, exhibits preformed nutrient concentrations of 5.86 ± 0.36 μmol N_{pref} kg⁻¹ and 0.54 ± 0.01 μmol P_{pref} kg⁻¹. In contrast, Eastern South Atlantic Central Water (ESACW) that dominates upwelling within the SBUS, holds higher preformed nutrient inventories of 8.61 ± 0.45 μmol kg⁻¹ N_{pref} and 0.75 ± 0.02 μmol P_{pref} kg⁻¹.

In order to quantify the possible impact of preformed nutrients on the air-sea gas exchange, we calculated pCO₂ on the basis of upwelling source water mass characteristics as obtained from our data (see Methods section) and sea surface conditions (Table 2) by applying the CO2SYS program^{39,40}, which is commonly used for analysing the marine carbonate system³⁹. Three different scenarios were taken into account to demonstrate that the chosen source water mass characteristics comply with the obtained pCO₂ climatology.

In an initial step, we estimated $p\text{CO}_2$ of SACW and ESACW when being upwelled by using the respective in situ source water mass characteristics such as temperature (SWT), salinity (SWS), DIC and total alkalinity (TA) within isopycnal ranges of σ_θ 26.5 – 27.0 (Table 2). In a next step, we quantified the effect of biologically-mediated CO_2 uptake at the surface by assuming the complete consumption of all nutrients in the cold upwelled water. Considering a constant Redfield-C:P ratio of 117:1⁴¹, the impact of biology can be estimated by subtracting the amount of P-associated carbon ($\text{P} \times 117$) from the source water's initial DIC concentration. Additionally, we consider its associated effect on TA using a factor of ~ 0.12 ⁴². This results in $p\text{CO}_2$ decreasing below atmospheric concentrations, to levels of $\sim 335 \mu\text{atm}$ in the NBUS and $\sim 284 \mu\text{atm}$ in the SBUS. In a third step, the warming of upwelled water was considered by using SST instead of the in situ SWT of the source waters. This indicates a temperature increase of approximately 7°C and led to enhancements in $p\text{CO}_2$ by $102 \mu\text{atm}$ and $97 \mu\text{atm}$ in the NBUS and SBUS, respectively. The resulting $p\text{CO}_2$ of $\sim 437 \mu\text{atm}$ (NBUS) and $\sim 381 \mu\text{atm}$ (SBUS) resemble our estimated annual mean $p\text{CO}_2$ of the NBUS ($487.6 \pm 2.9 \mu\text{atm}$) and SBUS ($379.2 \pm 1.74 \mu\text{atm}$). This gives confidence in our outlined source water mass properties and the obtained CO_2 climatology, and suggests that deviations from the used classical Redfield-C:P ratio of 117:1, e.g., due to anoxic processes in bottom waters and surface sediments on the NBUS shelf, are of minor importance on the regional scale in the BUS⁴³⁻⁴⁵. More importantly, it shows that, in contrast to the NBUS, the biologically-mediated drawdown of CO_2 overcompensates for the warming effect in SBUS, explaining the opposing CO_2 flux directions in the two subsystems. In comparison to SCAW, higher P_{pref} concentrations of ESACW explain, in turn, the stronger biologically-mediated drawdown of CO_2 in the SBUS.

To further quantify this effect, we subdivided the biological impact by individually considering the decrease in $p\text{CO}_2$ through the use of regenerated versus preformed nutrients (Supplementary Fig. 2). Compared to a $p\text{CO}_2$ of ~ 437 (NBUS) and ~ 381 (SBUS) μatm after complete nutrient consumption, the utilization of regenerated nutrients merely results in a $p\text{CO}_2$ of ~ 650 and $\sim 662 \mu\text{atm}$ in the NBUS and SBUS, respectively. This implies that the consumption of preformed nutrients reduces $p\text{CO}_2$ by about

32.8% ($650 - 437 = 213 \mu\text{atm}$) in the NBUS and 42.5% ($662 - 381 = 281 \mu\text{atm}$) in the SBUS, which resembles the mean share of P_{pref} to the total P of 29% in the NBUS and 47% in SBUS (Table 2).

In addition, we assessed the contribution of P_{pref} to the potential amount of CO_2 transfer into the ocean interior on the basis of new production rates from the BUS that were calculated e.g., by multiplying the volume of upwelling waters^{1,46} with corresponding nutrient inventories of the source water masses³¹. The resulting estimates for the NBUS ranged between 180 and $650 \text{ g C m}^{-2} \text{ year}^{-1}$ ³¹ which, given our defined upwelling area of 377.400 km^2 , amounts to 68 - 245 Tg C year^{-1} . With a mean share of P_{pref} to the total P of ~29% in the NBUS (Table 2), the utilization of P_{pref} results in a new production of around 19.7 - 71.1 Tg C year^{-1} . Considering the volume of upwelled waters in the SBUS³⁶ and corresponding nutrient inventories of ESACW (Table 2) which results in a new production of 0.279 Tg C year^{-1} , the contribution of P_{pref} to new production from the SBUS is low. However, in comparison to the CO_2 release during the formation of P_{pref} in the Atlantic sector of the Southern Ocean ($110 \text{ Tg C year}^{-1}$ ¹⁵⁻¹⁷), a new production driven by the utilization of P_{pref} of 19.7 - 71.1 Tg C year^{-1} implies that the biological carbon pump in the BUS counterbalances over 20% of the CO_2 loss in the subpolar South Atlantic region (Supplementary Fig. 3). This emphasizes the role of the BUS as a significant hub for restoring the CO_2 uptake efficiency of the biological carbon pump. Yet, its strength could be weakened with the predicted decrease in upwelling intensity in the NBUS^{47,48}. Whether the predicted enhancement in upwelling of the southern system^{36,49} could however compensate for the lowered biological carbon pump efficiency of the NBUS, remains open for debate. Thus, in addition to the Southern Ocean, also a better understanding of the BUSs response to climate change is essential to predict the CO_2 uptake efficiency of the biological carbon pump and the future evolution of CO_2 in the atmosphere.

Methods

Study region

The Benguela Upwelling System (BUS) stretches across the north of Cape Frio from ~15°S to Cape Agulhas (~35°S), while being bound by warm waters of the Angola and Agulhas current to its northern

207 and southern ends, respectively³⁸. Coastal upwelling is thereby controlled by south-easterly winds
 208 emanating from the interplay between the South Atlantic Anticyclone (SAA) and the continental low
 209 pressure trough, causing the emergence of distinct upwelling cells along the shoreline³⁷. The strongest
 210 upwelling cell is located at Lüderitz (~26°S) and separates the northern (NBUS) from the southern (SBUS)
 211 upwelling region⁵⁰. South Atlantic Central Water (SACW) is mainly formed in the Brazil-Malvinas
 212 confluence zone, where it is subducted beneath warmer subtropical water masses and transported
 213 eastward across the South Atlantic into the Cape basin⁵¹⁻⁵³. Here it converges with the Agulhas water
 214 from the Indian Ocean to form ESACW that enters the SBUS from the south⁵⁴⁻⁵⁶. The majority of the
 215 SACW circumvents the SBUS along the Benguela Current and enters the BUS from the north via the
 216 Angola-Benguela Frontal Zone (ABFZ) as a poleward undercurrent, which, unlike ESACW, is nutrient-
 217 enriched and oxygen-depleted⁵⁷. In addition, both SACW and ESACW comprise of subtropical mode
 218 waters which are characterized by preformed phosphate (P_{pref}) concentrations of ~0.6 μM , which in
 219 contrast to P_{pref} of Southern Ocean waters (~1.6 μM)⁵⁸ strongly correspond with the average P_{pref}
 220 concentrations of the NBUS (~0.54 μM) and SBUS (~0.75 μM).

221 Upwelling intensities in both subsystems exhibit a seasonal pattern which is mainly driven by temporal
 222 shifts of the SAA. As the SAA moves north-westward, the NBUS experiences maximum upwelling
 223 intensities during austral winter (June-August), while leading to a dominating westerly wind regime
 224 which weakens upwelling in the SBUS. With the south-eastward displacement of the SAA, upwelling in
 225 the SBUS is mainly confined to the summer season (September-March)^{38,59,60}. Additionally, the
 226 seasonality in upwelling is more pronounced in the SBUS as compared to the NBUS due to the perennial
 227 upwelling-favourable winds that reign in the northern region³⁷. This is also reflected in primary
 228 productivity which in the SBUS is twice as large in the summer than winter⁶¹. Furthermore, studies have
 229 shown a decrease and intensification of upwelling in the NBUS and SBUS, respectively, over the past
 230 decade in response to global warming^{36,49} and a southward shift of the SAA⁴⁷. The associated increase in
 231 sea water temperatures in the NBUS⁶² could potentially be related to the prominent decrease in

zooplankton size spectra in the BUS⁶³. Nevertheless, the productivity has remained consistently elevated, and still supports high fishery yields in both subsystems with sardine and horse mackerel as main target species in the SBUS and NBUS, respectively⁶⁴⁻⁶⁶.

Sea surface pCO₂ data collection

For the analysis and quantification of air-sea gas exchanges within the BUS, continuous underway measurements were carried out as described in ref. study³¹ during 2008 and 2019 (Supplementary Table 1). Additional quality-controlled measurements on sea surface fCO₂ from the Surface Ocean CO₂ Atlas (SOCAT) v2020²⁹ were converted into pCO₂⁶⁷ and embedded into our analysis. All pCO₂ measurements were normalized to a reference year (2020)⁶⁸ by using a mean yearly change rate in seawater pCO₂ of 1.5 µatm year⁻¹ and multiplying it with respective observations for both upwelling regions (NBUS, SBUS). All normalized measurements were distributed on a 0.1° x 0.1° grid, with each value representing the average of all observations falling into the same grid cell. Overall, the extended data set on pCO₂ records used in this study covers the shelf and coastal areas along the continental margin off Namibia and South Africa, while spanning a timeframe from 1986 to 2020 with over 250 000 data points within the area from approximately 5°E to 18.7°E, and 16°S to 34.5°S.

Carbon flux calculation

Differences in the partial pressure of carbon between the sea surface ($pCO_{2,sw}$) and atmosphere ($pCO_{2,at}$) were used to determine carbon flux rates (FCO_2) using equation (1):

$$FCO_2 = K_0 * k * (pCO_{2,sw} - pCO_{2,at}) \quad (1)$$

where K_0 is the solubility coefficient of CO₂⁶⁹ and k represents gas transfer velocity of CO₂⁷⁰. The gas transfer velocity k was calculated following equation (2):

$$k = 0.251 * u^2 * \left(\frac{Sc}{660}\right)^{-0.5} \quad (2)$$

Sc is the Schmidt number of CO₂ in seawater, 660 represents Sc at 20°C water temperature and u refers to the wind speed (m/s) at 10 m above the sea surface. Additional data on sea surface temperature (SST)

(°C) and salinity (PSU) were thereby required for the determination of Sc using the parameterization after Wanninkhof et al.⁷⁰. Data on wind speed, SST and salinity were based on shipboard measurements (Supplementary Table 1), which were distributed on a $0.1^\circ \times 0.1^\circ$ grid and averaged over each grid cell. The flux calculation was performed using the average sea surface pCO_2 , wind speed, SST and salinity of all grid cells within the defined NBUS and SBUS region. Seasonal flux estimates were calculated using gridded and averaged pCO_2 , wind speed, SST and salinity data collected during spring and summer (September till March), and austral autumn and winter (April till August), respectively.

Water column sampling and analysis

The analysis of water mass characteristics and biogeochemical settings in the BUS was based on data gathered during *RV Meteor* cruise M153, which covered on- and offshore areas of both subsystems and hence allowed a direct comparison of both upwelling zones for one given timeframe. We collected CTD profiles of temperature, salinity and oxygen, and defined the upwelling SACW and ESACW source waters by using the definition provided by ref.⁷¹. The analysis of dissolved inorganic nutrients (phosphate P, nitrate N) was carried out as outlined in ref.^{44,72}. Furthermore, the calculation of preformed and regenerated nutrients (P, N_{pref} and P, N_{reg} respectively) was performed following ref.⁵⁸ by including the apparent oxygen utilization (AOU) as well as the oxidation ratios $R_{P:O_2} = 1/138$ (phosphate) and $R_{N:O_2} = 14/138$ (nitrate), using equation (3) and (4):

$$P, N_{reg} = R_{P,N:O_2} AOU \quad (3)$$

$$P, N_{pref} = P, N - P, N_{reg} \quad (4)$$

These oxidation ratios are based on our nutrient analysis, with a N:P-ratio of ~14:1, in line with ref. study³¹. The analysis of dissolved inorganic carbon (DIC) was performed with a cavity ringdown spectrometer (Picarro G2201-I, 1510CFIDS2047_v1.0) attached to a Liaison A0301 and an AutoMate Prep device. For the analysis of total alkalinity (TA), samples were collected in 250 ml borosilicate bottles using silicone tubes (Tygon). The bottles were rinsed twice, filled from the bottom to avoid

bubbles and analysed on board using the VINDTA 2C system (Marianda, Kiel, Germany). The samples were titrated with a fixed volume of hydrochloric acid (HCl) by equal increments of HCl (0.1 N HCl). Both the Picarro and VINDTA 2C systems were calibrated using Certified Reference Material (CRM, batch #177) provided by A. Dickson (Scripps Institution of Oceanography, La Jolla, CA, USA), while affirming an accuracy of $\pm 12 \mu\text{mol kg}^{-1}$ and $\pm 4.3 \mu\text{mol kg}^{-1}$ for DIC and TA, respectively.

Statistical information

The uncertainty in average parameter calculations of pCO_2 , CO_2 exchange coefficients and fluxes (Table 1) as well as biogeochemical characteristics of source waters (Table 2) is presented as the standard error (s.e.), together with the number of values (n) used for the average calculation.

Data Availability

The underlying data of this study on pCO_2 and CO_2 flux coefficients were obtained from the Surface Ocean CO_2 Atlas (SOCAT) v2020 (<https://www.socat.info/index.php/data-access/>) and from the Pangaea data pool (access link to individual cruises provided in Emeis *et al.*³¹). Shipboard underway data of German research vessels are further accessible via the German Oceanographic Data Centre (DOD) under the Federal Maritime and Hydrographic Agency (https://www.bsh.de/EN/Home/home_node.html). Average atmospheric concentrations of CO_2 were obtained from the Global Monitoring Laboratory (GML) of the U.S. National Oceanographic and Atmospheric Administration (NOAA) Research (<https://www.esrl.noaa.gov/gmd/ccgg/trends/data.html>).

Code Availability

The CO2SYS program for Python as applied in this study is freely available and documented under <https://pypi.org/project/PyCO2SYS/>.

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Author contributions

TR and NL designed the study. The shipboard data collection and analysis in 2008 to 2014 was performed by TR, and in 2019 by TR and CS. CS led the writing of the manuscript. All authors discussed the results and contributed to the writing of the manuscript.

Ethics declarations

Competing interests

The authors declare no competing interests.

Supplementary information

Supplementary Figure 1-3

Supplementary Table 1

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Figures

Figure 1: Maps of pCO₂ measurements of research vessel cruises within the Benguela Upwelling System used in this study.

a Underlying cruise tracks of the 14 cruises (red) and those embedded in SOCAT v2020²⁹ (blue). **b** Recorded pCO₂ measurements (in μatm), normalized to the reference year 2020. **c** Normalized pCO₂ measurements (in μatm) interpolated on a 0.5° grid. Contour lines represent the atmospheric pCO₂ of the reference year 2020 based on Mauna Loa records (414 μatm). The interpolation was performed after the minimum curvature interpolation method⁷⁴ implemented in the Generic Mapping Tools (GMT).

Figure 2: Spatial and temporal variability of sea surface pCO₂.

a Average cross-shelf distribution of all pCO₂ measurements (in μatm) of the northern part of the BUS (NBUS) with distance to the coast (in km). The grey line marks the offshore boundary of the upwelling zone. Values were normalized to the reference year 2020, and averaged over intervals of 10 km coastal distance. **b** As in **a**, except for the southern part (SBUS). **c** Average latitudinal pCO₂ concentrations across the NBUS and SBUS regions, for austral spring and summer (blue) and austral autumn and winter (red). The dashed grey line in each subplot represents the atmospheric pCO₂ of the reference year 2020 based on Mauna Loa records (414 μatm). The shaded areas in each subplot represent the standard deviation.

515 Tables

516 **Table 1: CO₂ exchange coefficients and flux rates for the northern and southern Benguela**

517 **Upwelling System.**

518 Annual and seasonal mean air-sea CO₂ gas exchange rates, with positive flux values indicating CO₂
 519 outgassing. Each CO₂ exchange coefficient represents the average value calculated on the basis of
 520 shipboard data gathered during the 14 cruises and those embedded into SOCAT v2020²⁹, which we
 521 distributed on a 0.1°x0.1° grid. Uncertainties are given as the standard error. The corresponding *n* number
 522 of observations equals the number of grid cell values, each representing the average of all measurements
 523 falling into the same grid cell during the respective time period (Annual, September-March, April-
 524 August).

Parameter	NBUS			SBUS		
	Annual	Sep.-Mar.	Apr.-Aug.	Annual	Sep.-Mar.	Apr.-Aug.
Sea Surface Temperature (°C)	17.66 ±0.07 <i>n</i> =1549	18.01 ±0.07 <i>n</i> =1348	15.99 ±0.07 <i>n</i> =495	17.31 ±0.06 <i>n</i> =1004	17.69 ±0.07 <i>n</i> =855	15.62 ±0.07 <i>n</i> =335
Sea Surface Salinity (PSU)	35.25 ±0.01 <i>n</i> =1549	35.24 ±0.01 <i>n</i> =1348	35.24 ±0.01 <i>n</i> =495	34.88 ±0.03 <i>n</i> =979	34.85 ±0.03 <i>n</i> =831	34.99± 0.02 <i>n</i> =332
Wind Speed (m/s)	7.96 ±0.08 <i>n</i> =1543	7.85 ±0.10 <i>n</i> =1348	7.94 ±0.11 <i>n</i> =489	7.95 ±0.12 <i>n</i> =697	8.14 ±0.13 <i>n</i> =649	5.62 ±0.32 <i>n</i> =75
pCO _{2,sw} (µatm)	487.55 ±2.9 <i>n</i> =1535	475.42 ±3.07 <i>n</i> =1326	556.03 ±5.67 <i>n</i> =492	379.15 ±1.74 <i>n</i> =1017	378.15 ±2.24 <i>n</i> =894	381.91 ±1.99 <i>n</i> =335
Solubility coefficient of CO ₂ , <i>K₀</i>	0.0345 ±0.00006	0.0342 ±0.00007	0.0363 ±0.0001	0.0350 ±0.00007	0.0346 ±0.00007	0.0367 ±0.00008
Piston velocity, <i>k</i>	14.89 ±0.33	14.61 ±0.36	14.18 ±0.41	14.72 ±0.47	15.59 ±0.53	7.04 ±0.82
Carbon Flux rate (mol C m ⁻² yr ⁻¹)	3.31 ±0.19	2.68 ±0.19	6.39 ±0.41	-1.58 ±0.03	-1.77 ±0.05	-0.73 ±0.04

Table 2: Shipboard data on source water mass characteristics and sea surface conditions used for sea surface pCO₂ simulations with CO2SYS.

Properties of source water masses presented as average parameters within isopycnal ranges of σ_θ : 26.5 – 27.0 within the NBUS and SBUS, based on CTD casts and nutrient analysis during *RV Meteor* cruise M153. Sea surface temperature and salinity (<5 m water depth) are based on annual averages as estimated from shipboard data gathered during the 14 cruises, and those provided by SOCAT v2020²⁹. The effect on sea surface pCO₂ is estimated using CO2SYS^{39,40} with input parameters on total alkalinity (TA), dissolved inorganic carbon (DIC), source water and sea surface temperature (SWT, SST), salinity (SWS, SSS), total and preformed phosphate (P_{total}, P_{pref}). Uncertainties are given as the standard error, with *n* number of observations used for average calculations.

	Source water mass (σ_θ : 26.5 – 27.0)							Sea surface (< 5 m depth)	
	DIC ^{*1}	TA ^{*1}	P _{total} ^{*1}	P _{pref} ^{*1}	P _{pref} -%	SWT ^{*2}	SWS ^{*3}	SST ^{*2}	SSS ^{*3}
NBUS	2314.23	2298.73	1.87	0.54	28.88	11.5	35.04	17.66	35.25
	±6.9	±1.7	±0.03	±0.01	±0.1	±0.09	±0.01	±0.07	±0.01
	<i>n</i> =95	<i>n</i> =97	<i>n</i> =96	<i>n</i> =96	<i>n</i> =96	<i>n</i> =267	<i>n</i> =267		
SBUS	2246.17	2283.94	1.60	0.75	46.88	10.5	34.83	17.31	34.88
	± 6.2	± 2.6	±0.05	±0.02	±0.2	±0.06	±0.01	±0.06	±0.03
	<i>n</i> =102	<i>n</i> =101	<i>n</i> =79	<i>n</i> =79	<i>n</i> =79	<i>n</i> =265	<i>n</i> =265		

Units in ^{*1} μmol kg⁻¹, ^{*2} °C, ^{*3} PSU.

Figures

Figure 1

Maps of pCO₂ measurements of research vessel cruises within the Benguela Upwelling System used in this study. a Underlying cruise tracks of the 14 cruises (red) and those embedded in SOCAT v2020 29 (blue). b Recorded pCO₂ measurements (in μatm), normalized to the reference year 2020. c Normalized pCO₂ measurements (in μatm) interpolated on a 0.5° grid. Contour lines represent the atmospheric pCO₂ of the reference year 2020 based on Mauna Loa records (414 μatm). The interpolation was performed after the minimum curvature interpolation method⁷⁴ implemented in the Generic Mapping Tools (GMT).

Figure 2

Spatial and temporal variability of sea surface pCO₂. a Average cross-shelf distribution of all pCO₂ measurements (in μatm) of the northern part of the BUS (NBUS) with distance to the coast (in km). The grey line marks the offshore boundary of the upwelling zone. Values were normalized to the reference year 2020, and averaged over intervals of 10 km coastal distance. b As in a, except for the southern part (SBUS). c Average latitudinal pCO₂ concentrations across the NBUS and SBUS regions, for austral spring and summer (blue) and austral autumn and winter (red). The dashed grey line in each subplot represents the atmospheric pCO₂ of the reference year 2020 based on Mauna Loa records (414 μatm). The shaded areas in each subplot represent the standard deviation.

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