Contribution of hurricanes to the air-sea CO$_2$ flux: a global view

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Abstract. Previous case studies have illustrated the strong local influence of tropical cyclones (TCs) on CO$_2$ air-sea flux ($F_{CO_2}$), suggesting that they can significantly contribute to the global $F_{CO_2}$. In this study, we use a state-of-the-art global ocean biochemical model driven by TCs wind forcing derived from a historical TCs database, allowing to sample the $F_{CO_2}$ response under 1663 TCs. Our results evidence a very weak contribution of TCs to global $F_{CO_2}$, one or two order of magnitude smaller than previous estimates extrapolated from case studies. This result arises from several competing effects involved in the $F_{CO_2}$ response to TCs, not accounted for in previous studies. While previous estimates have hypothesized the ocean to be systematically oversaturated in CO$_2$ under TCs, our results reveal that a similar proportion of TCs occur over oversaturated regions (i.e. the North Atlantic, Northeast Pacific and the Arabian Sea) and undersaturated regions (i.e. Westernmost North Pacific, South Indian and Pacific Ocean). Consequently, by increasing the gas exchange coefficient, TCs can generate either instantaneous CO$_2$ flux directed from the ocean to the atmosphere (efflux) or the opposite (influx), depending on the CO$_2$ conditions at the time of the TC passage. A large portion of TCs also occurs over regions where the ocean and the atmosphere are in near equilibrium, resulting in very weak instantaneous fluxes. Previous estimates did also not account for any asynchronous effect of TCs on $F_{CO_2}$: during several weeks after the storm, oceanic pCO$_2$ is however reduced in response to vertical mixing, which systematically causes an influx anomaly. This implies that, contrary to previous estimates, TCs
weakly affect the CO$_2$ efflux when they blow over supersaturated areas because the instantaneous storm wind effect and post-storm mixing effect oppose with each other. In contrast, TCs increase the CO$_2$ influx in undersaturated conditions because the two effects add up. These compensating effects result in a very weak contribution to global $F_{CO_2}$ and a very modest contribution to regional interannual variations (up to 10%).
1. Introduction

Tropical cyclones (TCs), also often called typhoons in the Northwest Pacific or hurricanes on the eastern side of the dateline, are recurring extreme summertime weather events which strongly affect the thermal and physical structure of the upper ocean along their wake [e.g. D’Asaro, 2003; Cione and Uhlhorn, 2003]. One of the strong oceanic response commonly observed in the wake of TCs is the intense cooling of the surface layer that can reach up to 10°C [Chiang et al., 2011]. This cooling mainly ensues from the wind-driven vertical entrainment of cold subsurface waters [Price, 1981; Vincent et al., 2011a].

A few previous case studies have illustrated that TCs can cause enormous effluxes of CO\textsubscript{2} between the ocean and atmosphere (hereafter $F_{CO_2}$), large enough to influence the annual local efflux [Bates et al., 1998; Perrie, 2004; Nemoto et al., 2009]. Although TCs are short-lived and travel at considerable speed (typically 4-6 m/s), they affect a significant area of the tropical ocean each year, because of their large spatial extent [several ∼100 km, Willoughby et al., 2006] and frequent occurrence (∼ 100 TCs each year). It is thus likely that they exert a significant influence on the global air-sea flux of CO\textsubscript{2}. However, the limited availability of pCO\textsubscript{2} data under TCs (due to the severe weather conditions) has so far restricted the global quantification of the $F_{CO_2}$ response to TCs to qualitative extrapolations [Bates et al., 1998; Bates, 2007; Huang and Imberger, 2010].

The $F_{CO_2}$ response to a hurricane passage depends on various factors. $F_{CO_2}$ is a function of piston velocity and of the difference in the partial pressure of CO\textsubscript{2} (pCO\textsubscript{2}) between the atmosphere and sea surface. Oceanic pCO\textsubscript{2} depends on total dissolved inorganic carbon
DIC, total alkalinity (TA), temperature (T) and salinity (S) [Takahashi et al., 1993]. TCs increase the magnitude of the $F_{CO_2}$ because the piston velocity strongly increases at high wind speed [Liss and Merlivat, 1986; Wanninkhof, 1992; D’Asaro and McNeil, 2007; McNeil and D’Asaro, 2007]. TCs also impact $F_{CO_2}$ because their intense surface winds increase vertical entrainment of subsurface waters (hereafter referred to as mixing) and thereby change the surface properties of T, S, DIC and TA and thus surface oceanic pCO$_2$. In addition, the net $F_{CO_2}$ response to TCs is complicated by the fact that the TC-induced mixing (affecting the pCO$_2$) and TC-winds (affecting the piston velocity) are not synchronous [Nemoto et al., 2009]: cooling, the manifestation of vertical mixing, is observed to be maximum between one and three days after the TC-passage [Vincent et al., 2011b] and its intensity depends on the local vertical structure of the upper ocean [Vincent et al., 2011a].

These considerations highlight the difficulty in generalizing the $F_{CO_2}$ response to TCs from individual cases and illustrate why the few global TC-induced $F_{CO_2}$ estimates [Bates et al., 1998; Bates, 2007; Huang and Imberger, 2010] are so disparate: from $\sim$0.5 PgC to $\sim$0.05 PgC per year. This paper proposes an original methodology and a model framework for accurately quantifying the TC-induced $F_{CO_2}$ and the various processes into play. We use a global Ocean General Circulation Model [NEMO, Madec, 2008] coupled with a biogeochemical model [PISCES, Aumont and Bopp, 2006] with a modified version of CORE forcing [Large and Yeager, 2009] that includes an analytic formulation of 2-dimensional TC winds along observed TC tracks [Vincent et al., 2011a]. We compare two model simulations, with and without TCs, and provide quantitative diagnostics averaged along the
tracks of TCs. This enables us to provide a global picture of the effects of hurricanes on air-sea CO$_2$ transfers.

2. Model and methods

2.1. Coupled hydrodynamic and carbon model

The hydrodynamical model used here is the global configuration (ORCA2) of Nucleus for European Modelling of the Ocean [NEMO, v3.2, Madec, 2008]. The quasi-isotropic grid has a nominal resolution of 2° with increased 0.5° latitudinal resolution at the equator. In the vertical, 31 levels are used, with 10 levels in the upper 100 m. The mixed layer dynamics is parameterized using an improved Turbulent Kinetic Energy (TKE) closure scheme with a Langmuir cell, a surface wave breaking parameterization and an energetically consistent time and space discretization [Madec, 2008]. Additional subgrid-scale mixing parameterizations include a Laplacian viscosity, an iso-neutral Laplacian diffusivity and the use of a GM-scheme to mimic the effect of subgrid-scale eddy processes [Gent and McWilliams, 1990]. Complete description of ORCA2 can be found in Cravatte et al. [2007].

The Pelagic Interaction Scheme for Carbon and Ecosystem Studies (PISCES, Aumont and Bopp, 2006) is coupled to ORCA2. PISCES includes a simple representation of the marine ecosystem and describes the cycles of carbon and of the main marine nutrients (N, P, Fe and Si). The model has 24 compartments. Four living pools are represented: two phytoplankton size classes/groups (nanophytoplankton and diatoms) and two zooplankton size classes (microzooplankton and mesozooplankton). Fixed Redfield ratios are employed for N and P, while the ratios of both Si, and Fe, to C vary dynamically as a function of the phytoplankton functional group and environmental variables. The carbon-
ate chemistry follows the Ocean Carbon-Cycle Model Intercomparison Project (OCEMIP) protocols (www.ipsl.jussieu.fr/OCMIP). The air-sea CO$_2$ exchange is calculated with the bulk exchange formula:

$$F_{CO_2} = k_s DpCO_2$$  \hspace{1cm} (1)

where $F_{CO_2}$ is the air-sea CO$_2$ flux with positive values indicating CO$_2$ goes within the ocean, $k$ is the gas transfer velocity, $s$ is the CO$_2$ solubility and $DpCO_2 = pCO_2^{atm} - pCO_2^{sea}$ is the difference of pCO$_2$ between the ocean and atmosphere. In the following, we will use the notation pCO$_2$ for $pCO_2^{sea}$. There is still large uncertainty in $k$, particularly under hurricane winds [McNeil and D’Asaro, 2007; D’Asaro and McNeil, 2007]. Following the discussion by McNeil and D’Asaro [2007], we used the relationship of Wanninkhof [1992] to calculate $k$, which provides a lower limit for $k$ under extreme winds compared to other recent formulations [Perrie, 2004].

2.2. Model set up and experimental design

The surface boundary conditions used for the present ORCA2-PISCES simulations are based on the version 2 of the atmospheric data sets and formulations developed by Large and Yeager (2009) for global ocean-ice models and are referred to as Coordinated Ocean-ice Reference Experiments (CORE) forcing [Griffies et al., 2009]. The forcing data sets are based on a combination of NCEP/NCAR reanalysis products for the years 1958-2007 with various satellite data sets, and involve adjustments that correct global imbalances (e.g., produce near zero global mean heat and freshwater fluxes when used in combination with observed SSTs). Turbulent fluxes are computed from the CORE bulk formulae.
as a function of the prescribed atmospheric state and the simulated ocean surface state (SST and surface currents). A bound to 33 \( m s^{-1} \) has been introduced to the CORE dimensionless surface drag coefficient (\( C_D \)) to account for its observed saturation at strong winds following Donelan [2004]. Data are prescribed at six-hourly (wind speed, humidity and atmospheric temperature), daily (short- and long-wave radiation) and monthly (rain and snow) resolution, with inter-annual variability over the time range 1958 to 2007 except for runoff which are kept climatological. To avoid an artificial model drift due to a freshwater imbalance, the sea surface salinity is damped towards monthly-mean climatological values with a piston velocity of 50 m per 300 days [Griffies et al., 2009].

Boundary conditions for the biogeochemical model include atmospheric dust (Fe) deposition, rivers (Fe, N, P, Si and C) and sediment (Fe) mobilization. These sources are described in Aumont and Bopp [2006] and kept constant for the duration of the simulation.

Our simulations start in 1978. For the initial state, we used physical and biogeochemical tracers initialized from previous ORCA2-PISCES experiments: the spun-up physical state was obtained by running a 120-year simulation, starting from Levitus and forced by repeating the 50-year CORE forcing; the biogeochemical state was obtained by running a transient 1870-1977 simulation with increasing \( pCO_2^{atm} \) levels following historical records.

From 1978 onwards, we performed two simulations: one with the full strength of TCs (Cyclone or C run) and one without TCs (No cyclone or N run). For these simulations, the 10-m wind forcing from CORE has been modified following Vincent et al. [2011b]: In the N run, the effect of the weaker than observed TC-like vortices seen in the original CORE forcing has been filtered out by applying a 11-day running mean to the wind components within 600km of each cyclone (Fig. 1). In the C run, analytical TCs have
been superimposed to the filtered CORE winds. To do so, TC winds are parameterized using the idealized Willoughby and Rahn [2004] TC wind spatial pattern, which is based on a statistical fit to the observed TC winds [Willoughby et al., 2006]. This idealized wind pattern is interpolated in time at each model time step using the position and maximum wind speed of each cyclone in the IBTrACs database (www.ncdc.noaa.gov/oa/ibtracks). This analytical formulation enables to correct for the underestimation of the strength of TC-winds in CORE. Fig. 1 provides an illustration of the different wind forcings (CORE, N and C) under a TC in the south-west Indian Ocean.

Daily model outputs were saved from 1993 onward. Our analysis are thus based on the 1993-2007 period, and sample the ocean’s response to 1663 TCs.

2.3. Diagnostics of oceanic anomalies generated by TCs along their tracks

The anomalies of a given oceanic variable \( V \) (where \( V \) is for instance the temperature or \( pCO_2 \)) ensuing from the passage of TCs, noted \( \Delta V \), are computed under each TC track. To compute \( \Delta V \), the seasonal cycle of \( V \) is first removed from \( V \). TC track locations, available at 6h intervals, are used to retrieve \( V \) at the surface of the ocean. To characterize the amplitude of the ocean response around each TC-track position, we compute \( \mathcal{V} \), the average of seasonally-detrained \( V \) over a fixed radius of 200km (about 3-4 radius of maximum wind) around the track position. We do this averaging because the impact of TCs on sea-surface properties is not restricted to a narrow path along the storm’s center but rather can occur in a swath hundreds of kilometers wide, as reported for instance by Bond et al. [2011]. The reference unperturbed pre-storm conditions (\( \mathcal{V}_{before} \)) is defined as the average of \( \mathcal{V} \) over 10 to 3 days prior to TC passage. The evolution of the ocean response anomaly at time \( t \) to a TC is then defined as \( \Delta V(t) = \mathcal{V}(t) - \mathcal{V}_{before} \). The post-
storm anomalies $\Delta V(t)$ are generally maximum within 1 to 4 days after the TC passage (section 3.1). We define $\nu_{after}$ as the mean value of $\nu$ over days 1 to 4 after TC passage, and the wake anomaly as $\bar{\Delta} V = \nu_{after} - \nu_{before}$ (Fig. 2).

This methodology is applied to the model runs and to satellite observations of SST. In the N-run, where TCs have been smoothed out, $\Delta V_N$ is close to zero but is not strictly equal to zero because of the presence of variability not due to TCs. This background variability is present in both the C and N runs. To remove it from our estimates of $\Delta V$, for model variables we define $\Delta V$ as $\Delta V = \Delta V_C - \Delta V_N$, where subscripts N and C refer to the model solutions of the N and C runs, respectively. Removing $\Delta V_N$ does not change our main results and reduces the dispersion (not shown).

### 2.4. Air-sea flux anomaly diagnostics

Anomalies in $F_{CO_2}$ in response to TCs ($\Delta F_{CO_2}$) can result either from changes in the wind (and thus in gas transfer velocity $\Delta k$) or from changes in mixing (and thus in $\Delta (s DpCO_2)$):

$$
\Delta F_{CO_2} = \Delta k \times s DpCO_2 + k \times \Delta (s DpCO_2)
$$

Using the results from the C and N runs, the wind and mixing effects are approximated as:

$$
\Delta k \times s DpCO_2 \simeq (k_C - k_N) \times s_N (DpCO_2)_N
$$

and

$$
k \times \Delta (s DpCO_2) \simeq k_N \times (s_C (DpCO_2)_C - s_N (DpCO_2)_N)
$$
where subscripts N and C refer to the model solutions of the N and C runs, respectively. Note that the second order terms have been omitted in this approximation; a posteriori, our results show that they are smaller than the first order terms.

3. Model evaluation

3.1. Evaluation of TC-induced cooling

The C run captures very accurately the timing of the average observed cooling (Fig. 2a). In both model and observation, the temperature decrease starts a few days before the wind reaches its peak intensity at a given location (Day 0) and reaches its maximum amplitude 2-3 days after the TC passage. The cold anomaly persists for several weeks after the TC passage and is largely restored after 40 days. We note however that SST remains on average 0.2° colder than pre-storm conditions, in agreement with other observational analysis [Price et al., 2008; Lloyd and Vecchi, 2011; Vincent et al., 2011a]. Based on this composite picture, the average value of the e-folding time (time for the cold anomaly to be reduced by a factor e) is 15 days. Moreover, there is a 0.68 correlation between modeled and observed TC-induced cooling magnitudes (ΔT) in the wake of TCs at individual locations (Fig 2b), indicating that our simulation realistically samples the ocean response to the wide spectrum of TC characteristics. Because the analytic cyclone wind field formulation is fitted to an average of observed cyclone wind radial profiles [Willoughby and Rahn, 2004], we however do not expect each simulated cold wake to match the corresponding observed one, as illustrated by the spread on Fig. 2b. The model also successfully reproduces the observed spatial distribution of the TC-induced cooling (Fig. 3): the average cooling within TC-active regions is about 1°C in all basins, with maximum amplitude of about 2°C in the north-west Pacific region where amongst the most intense TCs occur.
The main model deficiency lies in the North-East Pacific basin where modeled coolings are overestimated by almost 1°C. This bias can be tracked back to a shallower than observed thermocline in this region. The use of a coarse horizontal resolution (2°) in our simulation do not hinder the realism of the ocean response to TCs: similar comparisons to observations have indeed been obtained with a simulation of higher (0.5°) resolution using similar boundary conditions and forcing strategy [Fig. 2 in Vincent et al., 2011a].

3.2. Evaluation of the DpCO$_2$

Simulated DpCO$_2$ has been evaluated against observations-based climatologies in Aubmont and Bopp [2006] both in terms of annual-mean, seasonal amplitude and phasing. Here, we focus on the mean DpCO$_2$ during the TC-season, i.e. from November to April in the southern hemisphere and from May to October in the northern hemisphere (Fig. 4). The TC-season DpCO$_2$ displays regions both with positive and negative values qualitatively well reproduced by the model. Regions with negative DpCO$_2$ (oceanic sources of CO$_2$ to the atmosphere) are in the tropical eastern Pacific, the tropical Atlantic, the Arabian sea. Regions with positive DpCO$_2$ (oceanic sinks for atmospheric CO$_2$) are found off the west coast of Australia, in the Southwest Pacific and in the mid-latitudes of all oceanic basins.

4. Results

4.1. Case of TC Felix at station BATS

The observations of Bates et al. [1998] at station BATS (31°10’N, 64°10’W) in the Western North Atlantic before and after the passage of TC Felix provide a unique opportunity to evaluate the response in pCO$_2$ predicted by our model. Around BATS, typical
summertime conditions are characterized by weak winds and seawater pCO$_2$ larger than atmospheric pCO$_2^{atm}$ levels, causing a small flux from ocean to atmosphere. Felix passed over the station on 14-15 August 1995. It was quite large (350 km in diameter) and had sustained winds of 40-45 m s$^{-1}$. For several weeks before the storm, the SST was close to 28°C and pCO$_2$ ranged from 400-420 μatm. A sharp decrease of SST and pCO$_2$ occurred following the TC passage, with similar amplitudes in both model and observations (resp. -3°C and -45 μatm). The SST and pCO$_2$ rebounded after passage of Felix, but did not return to pre-storm values (data from Bates et al., 1998, reproduced in Fig. 5a,b). The oceanic response to Felix is hence successfully captured by our simulation although the absolute pCO$_2$ values display a $\sim$ 10μatm systematic shift in our experiment (Fig. 5a,b), which is rather understandable given the coarse model resolution used here.

The TC passage results in significant changes of the modeled $F_{CO_2}$ (Fig. 5c). In the C-run, the $F_{CO_2}$ peaks very abruptly during the 2-days corresponding to the TC passage (August 14-15) reaching 58 mmole/m$^2$/day, then nearly vanishes (from August 16-29) until three other storms hit the area (in September). In the N-run, the $F_{CO_2}$ do not experience any large variations during the passage of the main TC and of the subsequent storms compared to the C-run, but is larger during the more quiet periods (Fig 5c,d).

To interpret the differences of $F_{CO_2}$ between the 2 runs, the wind and mixing effects have been separated according to equations 2, 3 and 4 applied at the fixed location of BATS (Fig. 5d). This diagnostics clearly reveals that the wind effect is the driver of the strong increase of $F_{CO_2}$ during the 2 days of the TC-passage and of the three subsequent storms (green curve in Fig. 5d). On the other hand, the mixing effect (blue curve in Fig. 5c) acts to increase the pCO$_2$ and thus to reduce the $\Delta pCO_2$, which reduces the
Moreover, the mixing effect is lagged and last longer compared to the wind effect, reducing $F_{CO_2}$ for more than 2 weeks reaching a maximum 2-3 days after the passage of Felix.

The $F_{CO_2}$ caused by Felix has been previously estimated by Bates et al. [1998], by assuming that the oceanic pCO$_2$ was constant in the lead up of the storm, then decreased linearly to the end of the hurricane. This led to a total flux of 40 mmoleC m$^{-2}$ during the two-days passage of the storm; this estimate did not account for the flux reduction during post-hurricane conditions. With our model, we predict a $\tilde{\Delta}F_{CO_2}$ of 62 mmoleC m$^{-2}$ during the storm. This larger number compared with to Bates et al. (1998) is due to the slower decrease of the oceanic pCO$_2$ in the model compared to the linear decrease hypothetized by Bates et al. (1998). Koch et al. (2009) also estimated the $\tilde{\Delta}F_{CO_2}$ caused by Felix during the two-days of the storm with a regional model, forced with daily NCEP winds. They found a lower value than Bates et al. (1998) of 32 mmoleC m$^{-2}$, which they attribute to the underestimation of their model wind forcing. In a second step, to account for the post-storm effect, we integrate the impact of TC Felix over two weeks. In this case, the increase of the $\tilde{\Delta}F_{CO_2}$ during the TC passage is counter-balanced by an increased during post-TC conditions by approximately half, which leads to the final estimate of 30 mmoleC m$^{-2}$. This post-storm effect was not accounted for in the estimates of Bates et al. [1998] and Koch et al. [2009].

A situation similar to that observed at BATS by Bates et al. [1998] was reported by Nemoto et al. [2009] in the coastal East China Sea, from continuous temperature and pCO$_2$ measurements at a moored buoy. Because their buoy was located in a coastal area, we cannot directly compare their observations with our 2º model results. However, Nemoto
et al. [2009] examined how the variations in wind and in pCO$_2$ affected the F$_{CO_2}$. They found, in agreement with our analysis at BATS, that the F$_{CO_2}$ efflux is increased during the passage of the storm, because of the increase in wind speed, and is decreased after the storm has passed, because of the decrease of pCO$_2$. Nevertheless, their estimate of the impact of TCs on the F$_{CO_2}$, approximately 30 mmoleC m$^{-2}$ per typhoon, only accounts for the wind effect.

4.2. Sign of DpCO$_2$ under TCs

The efflux observed at BATS during TC Felix results from the negative DpCO$_2$ at the time $t_0$ of the TC passage. Previous studies that attempted to estimate the global impact of TC on F$_{CO_2}$ have generalized this observation, hypothesizing that DpCO$_2$ at $t_0$ was negative for all TCs. Our numerical experiment allows to check this hypothesis (Fig. 6): we found almost as many occurrences of negative and positive DpCO$_2$ at $t_0$ (25773 versus 21178 cases). The immediate consequence is that TCs do not systematically cause CO$_2$ effluxes at $t_0$ (as in BATS) but also be responsible for influxes, in similar proportion (Fig. 6b). Moreover, $|DpCO_2|$ at $t_0$ is larger than 10 $\mu$atm in $\sim$ 50% of the cases, and cause the largest F$_{CO_2}$ anomalies (tails of the distribution in Fig. 6b). In the other $\sim$ 50% cases, the impact of TC on the F$_{CO_2}$ is small because DpCO$_2$ is close to zero.

The spatial distribution of the positive and negative DpCO$_2$ cases under TCs in the C-run is displayed in Fig. 7a,b. Areas with negative DpCO$_2$ during the cyclonic seasons are found mostly in the Western North Atlantic, Western and Eastern North Pacific, Arabian Sea and South West Indian Ocean, while areas of positive DpCO$_2$ are mostly in the Westernmost North Pacific, central south Indian Ocean, Bay of Bengal and South western Pacific, in broad agreement with the pCO$_2$ distribution during the cyclonic season (Fig.
The same diagnostic applied to the climatology of Takahashi et al. [2009] interpolated in time reveals similar regional patterns (Fig. 7c,d). The differences are due both to model imperfections and to the absence of interannual variations in the climatology.

### 4.3. TCs induced DpCO$_2$ and $F_{CO_2}$ anomalies

Since the $F_{CO_2}$ response differ depending on the sign of the DpCO$_2$ background conditions at the time of TCs passage, two different cases are considered in the following: the case where DpCO$_2$ is negative at $t_0$, causing an eflux to the atmosphere (case of BATS), and the opposite case, where DpCO$_2$ is positive at $t_0$, causing an influx to the ocean.

Figure 8 shows the composite evolution of wind speed, SST, DpCO$_2$ and $F_{CO_2}$ anomalies associated with the passage of all TCs falling in each category, and computed following section 2.3.

In the two cases, the composite time-evolution of the wind, SST and DpCO$_2$ TC-induced anomalies are consistent with the results obtained in the case of TC Felix at BATS: TCs induce a cooling of the SST (Fig. 8b), which reaches its maximum amplitude two to three days after the maximum wind speed intensity and slowly returns to its pre-storm value during the month that follows the passage of the TC. In our simulation, the maximum coolings reach 4°C, and the mean cooling when averaged over all TCs is less than 1°C. In agreement with Vincent et al. [2011b], this cooling mainly results from the entrainment of cold sub-surface waters driven by the storm, especially for the strongest wind forcing. The mean cooling is larger in the negative DpCO$_2$ case, because different oceanic regions are concerned in the two cases. Vertical mixing not only decreases SST but also, in agreement with the theoretical pCO$_2$ change with temperature [Sarmiento and Gruber, 2006], decreases pCO$_2$ and thus increases DpCO$_2$ (Fig. 8c). Moreover, as for SST, the
maximum magnitude of the pCO$_2$ drawdown is reached 2 to 3 days after the passage of the storm (Fig. 8c), and the change is larger in the negative DpCO$_2$ case.

Because the pre-storm air-sea fluxes have different signs in the two cases, the direct impact of TC winds on the air-sea CO$_2$ fluxes are opposite. At $t_0$, TC winds acts to increase the amplitude of the flux in either direction, thus enhancing effluxes in the case of negative DpCO$_2$, and enhancing influxes in the case of positive DpCO$_2$ (Fig. 8). After the storm, the decrease of pCO$_2$ causes $|DpCO_2|$ to decrease in the negative DpCO$_2$ (undersaturated) case (and eventually change sign), and to increase in the positive DpCO$_2$ (oversaturated) case (Fig. 9); consequently, the amplitude of the flux is reduced in the oversaturated case (and eventually reverses), and is increased in the undersaturated case, in comparison with pre-storm conditions (Fig. 8 and 9).

4.4. Respective influence of TCs wind and induced mixing on $F_{CO_2}$

To assess and quantify the impact of TCs on $F_{CO_2}$, the wind and mixing effects have been separated according to equations 2, 3 and 4 for each TC, and averaged out for all oversaturated (resp. undersaturated) cases (Fig. 10). In both cases, the wind effect explains most of the increase of the flux in either direction during the TC-passage, while the mixing effect explains the modifications of the flux after the storm, in opposite sign in the case of oversaturation, and in same sign in the case of undersaturation.

In the oversaturated case, when integrated over 30 days, the time integral of the TC-induced flux anomaly is 1.1 mmole/m$^2$ and is positive. This means that, on average over all TCs, the negative wind-driven flux anomaly during the storm is more than offset by the post-storm, positive, mixing-driven flux anomaly. In the undersaturated case, the time
integral of the TC-induced flux anomaly is 10.1 mmole/m$^2$; in this case, the wind-driven
and mixing-driven anomalies are both positive and add up.

4.5. Regional and global impact of TC on $F_{CO_2}$

In Figure 11, the storm and post-storm impacts of TCs on $F_{CO_2}$ are evaluated regionally.
This is done by integrating $\Delta F_{CO_2}$ over two time periods.

During the storm peak intensity ($t_0-1.5d$ to $t_0+1.5d$, Fig. 11b), the averaged TC-driven
flux anomaly is negative in the North West Atlantic, North West Pacific, North East
Pacific, Arabian Sea and South West Indian, and is positive in the Westernmost North
Pacific, central south Indian Ocean, Bay of Bengal and South western Pacific. These
patterns are due to the first (resp. second) regions being predominantly supersaturated
(resp. undersaturated) during the cyclonic season (Fig. 4). After the storm ($t_0+1.5d$ to
$t_0+30d$, Fig. 11c), the flux anomaly is systematically positive (except for a small region
in the north eastern Pacific, where $pCO_2$ is enhanced by vertical mixing) and its pattern
is that of the TC density. When considering the whole period ($t_0-1.5d$ to $t_0+30d$, Fig.
11a), the storm and post-storm effects often balance when they are of opposite signs.

In consequence, in the North West Atlantic and Arabian Sea, which are mostly supersat-
urated during the cyclonic season, the net effect of TC on $F_{CO_2}$ during the cyclonic season
is very weak because the storm and post-storm effects almost compensate. The situation
is different in the North West Pacific, where the strongest SST anomalies occur (Fig.
3); because of these strong SST anomalies, the post-storm effect prevails over the storm
effect, with the consequence of a net TC-induced CO$_2$ influx anomaly. In regions which
are predominantly undersaturated during the cyclonic season, the storm and post-storm
effects reinforce each other. Consequently, the mean effect of TC in the Westernmost
North Pacific, central south Indian Ocean, Bay of Bengal and South western Pacific is to increase the uptake of CO$_2$ by the ocean during the cyclonic season.

Tables 1 and 2 provide more quantitative numbers of the impact of TCs on $F_{CO_2}$ over the large regions delimited in Fig. 11. Except in the North West Pacific and in the Bay of Bengal, the flux due to TC is generally less than $\pm 3\%$ of the total flux. In the North West Pacific and Bay of Bengal the percentage is larger ($+33.4\%$ and $-13.6\%$, respectively), but they concern regions which weakly (15.5 and -0.6 TgC, respectively) contribute to the total $F_{CO_2}$ during the cyclonic season (-305 TgC). Note also that in most regions, the flux due to TCs is not systematically in the same direction (Table 2); for instance, the North Atlantic is submitted to an average of 80 TC-days over supersaturated conditions, contributing to a TC efflux of -1.6 TgC, but also to 16 TC-days over undersaturated regions, which cause an influx of +1 TgC. Under undersaturated conditions, the TC induced flux is always positive, while it can be either positive or negative under supersatured conditions (depending on the relative strength of the storm and post-storm effects) (Table 2). The North West Pacific and Bay of Bengal are predominantly undersaturated, which explains the larger $\%$ of the TC-flux to the total flux.

5. Discussion

5.1. Comparison with previous estimates

The global impact of TCs on air-sea CO$_2$ fluxes estimated in this study ($0.007\ Pg\ C\ y^{-1}$, i.e. 2$\%$ of the mean, subtropical flux during the cyclonic season, Table 1) is one order of magnitude less than the lowest previous estimates [Perrie, 2004; Bates, 2007; Koch et al., 2009; Huang and Imberger, 2010] and is in opposite direction (influx anomaly). This is because previous estimates did not account for two important aspects: first, in supersat-
urated regions, they did not account for the post-storm mixing-effect, which balances the
storm wind-effect; second, they did not consider the impact of TCs over undersaturated
regions.

Previous estimates of air-sea $CO_2$ fluxes under TCs mainly focused on hurricanes in
the North West Atlantic \cite{Bates et al., 1998; Perrie, 2004; Bates, 2007; Koch et al., 2009;
Huang and Imberger, 2010}, with the exception of one study in the East China Sea \cite{Nemoto
et al., 2009}. These studies took place in undersaturated regions. In most of them, it was
attempted to extrapolate the results from a few TCs (one to three TCs, typically) to the
global ocean, by using TCs database and making assumptions about storm area, wind
speed and oceanic pCO$_2$ values. Moreover, it was generally assumed that the oceanic
conditions at the time and location of TC passage did not significantly deviate from the
case study that was examined. The different assumptions and methods used led to a wide
range of estimates. The first global estimate by Bates et al. \cite{1998} was a TC-induced
efflux of up to 0.51 Pg C. Later, Bates \cite{2007} estimated a smaller efflux of 0.04-0.08 Pg C,
but no details on the method were provided. The method of Huang and Imberger \cite{2010},
with different extrapolation assumptions, lead to an efflux of 0.047-0.141 Pg C. Nemoto
et al. \cite{2009} restricted their extrapolation to the western subtropical North Pacific, and
found a contribution of TCs equal to 76\% of the summer efflux, a value even larger than
the 20-54\% estimate of Bates et al. \cite{1998}. Our estimate in for North West Atlantic is
close to zero (1\%, Table 1), because the wind and mixing effect are in balance. In the
North West Pacific, we found a larger impact ( -37\%, Table 1) but with opposite sign,
because the mixing effect is larger than the wind effect. In that respect, our results are
thus significantly different from the previous estimates.
Moreover, it was hypothesized that the year-to-year differences in TC frequency and intensity might be an important mechanism for controlling interannual variability in $F_{CO_2}$ [Bates et al., 1998; Bates, 2007], although this hypothesis was not supported by the model result of Koch et al. [2009] in the subtropical North Atlantic. To examine this question over different ocean sub-basins, we compared the standard deviation of the year-to-year regional $F_{CO_2}$ budget to the standard deviation of the year-to-year regional TC-induced $F_{CO_2}$ flux (Table 1). Our result extend the conclusion of Koch et al. [2009]: over the different sub-basins, TC explain between 0 and 10% of the interannual flux changes, with a global mean of $\sim 5\%$.

5.2. The post-storm, mixing effect

One of the important process highlighted in this study is the impact of TC-induced vertical mixing with thermocline waters, which occurs essentially after the storm, cools the SST and reduces the pCO$_2$. The importance of this processes is revealed by the persistence of the cold SST anomaly after the passage of the storm that can last from a few days to over a month. This persistence is accurately reproduced by our model when compared with satellite AMSR observations. The link between the cooling and the change in pCO$_2$ is however not straightforward. Figure 12 shows the distribution of the change in pCO$_2$ after the passage of TC, against the change in SST (computed following section 2.3 as $\Delta pCO_2$ versus $\Delta T$). It confirms that in most cases, pCO$_2$ decreases after the passage of TCs; the average decrease (-7.6 $\mu$atm per °C) is however smaller than the theoretical decrease due to the temperature effect alone (-13 $\mu$atm per °C, Sarmiento and Gruber [2006]). This feature can be explained by the entrainment of higher DIC water from the thermocline into the surface mixed-layer which increases the pCO$_2$ and partly offsets the
temperature effect; TC can also alleviate nutrient limitation and promote phytoplankton blooms [Babin et al., 2004; Walker et al., 2005; Son et al., 2006; Patra et al., 2007; Liu et al., 2009; Hanshaw et al., 2008; Gierach and Subrahmanyam, 2008], thus decreasing the pCO2. How mixing of these different elements impact pCO2 depends on their respective vertical gradients and is spatially and temporally variable [Mahadevan et al., 2011]. Our model results are in that respect consistent with the analysis of Mahadevan et al. [2011], based on climatological distributions of T, S, DIC, nutrients and TA, which suggest that the temperature impact on pCO2 prevails upon the change of the other variables (S, TA, DIC) in the tropical oceans and during the summer cyclonic season. However, a sharp rise in pCO2 has recently been reported after the passage of typhoon Choi-Wan in 2009 in the North West Pacific, which preceded the drop in temperature by about 6h [Bond et al., 2011]. Such short-term sequence could not be examined with the daily resolution of our model outputs and deserves further investigation.

5.3. Caveats

Our results rely on an number of assumptions that were made to build the model. Decreasing atmospheric pCO2 during the passage of TCs, associated with the decrease atmospheric pressure, have been reported [Bates et al., 1998; Huang and Imberger, 2010], with values up to 20 μatm (Sarma, unpublished data). In our model, the atmospheric pCO2 is increased from year to year but is not impacted by TCs. The model study of Huang and Imberger [2010], which accounts for that effect, suggest that it is very short-lived (<12h) and limited to the core of the TC (in a 50 km radius); we thus expect this effect to be weak when averaged over the large oceanic area influenced by TCs.
Our results are also certainly dependent on the choice of the parameterization of the gas transfer velocity at high wind speed, as suggested by the sensitivity analysis of Perrie [2004]. A different choice might imply larger absolute values of the fluxes, and modulate the strength of the compensations.

Another factor that was not accounted for in this study is the intense rain falls that often come with TCs, and can dilute the salinity and DIC, with possible modifications of the surface pCO$_2$. A more precise estimate of the TC induced $F_{CO_2}$ would certainly require to examine this aspect.

6. Conclusion

In this study, we used a state-of-the-art global ocean biogeochemical model driven by TC wind forcing derived from a historical TC database. This allows us to examine the ocean response along 1,663 TC tracks over the 1993-2007 period. The resulting modeled SST response to TCs compares very well with satellite estimates during the same period. The model also accurately reproduces the sharp pCO$_2$ drawdown recorded after the passage of Hurricane Felix in 1995 in the NW Atlantic.

Previous case studies have illustrated the strong influence of TCs on ocean-atmosphere CO$_2$ fluxes by increasing the gas exchange coefficient and decreasing the SST and oceanic pCO$_2$. Moreover, it was generally assumed that the ocean is oversaturated in CO$_2$ under TCs, because TCs blow essentially in the tropics during the summer season. Thus, based on these few observations, it has been suggested that TCs significantly increase the CO$_2$ efflux from the ocean to the atmosphere. However, limited availability of pCO$_2$
observations under TCs harsh conditions has so far restricted global quantification of the TC-induced $F_{CO_2}$ to hazardous extrapolations.

We found a similar proportion of TCs over undersaturated regions ($\sim 20\%$ of TC locations are over regions where $DpCO_2 < -10\mu atm$) and oversaturated regions ($\sim 30\%$ for $DpCO_2 > 10\mu atm$), with a large proportion ($\sim 50\%$) of TCs over regions where the ocean and atmosphere are in near equilibrium. This estimation, based on our model outputs, is consistent with a similar estimation from observed climatological pCO$_2$ variations. The consequence is that TCs can generate instantaneous CO$_2$ fluxes directed from the ocean to the atmosphere (efflux) or vise-versa (influx), depending on the oceanic condition at the time of the TC passage; this instantaneous flux is very weak in $\sim 50\%$ of the cases.

Moreover, we identify two competing effects of TCs on $F_{CO_2}$ that are not synchronous. During the storm and depending on the sign of the difference of pCO$_2$ between the ocean and atmosphere, TCs are responsible for large efflux or influx anomalies due to the strong winds. During several weeks after the storm, oceanic pCO$_2$ is reduced in response to vertical mixing, which systematically causes an influx anomaly. Generally, the storm wind-effect and post-storm mixing effect have the same order of magnitude. This implies that, contrary to previous estimates, TCs weakly impacts the CO$_2$ efflux because the two effects oppose with each other when they blow over supersaturated areas (typically in the North Atlantic, North-East Pacific, Arabian Sea). In contrast, TCs increase the CO$_2$ influx because the two effects add up under undersaturated conditions (e.g. in the Westernmost North Pacific, Bay of Bengal, South Indian and Pacific Ocean). In total, we find that TCs account for $\sim 2\%$ of the $F_{CO_2}$ during the cyclonic season over the tropical ocean ($40^\circ N-40^\circ S$). This is an order of magnitude less than the lowest previous estimates.
Moreover, we find that regionally, TC account for 0 to 10% of the year-to-year variations of the $F_{CO_2}$.

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Total flux | TC-flux | % TC/Total | STD (Total/TC)
---|---|---|---
North Atlantic | -65.7 (-2699) | -0.6 | +1% | 11.1/1.8
North East Pacific | -130.0 (-4988) | -3.0 | +2.3% | 24.3/2.9
North West Pacific | +15.5 (+384) | +5.2 | +33.4 % | 21.8/2.3
Bay of Bengal | -0.6 (-117) | +0.1 | -13.6 % | 4.3/0.3
Arabian Sea | -43.4 (-2727) | -0.1 | +0.2 % | 6.1/0.2
South Indian | +93.1 (+2838) | +2.7 | +2.9 % | 26.8/1.2
Australia | +18.1 (+2017) | +0.5 | +2.5 % | 5.0/0.4
South Pacific | +53.3 (+2492) | +1.7 | +3.2% | 18.2/ 1.0
Total | -305.0 (-1258) | +6.7 | -2.2% | 85.0/5.0

Table 1. Regional and global CO₂ fluxes and contribution of TCs. Mean and standard deviation of total and TC-induced carbon flux cumulated over the cyclonic season and over different oceanic basins (delimited in Fig. 11), in Tera gC/cyclonic season. The STD represents interannual variations. The Total flux is computed over the region 40S-40N for the cyclonic season (NDJFMA for Southern Hemisphere and MJJASO for Northern Hemisphere). Value in parenthesis in the first column are in mgC/m².

<table>
<thead>
<tr>
<th>TC-flux ΔpCO₂ &gt; 0</th>
<th>TC-flux ΔpCO₂ &lt; 0</th>
<th>Days ΔpCO₂ &gt; 0</th>
<th>Days ΔpCO₂ &lt; 0</th>
</tr>
</thead>
<tbody>
<tr>
<td>North Atlantic</td>
<td>+1.0</td>
<td>-1.6</td>
<td>16</td>
</tr>
<tr>
<td>North East Pacific</td>
<td>+0.3</td>
<td>-3.3</td>
<td>14</td>
</tr>
<tr>
<td>North West Pacific</td>
<td>+5.2</td>
<td>0.0</td>
<td>168</td>
</tr>
<tr>
<td>Bay of Bengal</td>
<td>+0.1</td>
<td>0.0</td>
<td>15</td>
</tr>
<tr>
<td>Arabian Sea</td>
<td>0.0</td>
<td>-0.1</td>
<td>0</td>
</tr>
<tr>
<td>South Indian</td>
<td>+2.5</td>
<td>+0.2</td>
<td>77</td>
</tr>
<tr>
<td>Australia</td>
<td>+0.5</td>
<td>0</td>
<td>19</td>
</tr>
<tr>
<td>South Pacific</td>
<td>+1.5</td>
<td>+0.2</td>
<td>45</td>
</tr>
<tr>
<td>Total</td>
<td>11.3</td>
<td>-4.6</td>
<td>354</td>
</tr>
</tbody>
</table>

Table 2. Global and Regional CO₂ fluxes due TCs depending on oceanic state. Mean TC-induced carbon flux cumulated over the cyclonic season (NDJFMA for Southern Hemisphere and MJJASO for Northern Hemisphere) and over different oceanic basins (delimited in Fig. 11). For each region, we count the number of TCs occurrences (In days/cyclonic season) and associated carbon flux (in Tera gC/cyclonic season) over undersaturated (versus oversaturated) oceanic state.
Figure 1. Illustration of forcing strategy to include TCs-related winds. Snapshot of the wind forcing on the 7th of April 1978 in the South-West Indian Ocean for: (a) CORE forcing; (b) along-track filtered CORE forcing (N-run); and (c) along-track filtered CORE forcing superimposed with analytic TC vortices computed using IBTRACS TC location and maximum wind (C-run).
Figure 2. Evaluation of the timing and amplitude of surface temperature response to TCs (a) Temporal evolution of the mean TC-induced cooling along TC tracks for model and observations over the period 1998-2007. Shading indicates the spread around the mean value, calculated as ± 1/2 quartile. (b) Probability Density function of simulated (C-run minus N-run) versus observed (TMI-AMSR) cold wake anomalies for the same period.
Figure 3. Evaluation of the spatial distribution of the surface temperature response to TCs. Spatial map of averaged cold wake anomaly amplitude (in °C) during the cyclonic season over the 1998-2007 period for (a) TMI-AMSR observations and (b) C-run.
Figure 4. Evaluation of DpCO$_2$ during cyclonic season. DpCO$_2$ climatology (in μatm) during the cyclonic season (MJJASO in the northern hemisphere and NDJFMA in the southern hemisphere) from (a) Observed estimates [Takahashi et al., 2009] and (b) the model (C run) over the 1993-2007 period. The grey band is to mark the time discontinuity between the two hemispheres.
Figure 5. Hurricane Felix test case (BATS, August 1995). Temporal evolution of (a) SST and (b) oceanic pCO2 (in μatm) in the C run (red), the N run (blue) and shipboard measurements from Bates et al. [1998] (dots) during the passage of Hurricane Felix in the Sargasso Sea near Bermuda in August 1995. (c) Temporal evolution of the air-sea CO2 flux (negative when oriented from the ocean to the atmosphere in mmole/m²/day) for the C run (red) and the N run (blue) (d) Temporal evolution of the flux difference between the C and N runs (red), flux anomaly related to the TC wind effect (green), flux anomaly related to the TC mixing effect (blue) (computed according to section 2.3).
Figure 6. Distribution of DpCO$_2$ under TCs. Histograms of (a) DpCO$_2$ (in $\mu$atm) and (b) TC-related air-sea flux anomaly (in mmole/m$^2$/day) under TCs at time of maximum wind intensity for 46951 TC locations along 1663 TC tracks over the 1993-2007 period. Blue (red) curve on panel (b) correspond to the histogram of TC-related air-sea flux anomaly under cyclones where DpCO$_2$ < 10 $\mu$ atm (DpCO$_2$ > 10 $\mu$ atm) corresponding to 30.2% (19.8%) of the cases.
Figure 7. Locations where DpCO$_2$ is positive and negative under TCs. Density distribution (in cyclone-day/year) for TCs where (a) DpCO$_2$ > 0 and (b) DpCO$_2$ < 0 at t=t$_0$ in the C-run. Densities are computed over the 1993-2007 period. t$_0$ is the time of maximum wind intensity. (c) and (d): same diagnostics applied to the DpCO$_2$ climatology of Takahashi et al. [2009] interpolated in time.
Figure 8. Composite time evolution of wind speed, SST, DpCO₂ and air-sea flux anomalies generated by the passage of TCs. Blue color: Mean composite of all cases (25773) where DpCO₂ < 0 at t=t₀, with t₀ the time of maximum wind intensity. Red color: Mean composite of all cases (21178) where DpCO₂ > 0 at t=t₀. Shading show the ± 1/2 standard deviation around the mean composite value. Composites are computed over the 1993-2007 period. Anomalies are computed according to section 2.3.
Figure 9. **Schematic representation of the impact of TCs on pCO$_2$ and air sea flux.**

Schematic time evolution of oceanic pCO$_2$ and air-sea CO$_2$ flux associated with the passage of a TC (black curves). For reference, the grey curves show the stationary situation in the absence of TC. Day $t_0$ is the time of maximum wind intensity during the passage of TC. The left column is the case where the oceanic pCO$_2$ is initially larger than the atmospheric pCO$_2$ and the air-sea flux is directed from the ocean to the atmosphere (case of BATS). The right column is when the atmospheric pCO$_2$ is initially larger than the oceanic pCO$_2$ and the air-sea flux is directed from the atmosphere to the ocean.
Figure 10. Composite time evolution of TC-induced air-sea flux anomalies: controlling mechanisms. Mean of all cases where (a) DpCO\textsubscript{2} < 0 at t=t\textsubscript{0} and (b) DpCO\textsubscript{2} > 0 at t=t\textsubscript{0}. Anomalies are computed over the 1993-2007 period. The black curve is the total TC-induced flux anomaly, the blue curve is the wind effect on the flux (section 2.3, equ. 2), the red curve is the mixing effect on the flux (section 2.3, equ. 3).
Figure 11. Climatological TC-induced air-sea flux. Climatological average of seasonally cumulated air-sea flux (in mmol/m$^2$/season) induced by cyclones along their tracks in the C-run between (a) $t_0-1.5d$ and $t_0+30d$ (b) $t_0-1.5d$ and $t_0+1.5d$ (c) $t_0+1.5d$ and $t_0+30d$. Dashed lines on panel (a) indicate the boundaries of the regions discussed in Tables 1 and 2.
Figure 12. TC-induced DpCO$_2$ anomaly versus SST anomaly. Probability Density function of TC-induced DpCO$_2$ wake anomaly versus SST wake anomaly simulated in the model. Thick black line indicates the theoretical relationship of Gruber and Sarmiento, with a slope of -13. The blue line is the mean $\bar{\Delta}$SST per bin of $\bar{\Delta}$DpCO$_2$. The black line is the linear fit to the blue line and as a slope of -7.6. Anomalies are computed over the 1993-2007 period along TC tracks for run C.