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## Premonsoon/postmonsoon Bay of Bengal tropical cyclones intensity: Role of air-sea coupling and large-scale background state

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# Geophysical Research Letters

## RESEARCH LETTER

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### Key Points:

- Bay of Bengal tropical cyclones are stronger before than after the monsoon, due to more favorable large-scale background conditions
- Air-sea coupling alleviates the effect of large-scale background conditions, by inhibiting premonsoon tropical cyclones
- The air-sea coupling negative feedback on tropical cyclones is weak after the monsoon, due to a fresher, less thermally stratified BoB

### Supporting Information:

- Supporting Information S1
- Figure S1
- Figure S2
- Figure S3
- Figure S4
- Figure S5
- Figure S6
- Figure S7

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## Premonsoon/Postmonsoon Bay of Bengal Tropical Cyclones Intensity: Role of Air-Sea Coupling and Large-Scale Background State

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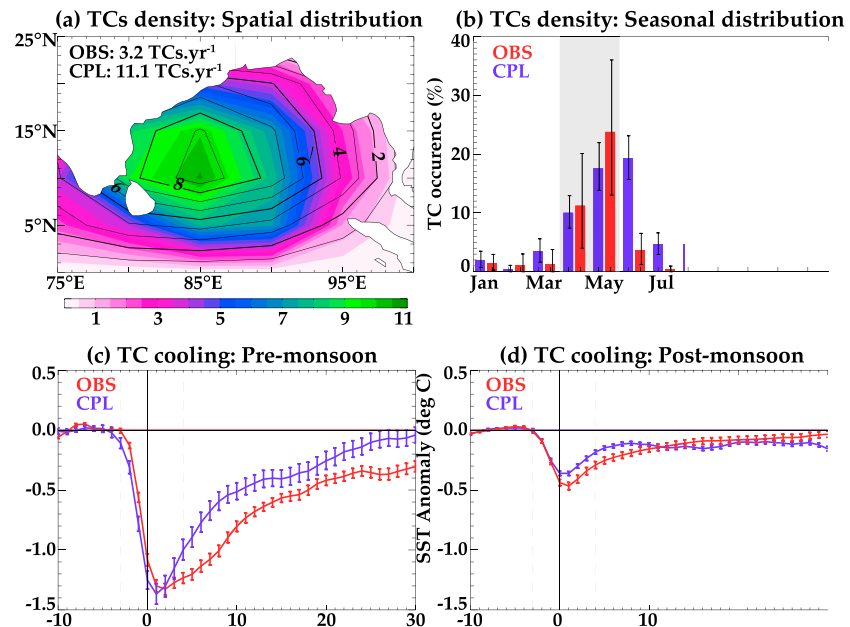
**Abstract** The densely populated Bay of Bengal (BoB) rim witnesses the deadliest tropical cyclones (TCs) globally, before and after the summer monsoon. Previous studies indicated that enhanced salinity and reduced thermal stratification reduce cooling under BoB TCs after the monsoon, suggesting that air-sea coupling may favor stronger TCs during that season. Using observations and simulations from a one fourth degree regional ocean-atmosphere model, we show that BoB TCs are stronger before the monsoon due to a more favorable large-scale background state (less vertical wind shear and higher sea surface temperature). Air-sea coupling however alleviates this background state influence, by reducing the number of premonsoon intense TCs, because of larger cooling and reduced upward enthalpy fluxes below TCs during that season. As the impact of air-sea interactions on BoB TCs is largest for intense TCs, it should be further investigated for Category 3 and above TCs, which are not reproduced at one fourth degree resolution.

**Plain language summary** Tropical cyclones that develop in the Bay of Bengal are amongst the most lethal globally, owing to the dense and vulnerable coastal population living along its rim. These cyclones mostly occur in April–May and October–December, that is, before and after the Indian summer monsoon. In this study, we show that there are more intense cyclones before than after the monsoon, because of more favorable large-scale background conditions. We however also show that more intense interactions between the cyclone and the ocean tend to reduce risks for intense cyclones before the monsoon, hence opposing the effect of background conditions. Since air-sea coupling clearly contributes to the Bay of Bengal cyclone intensity, in particular before the monsoon, it should be accounted for in operational forecasts.

## 1. Introduction

Tropical cyclones (TCs) are the most devastating natural disasters in coastal regions (Blake et al., 2011; Pielke et al., 2008). Although the Bay of Bengal (BoB) accounts for only 5% of TCs globally, its rim gathers 80% of TC-induced fatalities worldwide (Needham et al., 2015). The Category 3 Bhola TC in November 1970 was, for instance, the deadliest cyclone in the recent history, killing ~500,000 people in Bangladesh (Islam & Chik, 2011; Madsen & Jakobsen, 2004). The BoB rim is particularly vulnerable to TCs because of its densely populated low-lying areas and a very shallow continental shelf that amplifies storm surges (up to 14 m; Jakobsen & Azam, 2006). This exposure to coastal flooding is expected to further increase over the coming decades due to more numerous extreme TCs and rising sea level associated with climate change (e.g., Karim & Mimura, 2008). These tremendous socio-economic implications call for a better understanding of the mechanisms of TC intensification in the BoB, in order to improve TC intensity operational forecasts and projected changes in a warming world.

TCs mostly occur over the southwestern BoB (Figure 1a), with a clear seasonality characterized by a double peak during the premonsoon (April–May) and postmonsoon (October–December, Figure 1b). The postmonsoon exhibits a larger TC density (~18 TC days/year) than the premonsoon (~11 TC days/year), with very few events during the rest of the year (Figure 1b). This distinct bimodal distribution is attributed to seasonal changes in the main large-scale environmental parameters that affect TC development. Large vertical wind shear is a long-known cause for suppressing TCs and has been underlined as a major player for the absence

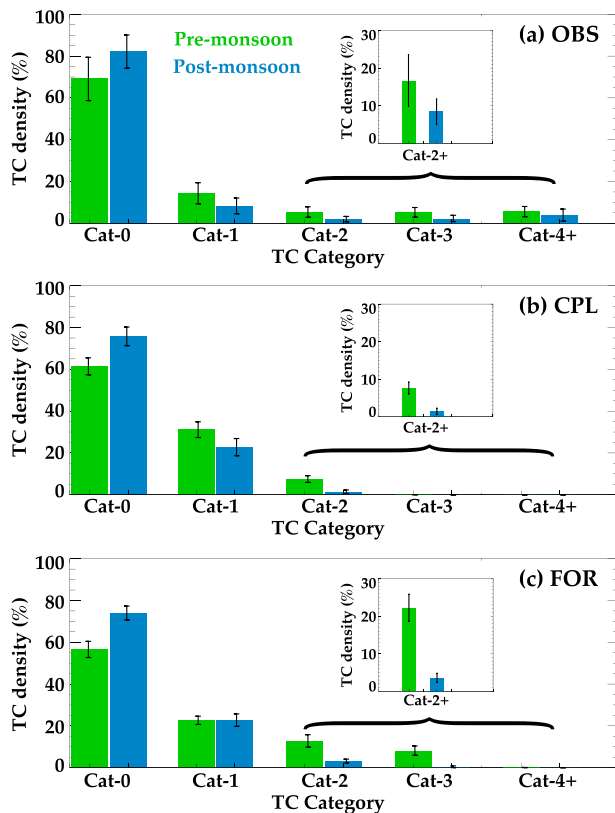


**Figure 1.** (a) Climatological distribution of normalized TC density (i.e., percent of the total occurrence rate at each location) for observations (color) and CPL simulation (contour). (b) Climatology of the fraction of TCs that occur at a given month in the BoB for observations (red bars) and CPL (blue bars). The two shaded regions highlight the premonsoon and postmonsoon TC season definitions used in this paper. Composite evolution of the SST cooling within 200 km of all TCs in the BoB (in degree Celsius) during the (c) premonsoon and (d) postmonsoon for observations (red) and CPL simulation (blue). The whiskers in Figures 1b–1d display the 90% confidence intervals based on a bootstrap method. TCs = tropical cyclones; CPL = coupled model simulation; BoB = Bay of Bengal.

of TCs during the southwest monsoon (Li et al., 2013; Xing & Huang, 2013; Yanase et al., 2012). Li et al. (2013) however stressed out that the decreased absolute vorticity and sea surface temperature (SST) during the monsoon also contribute to suppress TCs during the monsoon (a colder SST reduces the TC maximum potential intensity (MPI); e.g., Emanuel, 1986). On the other hand, the midtropospheric humidity increase during the monsoon would tend to favor more TCs, offsetting the effect of the vertical shear, vorticity, and SST changes (Li et al., 2013).

While the above studies have mostly focused on changes in large-scale environmental parameters, some studies have argued that seasonal changes in oceanic stratification may also matter, by influencing air-sea interactions below TCs. TCs cool the ocean under their passage, mostly through vertical mixing (e.g., Price, 1981; Vincent, Lengaigne, Madec, et al., 2012). This cooling feeds back negatively on the TC intensity, since the colder SST reduces the upward enthalpy fluxes that allow TCs to intensify (e.g., Cione & Uhlhorn, 2003; Schade, 2000). In the BoB, this TC-induced cooling is 3 times larger before than after the monsoon, for a given TC strength (Neetu et al., 2012; Sengupta et al., 2008). Neetu et al. (2012) attributed this to the seasonal changes in the upper ocean stratification between the premonsoon and postmonsoon. The postmonsoon indeed both displays a stronger salinity stratification (due to the large freshwater flux into the Bay during and shortly after the monsoon; Akhil et al., 2014; Chaitanya et al., 2014; Shenoi et al., 2002) and a deeper thermocline along the BoB rim (in response to coastal Kelvin waves originating from the equatorial region; McCreary et al., 1996) than the premonsoon. Balaguru et al. (2012) have underlined that TCs that travel over a strong salinity stratification tend to be associated with weaker cold wakes, stronger upward enthalpy fluxes, and stronger intensification rates. Sengupta et al. (2008) have more specifically shown that BoB TCs yield a weaker cold wake after the monsoon and wondered whether this should favor more intense TCs by reducing the negative feedback due to air-sea coupling. Observations however indicate stronger TCs before the monsoon (Figure 2a; Yanase et al., 2012; Li et al., 2013), hence calling for a more detailed investigation in order to ascertain whether the negative feedback on BoB TCs is indeed weaker after the monsoon.

In a recent study, Lengaigne et al. (2018) investigated the influence of air-sea coupling on TC characteristics in both hemispheres of the Indian Ocean, comparing a one fourth degree regional coupled ocean-



**Figure 2.** Percentage of BoB TCs as a function of their mean wind intensity during the premonsoon (green) and postmonsoon (blue) for (a) observations, (b) CPL, and (c) FOR simulations. The inset indicates the overall percentage of intense TCs (defined as Category 2 and above). The whiskers display 90% confidence intervals based on a bootstrap method.

TCs = tropical cyclones; CPL = coupled model simulation; BoB = Bay of Bengal; FOR = uncoupled atmospheric simulation.

atmosphere model 20-year-long simulation with a twin forced atmospheric experiment. They found that coupling reduces the number of TCs and their intensity in both hemispheres, as the TC-induced cooling reduces the upward enthalpy fluxes under TCs. They further found that this reduction was particularly large for the northern Indian Ocean most intense TCs, due to higher ambient surface temperatures and larger TC-induced cooling there. Lengaigne et al. (2018) hence suggest that air-sea coupling is very efficient in modulating northern Indian Ocean TC intensity. They did however not investigate if air-sea coupling allowed stronger postmonsoon TCs due to seasonal changes in ocean stratification, as hinted by Sengupta et al. (2008) and Balaguru et al. (2012).

While a couple of previous studies did investigate the effect of atmospheric parameters on BoB premonsoon and postmonsoon cyclones (e.g., Li et al., 2013; Yanase et al., 2012), they did not examine the effect of air-sea coupling in detail. The present study aims at providing an in-depth analysis of the premonsoon/postmonsoon differences in air-sea coupling below TCs and the underlying mechanisms. More specifically, we quantify the respective influences of large-scale background conditions and air-sea coupling on the premonsoon and postmonsoon BoB TC intensity distribution using the same set of coupled and forced experiments as in Lengaigne et al. (2018). Considering the effect of air-sea coupling and large-scale background state together will allow to understand why stronger TCs occur *before* the monsoon, while previous studies (Neetu et al., 2012; Sengupta et al., 2008) hint toward air-sea coupling favoring stronger cyclones *after* the monsoon. The paper is organized as follows. Section 2 summarizes the modeling framework and observed data sets. Section 3 assesses the impact of large-scale background conditions and air-sea coupling on BoB TC intensity. A discussion of our results against relevant literature is provided in section 4.

## 2. Data Set and Methods

We use an Indian Ocean [25.5°E–142.5°E, 34.5°S–26°N] configuration of the NEMO-OASIS3-WRF regional coupled model at one fourth degree horizontal resolution. NEMO-OASIS3-WRF couples the NEMO (Nucleus for European Modeling of the Ocean) ocean component (Madec, 2008) to the WRF (Weather Research and Forecasting Model) atmospheric component (Skamarock & Klemp, 2008) through the OASIS3 coupler (Valcke, 2013). This configuration has been extensively described and validated in Samson et al. (2014) and Lengaigne et al. (2018). We therefore only provide a brief summary here. The ocean component has 46 vertical levels, with an enhanced 5-m resolution in the upper ocean. The atmospheric component has 28 sigma vertical levels, with a higher resolution of 30 m near the surface, and uses the updated Kain-Fritsch atmospheric convective scheme (Kain, 2004). Lateral boundary conditions are, respectively, supplied from a global simulation forced by the Drakkar data set (Brodeau et al., 2010) for NEMO and from the European Centre for Medium-Range Weather Forecasts ERA-Interim reanalysis (Dee et al., 2011) at 6 hourly for WRF.

We use the same twin-experimental setup as in Lengaigne et al. (2018). A 21-year coupled simulation (CPL) is first performed using the above 1989–2009 lateral boundary conditions. To isolate the effect of air-sea coupling on TCs, an uncoupled atmospheric simulation (FOR) is performed with the same atmospheric boundary conditions as CPL, but forced by SST from the CPL experiment, from which TC-induced cold wakes have been suppressed (see Lengaigne et al., 2018 for details). As a result, the FOR simulation does not account for any oceanic feedback on TCs. The two experiments have a very similar atmospheric background state over the BoB, in particular for variables that have a well-documented influence on TCs (supporting information Figure S1), so that the TC differences between the two experiments are dominated by the effects of air-sea coupling (also see Lengaigne et al., 2018). Due to chaotic nature of the atmosphere, the small

perturbations induced by the removal of TCs cold wakes yield different TC tracks between the FOR and CPL simulations. While individual TC tracks are different in the two simulations, we will examine statistical differences in, for example, the number of cyclones or cyclone intensity distribution in the rest of the paper. Lengaigne et al. (2018) indeed demonstrated that comparing the statistics of simulated TCs in these twin experiments allows diagnosing the influence of air-sea coupling on TCs.

The first year of each experiment is discarded: the simulations are hence analyzed over a 20-year period. TCs from both simulations are tracked using the same methodology as in Samson et al. (2014) and Lengaigne et al. (2018) using criteria similar to those considered in previous studies (Jourdain et al., 2011; Jullien et al., 2014):

1. 10-m wind  $>17.5$  m/s associated with a local minimum in sea-level pressure,
2. 850-hPa vorticity  $>3 \times 10^{-4} \text{ s}^{-1}$ , and
3. 700- to 300-hPa mean temperature anomaly  $>1^\circ\text{K}$ .

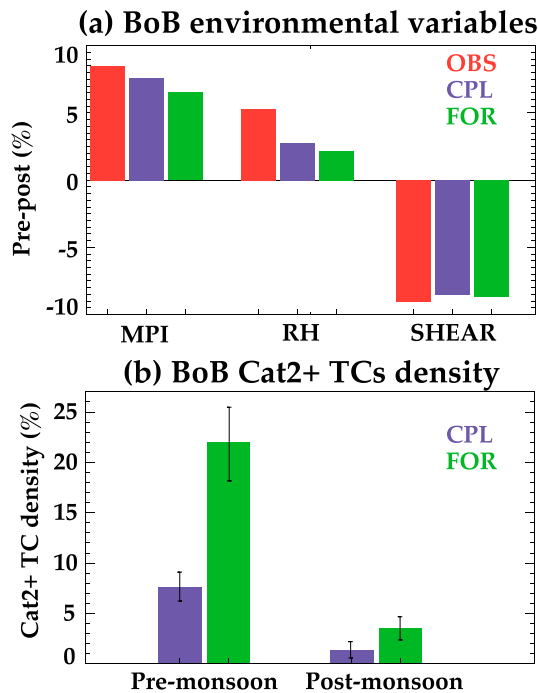
TCs are defined as sustained 10-m winds more than 17.5 m/s, and TC intensities are classified using the Saffir-Simpson scale. TC temperature anomalies are calculated with respect to their large-scale environment: the TC region is defined as within 3 radii of maximum wind around TC center, while the environmental temperature is averaged between 6 and 9 radii of maximum wind. TC tracks shorter than 1 day are eliminated.

The large-scale background control on TC intensification is assessed by analyzing the three major parameters known to influence TC intensification in the BoB (Li et al., 2013; Xing & Huang, 2013; Yanase et al., 2012): the vertical wind shear between 850 and 200 hPa in meter per second (SHEAR), the midtropospheric relative humidity (RH) at 600 hPa, and the MPI (calculated using a routine available at <http://emanuel.mit.edu/products> and closely related to SST, with high SSTs favoring larger MPIs). Following Knaff et al. (2005), Knaff and Sampson (2009) and Neetu et al. (2017) RH and SHEAR are averaged over an annual region within 200–800 km from the center of the storm. This averaging accounts for the fact that environmental parameters can influence the storm far away from its center, while excluding TC signals close to the TC center. These parameters are calculated over 1990–2009 from simulations and from the ERA-Interim atmospheric Reanalysis (Dee et al., 2011). Observed TC positions and magnitudes are derived from the International Best Track Archive for Climate Stewardship (Knapp et al., 2010). Observed SST response under TCs is characterized using a blend of Tropical Rainfall Measuring Mission Microwave Image and advanced Microwave Scanning Radiometer SST data (available from [http://www.ssmi.com/sst/microwave\\_o\\_i\\_sst\\_data\\_description.html](http://www.ssmi.com/sst/microwave_o_i_sst_data_description.html)) as in Neetu et al. (2012), Vincent, Lengaigne, Madec, et al. (2012), and Vincent, Lengaigne, Vialard, et al. (2012).

### 3. Results

We first assess the ability of the reference coupled model simulation (CPL) to reproduce the main seasonal characteristics of BoB TCs. In observations, the two TC-prone periods (defined as months with more than 5% of the annual average TC-day count) are April–May and October–December (Figure 1b), and we will use those definitions in the following for the premonsoon and postmonsoon seasons. In line with observations, the model also displays a clear bimodal TC seasonal distribution, with few TCs from January to March and during the core of the summer monsoon (July–August). The model seasonal TCs distribution in the BoB however displays some systematic biases compared to observations, including an overestimation of TCs occurrence during the summer monsoon (June–September) and an underestimation in November–December. All the results in this study remain qualitatively similar if the premonsoon and postmonsoon seasons are rather defined based on the model seasonal TC distribution (e.g., April–June and September–December). Aside from the bimodal seasonal distribution, the CPL experiment is also able to reproduce TCs spatial distribution very well (0.94 pattern correlation between observation and CPL; Figure 1a). As already mentioned in previous papers (Samson et al., 2014; Lengaigne et al., 2018), the model however strongly overestimates cyclogenesis in the BoB, with an average 11.1 TCs/year compared to 3.2 TCs/year in observations. We will discuss the potential implications of our model biases in section 4.

We will now assess how the CPL experiment captures salient features of the observed BoB TC-induced cooling, especially their premonsoon and postmonsoon amplitude difference. Figures 1c and 1d display



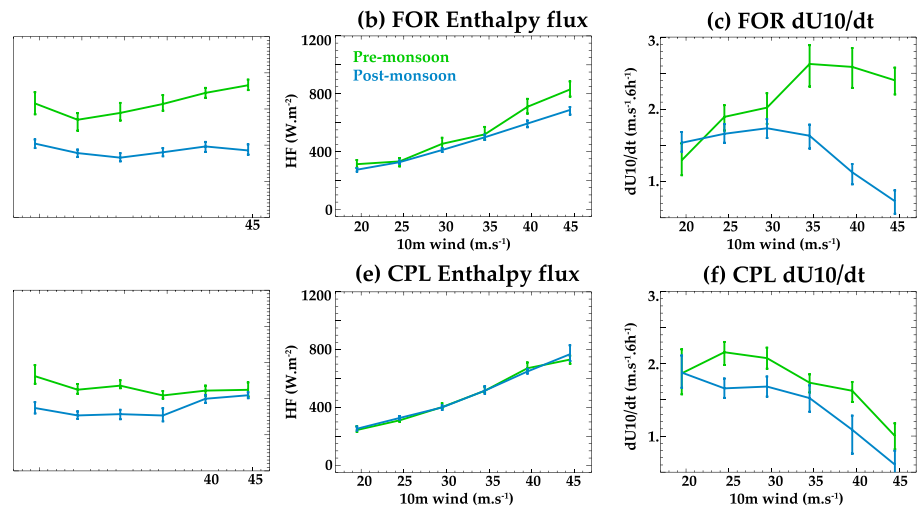
**Figure 3.** (a) Premonsoon minus postmonsoon BoB-averaged maximum potential intensity (MPI), relative humidity at 600 hPa (RH), and vertical wind shear (SHEAR) in percent for ERA-I (red), CPL (blue), and FOR (green). (b) Percentage of BoB intense TCs (Category 2 and above) for CPL and FOR during the premonsoon and postmonsoon. The whiskers display 90% confidence intervals based on a bootstrap method. TCs = tropical cyclones; CPL = coupled model simulation; BoB = Bay of Bengal; FOR = uncoupled atmospheric simulation.

premonsoon and postmonsoon composites of the cooling under TCs in observations and the CPL simulation. In agreement with previous results (Neetu et al., 2012; Sengupta et al., 2008), the observed BoB TC-induced cooling is on average 2 to 3 times larger before (up to  $-1.3$  °C; Figure 1c) than after the monsoon (up to  $-0.5$  °C; Figure 1b). The CPL simulation accurately captures the timing and seasonal contrast of the SST response to TCs, with a weaker cooling during the postmonsoon (up to  $-0.4$  °C) than during the premonsoon (up to  $-1.3$  °C). Neetu et al. (2012) attributed this behavior to stronger upper ocean stratification during the postmonsoon. In agreement with their analyses (their Figure 9), the SST response in CPL simulation is larger during the premonsoon than during the postmonsoon for a given amount of energy transferred to the ocean by a TC (Figures S2a and S2b in the supporting information), indicating that the premonsoon oceanic stratification favors a larger cooling through vertical mixing (Neetu et al., 2012; Vincent, Lengaigne, Vialard, et al., 2012). The *cooling inhibition index*, a proxy of the oceanic resistance to mixing-induced cooling (Vincent, Lengaigne, Madec, et al., 2012, more details in the supporting information), is indeed higher during the postmonsoon in CPL, in good agreement with observations (Figures S2c and S2d), associated with a stronger salinity and weaker thermal stratification (Figure S3).

We finally assess how the CPL experiment captures the BoB premonsoon and postmonsoon TC intensity distributions (Figures 2a and 2b). In line with previous studies (Li et al., 2013; Yanase et al., 2012), Figure 2a indicates that observed TCs are stronger before than after the monsoon. The CPL simulation reproduces this seasonal contrast, with a larger proportion of strong (Cat-2 and above) TCs before than after the monsoon (8% vs 2% in the model and 17% vs 8% in observations). Due to the limited sample size, this difference is only significant at the 84% level in observations while the larger model sample (due to the overestimated model TC occurrence rate) yields a 99% significance. While there are TCs up to Cat-5 in observations, the model rarely simulates TCs above Cat-2. As acknowledged in other studies using coupled models with a similar spatial resolution in other basins (e.g., Jullien et al., 2014; Ogata et al., 2015), this inability to simulate the most intense TCs is likely related to the relatively coarse horizontal resolution of our atmospheric component (one fourth degree): an atmospheric resolution below 10 km is indeed needed to represent the most intense TCs (Jung et al., 2012; Manganello et al., 2012; Murakami et al., 2015; Satoh et al., 2012).

Despite some shortcomings, the CPL simulation reproduces the BoB TCs spatial and seasonal distribution reasonably well, including the larger cooling under TCs and more intense TCs before the monsoon. This allows us to investigate the relative roles of air-sea coupling and large-scale background parameters in controlling the premonsoon/postmonsoon TC intensity contrasts in this model. Premonsoon TCs are stronger despite a more intense cooling, in both observations and the CPL experiment. This indicates that if air-sea coupling plays any role in the pre-post monsoon TC intensity contrasts, it is not the dominant factor. The FOR simulation also confirms that by exhibiting stronger TCs before the monsoon (Figure 2c). This indicates that the large-scale background state is the main factor that yields more intense TCs before the monsoon.

The premonsoon minus postmonsoon differences of the most important large-scale background parameters affecting TC intensification in the BoB are further analyzed in Figure 3a. Large MPI, midtropospheric RH, and weak SHEAR are favorable to TC intensification (DeMaria & Kaplan, 1994; Merrill, 1988). The premonsoon/postmonsoon changes in MPI, RH, and SHEAR are very consistent in CPL, in FOR, and in observations. The seasonal changes in these three environmental parameters favor more intense TCs before the monsoon:  $\sim 10\%$  higher MPI,  $\sim 10\text{--}15\%$  higher RH, and  $\sim 18\%$  lower SHEAR. The most likely cause for stronger premonsoon cyclones is thus the larger MPI, more humid midtroposphere, and weaker vertical wind shear during that season. However, note that the effect of tropospheric humidity is not robust as it



**Figure 4.** Mean inner-core (i.e., within 200 km of the TC center) (left column) SST, (middle column) upward surface enthalpy flux, and (right column) TC intensification rate as a function of the TC maximum wind speed for intensifying BoB TCs in (top row) FOR and (bottom row) CPL during the premonsoon (green) and postmonsoon (blue). This figure was obtained through averaging within 5 m/s TC maximum wind bins. The intensification rate is calculated as a centered difference using the TC maximum wind speed 12 hr before and after any given time. TCs = tropical cyclones; CPL = coupled model simulation; BoB = Bay of Bengal; FOR = uncoupled atmospheric simulation; SST = sea surface temperature.

depends on the averaging strategy to produce Figure 3a, as we will discuss in section 4. The increase in MPI is related to  $\sim 1.2$  °C warmer SST before the monsoon, in both observations and the model.

While the analyses above indicate that large-scale changes dominate the premonsoon/postmonsoon changes in TC intensity, this does not imply that air-sea coupling does not affect the BoB TC intensities. The effect of air-sea coupling on the TC intensity changes between premonsoon and postmonsoon can be isolated by comparing the TC statistical properties in the CPL and FOR twin experiments. Air-sea coupling under TCs does not affect the BoB TC spatial distribution but slightly reduces the overall frequency of occurrence of TCs (Figures 1a and S3a), as already documented in Lengaigne et al. (2018). This reduction is in particular a bit more prominent during the southwest monsoon, yielding a slightly more realistic bimodal seasonal distribution in CPL than in FOR (Figures 1b and S3b).

Let us now investigate the premonsoon/postmonsoon contrasts in air-sea coupling under TCs. The larger oceanic cooling before the monsoon (Figures 1c and 1d) is indicative of a stronger air-sea coupling negative feedback on TC intensity before the monsoon. Figure 3b displays the percentage of intense TCs (Cat-2 and above) during the premonsoon and postmonsoon for the CPL and FOR experiments. Air-sea coupling yields a much larger reduction of the simulated intense TC proportion before than after the monsoon. This suggests a strong negative feedback of air-sea coupling on TCs before the monsoon, while this effect is weaker after the monsoon. The effect of air-sea coupling is hence to reduce the intensity of premonsoon TCs. This effect is however offset by the stronger impact of changes in MPI, RH, and SHEAR, which all tend to induce stronger premonsoon TCs.

Figure 4 displays FOR and CPL TC composites of SST, upward enthalpy fluxes (i.e., the source of energy feeding TCs), and intensification rate under premonsoon and postmonsoon intensifying TCs as a function of the TC intensity. There is a decrease of the intensification rate for strongest TCs in FOR, both before and after the monsoon, which is simply due to the fact that very strong TCs rarely further intensify. In FOR, the differences between the premonsoon and postmonsoon TC intensity are entirely due to environmental parameters, that is, the weaker vertical shear, more humid troposphere, and warmer SST (Figure 3a) before the monsoon. The warmer premonsoon SST is also visible in Figure 4a. Those more favorable environmental parameters yield a larger intensification rate for most TCs, especially the most intense (Figure 4c). Warmer SSTs before the monsoon (Figure 4a) induce stronger evaporation rates and upward latent heat fluxes for a given wind intensity (Figure 4b), due to the exponential dependence of saturation

vapor pressure on SST in the Clausius-Clapeyron relation. This provides a physical explanation for the process by which the higher MPI yields more TC intensification before the monsoon (Figure 3).

The most relevant large-scale parameters for BoB TC intensification are almost identical in CPL and FOR simulations (Figure 3a and supporting information Figure S1), so that the TC differences between those simulations can be attributed to the effect of air-sea coupling under TCs. Composite maps indicate an average cooling of up to  $-1.2$  °C under TCs during the premonsoon, which is strongly reduced during the postmonsoon (Figures S5f and S6f). Echoing this, the upward enthalpy flux is much more reduced by coupling before (up to  $-180$  W/m<sup>2</sup>) than after (up to  $-100$  W/m<sup>2</sup>) the monsoon (Figures S5i and S6i). This larger premonsoon cooling under TCs offsets the warmer large-scale temperatures. This results in a much weaker difference between premonsoon minus postmonsoon SST under TCs, especially for the strongest simulated TCs (Figure 4d). As a result, the inner core upward enthalpy fluxes are strongly reduced by air-sea coupling during the premonsoon and match those during the postmonsoon (Figure 4e). Consequently, TC intensification rates are reduced during the premonsoon (Figure 4f), especially for strongest TCs (wind speed  $>35$  m/s). Air-sea coupling hence significantly contributes to reduce the number of premonsoon intense TCs (Figures 3b, 4c, and 4f). After the monsoon, the impact of coupling on TCs intensification rates, and hence intensities, is weaker (Figures 3b, 4c, and 4f).

#### 4. Conclusion and Discussion

Based on eight TCs cases, Sengupta et al. (2008) showed that the strong postmonsoon BoB haline stratification inhibits cooling under TCs. This should hence favor more intense postmonsoon TCs, owing to a reduction of the negative feedback associated with air-sea coupling under TCs. Here we use a regional coupled model to investigate whether this is indeed the case. Our coupled and forced simulations confirm that air-sea coupling reduces the premonsoon TC intensity, as stronger TC cold wakes reduce the upward enthalpy fluxes, leading to weaker intensification rates. This negative effect of air-sea coupling on TCs intensity is however offset by the stronger positive effect of seasonal changes in large-scale environmental parameters. In line with Li et al. (2013), TCs are indeed more intense before than after the monsoon in our model and observations, and those more intense premonsoon TCs can be traced back to a warmer ocean (and thus higher MPI), more humid midtroposphere, and weaker vertical shear before than after the monsoon. By quantifying the relative effects of air-sea coupling and large-scale background conditions, our study thus reconciles previous studies, suggesting a weaker effect of air-sea coupling after the monsoon (e.g., Sengupta et al., 2008) with observations of stronger cyclones before the monsoon (e.g., Li et al., 2013).

While the effect of air-sea coupling on premonsoon/postmonsoon TC intensity contrast is weaker than that of environmental parameters, it is far from negligible, especially before the monsoon when the percentage of strong TCs (Cat-2 and above) exhibits a drop from 23% in the forced simulation to 7% in the coupled one. An important implication of our result is that the operational TC forecasts for the northern Indian Ocean should include an oceanic component.

Let us now discuss our results against those of Li et al. (2013). First, Li et al. (2013) suggested higher background RH after than before the monsoon, in contrast to our result. Supporting information Figure S7 demonstrates that the premonsoon minus the postmonsoon RH is sensitive to whether land values are considered or not. The role of MPI and shear in favoring more intense premonsoon TCs is however robust. Li et al. (2013) also partly attributed the more intense observed premonsoon TCs to a greater oceanic heat content during that season. Supporting information Figure S8 demonstrates that the oceanic heat content is indeed higher *before* the monsoon in the model and observations, but that the cooling inhibition of Vincent, Lengaigne, Vialard, et al. (2012)—which tracks the effect of changes in salinity stratification and in the top of the thermocline better—is higher *after* the monsoon, in relation with both thermal and haline stratification changes (Neetu et al., 2012).

Another factor that can influence the TC intensity is the TC translational speed. Slower TCs, for instance, contribute to enhance the effect of air-sea coupling as they induce larger cooling and also feel the effect of this cooling longer. TCs translation speeds are very similar in the forced and coupled runs and cannot explain differences between the two. The premonsoon/postmonsoon differences in TC translation speed are also small in both model experiments and observations and are hence unlikely to play any strong role.



The robustness of our results should be assessed in other coupled models. While the overestimated TC count in our model strengthens the statistical significance of our results, it also suggests that the cumulus parameterization we use (Kain-Fritsch; Kain, 2004) may be overly active and/or too easily triggered by the environmental forcing when the atmospheric column is destabilized, as already discussed in Samson et al. (2014). We tried to assess the robustness of our results by analyzing similar twin experiments using the Bett-Miller-Janjic convective parameterization (Janjić, 1994). This parameterization leads to a more realistic number of TCs in the BoB (Samson et al., 2014) but unfortunately fails to simulate TCs stronger than Cat-1, preventing a meaningful assessment of the impact of air-sea coupling on TCs intensity. However, the fact that Lengaigne et al. (2018) showed that these two parameterizations yield to qualitatively similar results about the air-sea coupling influence on TCs characteristics, including a reduction in the TC count, is reassuring. Air-sea coupling indeed reduces the number of TCs more with the Bett-Miller-Janjic parameterization (65%) than with the Kain-Fritsch scheme (20%; Lengaigne et al., 2018). This suggests an even stronger sensitivity of TCs to air-sea coupling with a different convective scheme to the one used here.

Finally, irrespective of the convective parameterization, our one fourth degree atmospheric model resolution is not sufficient to simulate the strongest observed TCs as this resolution smooths out the sharp eyewall structure of tangential winds (Gentry & Lackmann, 2010). Numerous studies using similar horizontal resolutions in their atmospheric models reported similar issues (e.g., Jullien et al., 2014; Ogata et al., 2015). This is problematic, as our results point to a far larger impact of air-sea coupling on the strongest simulated TCs in the BoB (above 35 m/s maximum winds, Figures 4c and 4f) and because these TCs have the most devastating impacts. A number of studies reported that a 10 km or finer resolution is needed to allow simulating the most intense TCs (Jung et al., 2012; Manganello et al., 2012; Murakami et al., 2015; Satoh et al., 2012). This study will have to be pursued at such finer resolution. It is for instance likely that the effect of air-sea coupling will be larger at higher resolution. It is also possible that air-sea coupling also contributes to the characteristics of the most intense postmonsoon TCs, as those cyclones are probably strong enough to break through the haline stratification and produce a significant cooling.

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