What controls the interannual variation of tropical cyclone genesis frequency over Bay of Bengal in the post-monsoon peak season?

Zhi Li,1 Tim Li,2,3* Weidong Yu,1 Kuiping Li1 and Yanliang Liu1

1Center for Ocean and Climate Research, First Institute of Oceanography, SOA, Qingdao, China
2IPRC and Department of Meteorology, University of Hawaii, Honolulu, HI, USA
3International Laboratory on Climate and Environment Change and Key Laboratory of Meteorological Disaster, Nanjing University of Information Science and Technology, China

Abstract

Tropical cyclone (TC) over Bay of Bengal (BoB) during its climatologic maximum peak season (October–November, post-monsoon season) exhibits a significant interannual variation between a negative Indian Ocean dipole (NIOD) and a positive IOD (PIOD) phase but not between El Niño and La Niña phase. Diagnosis of observational data reveals that the most important parameter that determines the interannual variation of BoB TC is the interaction between the mid-tropospheric relative humidity and the long-term mean states of absolute vorticity, vertical wind shear, and potential intensity. The change of mid-tropospheric moisture is primarily determined by vertical advection associated with low-level vorticity anomalies during IOD.

Keywords: tropical cyclone; Bay of Bengal; IOD; interannual variation

1. Introduction

The North Indian Ocean (NIO) is one of the main tropical cyclone (TC) genesis ocean basins. TCs over the NIO impose threats to a billion people each year in Arabia, India, Myanmar, Bangladesh, and Southeast Asian region (SEAR) (Webster, 2008; Lin et al., 2013). Among the NIO TCs, almost 3/4 of those occur in Bay of Bengal (BoB). The climatological annual cycle of TC over BoB exhibits a marked bimodal character (Camargo et al., 2009; Yanase et al., 2012). The two peaks of BoB TC annual cycle take place in before and post-monsoon periods, April–May and October–November, respectively. The dynamic and thermodynamic factors that cause the bimodal characteristic were studied by Li et al. (2013b) based on the diagnosis of environmental parameters associated with the TC genesis potential index (GPI, Emanuel and Nolan, 2004). It was noted that due to the effect of mid-tropospheric relative humidity (RH), the overall TC genesis number during the post-monsoon peak is greater than that during the pre-monsoon peak (Li et al., 2013b).

Besides the distinctive annual cycle, the BoB TC also undergoes a significant interannual variation. As dominant interannual modes in the tropical oceans, the Indian Ocean Dipole (IOD) and El Niño Southern Oscillation (ENSO) may exert a great influence on the frequency, intensity, and track of TCs over the Indian Ocean through induced large-scale atmosphere circulation (Wang and Chan, 2002; William and Young, 2007; Wing et al., 2007; Eric and Chan, 2012; Clifford et al., 2013; Sumesh and Kumar, 2013). Unlike ENSO whose peak phase typically occurs in boreal winter, the peak phase of IOD occurs in October–November and it almost overlaps the maximum peak of BoB TC frequency.

BoB TC frequency exhibits significant difference during October–November between positive IOD (PIOD) and negative IOD (NIOD) rather than between El Niño and La Niña. It may be attributed to the phase-locking relationship between IOD and BoB TC maximum peak phase. IOD, a tightly coupled air–sea interaction phenomenon, may exert a great impact on the interannual variation of October–November BoB TC (Singh et al., 2008; Yuan and Cao, 2013). It was shown that more frequent BoB TC genesis in October–November occurred when there was a NIOD, and vice versa. While these observational studies pointed out the close relationship between IOD and BoB TC, it is not clear what large-scale controlling parameters associated with IOD influence the BoB TC frequency in the post-monsoon peak season. Yuan and Cao (2013) speculated that local warm sea surface temperature anomaly (SSTA) and low-level cyclonic vorticity associated with IOD might affect TC genesis. In this study we will use a quantitative analysis approach to reveal the relative roles of various environmental factors such as vertical wind shear (VWS), RH, vorticity, and sea surface temperature (SST) in controlling the interannual variation of TC frequency over BoB. Because the GPI is widely used in TC community and it well reflects to the characters of TC frequency at each ocean basin (Emanuel and Nolan, 2004; Camargo et al., 2009; Yanase et al.,

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2. Data and methods

TC best-track data from the Joint Typhoon Warning Center (JTWC) is utilized for determining TC number. Daily and monthly wind, air temperature, air-specific humidity, relative humidity from National Centers for Environmental Prediction (NCEP)—National Center for Atmospheric Research (NCAR) reanalysis and monthly SST from National Oceanic and Atmospheric Administration (NOAA) observed interpolated (OI) datasets are used to describe the large-scale environmental processes. Except for the SST data that have a horizontal resolution of 2° latitude by 2° longitude, the rest data sets, including wind, specific humidity, relative humidity, and air temperature, have a resolution of 2.5° latitude by 2.5° longitude.

It is well-known that TC genesis depends on several environmental factors including vorticity, VWS, SST, and water vapor content (Gray, 1968, 1979). Emanuel and Nolan (2004) further refined the TC genesis condition and proposed a GPI:

\[
\text{GPI} = \text{Term1} \times \text{Term2} \times \text{Term3} \times \text{Term4}
\]

where, \( \text{Term1} = 105 \eta^{3/2} \), \( \text{Term2} = (1 + 0.1 V_{\text{shear}})^{-2} \), \( \text{Term3} = \left( \frac{H}{g} \right)^3 \), \( \text{Term4} = \left( \frac{V_{\text{shear}}}{V_{\text{pot}}} \right)^3 \), \( \eta \) is absolute vorticity at 850 hPa (hereafter, vorticity), \( V_{\text{shear}} \) is the magnitude of VWS between 200 and 850 hPa, \( H \) is RH at 600 hPa , and \( V_{\text{pot}} \) is the maximum TC potential intensity (PI) defined by Emanuel and Nolan (2004):

\[
V_{\text{pot}}^2 = C_p \left( T_s - T_o \right) \frac{T_s}{T_o} C_k \left( \ln \theta_e^* - \ln \theta_e \right)
\]

In the PI formula, \( C_p \) is the heat capacity at constant pressure, \( T_s \) is the ocean temperature, \( T_o \) means outflow temperature, \( C_k \) is the exchange coefficient for enthalpy, \( C_D \) is the drag coefficient, \( \theta_e^* \) is the saturation equivalent potential temperature at ocean surface, and \( \theta_e \) is the boundary layer equivalent potential temperature. In order to quantitatively assess the relative contribution of

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Each term, Li et al. (2013b) developed an analysis strategy by taking a natural logarithm at both sides of the GPI formula first and then applying a total differential to both sides. Thus the change of GPI may be separated into four terms as below:

$$\delta \text{GPI} = \alpha_1 \delta \text{Term1} + \alpha_2 \delta \text{Term2} + \alpha_3 \delta \text{Term3} + \alpha_4 \delta \text{Term4}$$  \hspace{1cm} (3)

where

$$\alpha_1 = \begin{cases} \frac{\text{Term2} \: \text{Term3} \: \text{Term4}}{\text{Term2} \: \text{Term3} \: \text{Term4}} & \text{method 1} \\ \frac{\text{Term2} \: \text{Term3} \: \text{Term4}}{\text{Term2} \: \text{Term3} \: \text{Term4}} & \text{method 2} \end{cases}$$  \hspace{1cm} (4)

$$\alpha_2 = \begin{cases} \frac{\text{Term1} \: \text{Term3} \: \text{Term4}}{\text{Term1} \: \text{Term3} \: \text{Term4}} & \text{method 1} \\ \frac{\text{Term1} \: \text{Term3} \: \text{Term4}}{\text{Term1} \: \text{Term3} \: \text{Term4}} & \text{method 2} \end{cases}$$  \hspace{1cm} (5)

$$\alpha_3 = \begin{cases} \frac{\text{Term1} \: \text{Term3} \: \text{Term4}}{\text{Term1} \: \text{Term3} \: \text{Term4}} & \text{method 1} \\ \frac{\text{Term1} \: \text{Term3} \: \text{Term4}}{\text{Term1} \: \text{Term3} \: \text{Term4}} & \text{method 2} \end{cases}$$  \hspace{1cm} (6)

$$\alpha_4 = \begin{cases} \frac{\text{Term1} \: \text{Term2} \: \text{Term3}}{\text{Term1} \: \text{Term2} \: \text{Term3}} & \text{method 1} \\ \frac{\text{Term1} \: \text{Term2} \: \text{Term3}}{\text{Term1} \: \text{Term2} \: \text{Term3}} & \text{method 2} \end{cases}$$  \hspace{1cm} (7)

Through the above diagnosis method, one may quantitatively estimate the impacts of each environmental factor on TC frequency change between the negative and positive IOD phases.

A moisture budget analysis is further conducted to examine specific processes that give rise to the specific humidity anomaly ($q'$) in PIOD and NIOD. Following Yanai et al. (1973), Hsu and Li (2012), and Li et al. (2013a), a tendency equation for interannual specific humidity anomaly ($q'$) may be written as:

$$\frac{\partial q'}{\partial t} = -(V \nabla q)' - \left( \omega \frac{\partial q'}{\partial p} \right)' - (Q_2/L)'$$  \hspace{1cm} (8)

where $V$ is the horizontal velocity, $\omega$ is the p-vertical velocity, $Q_2$ is apparent moisture sink, and $L$ is latent heat constant. In the equation above, $-(V \nabla q)'$ denotes anomalous horizontal moisture advection, $-\left( \omega \frac{\partial q'}{\partial p} \right)'$ indicates anomalous vertical moisture advection, and $-(Q_2/L)'$ represents the anomalous moisture source or sink ($Q_2$ is primarily determined by surface evaporation and atmospheric condensation).

3. Analysis results

Six PIOD cases (1982, 1986, 1994, 1997, 2006, and 2007) and six NIOD cases (1984, 1992, 1996, 1998, 2005, and 2010) are selected during the period of 1981–2010 according to the original definition of IOD (Saji et al., 1999; Webster et al., 1999). In the same time, 1982, 1986, 1991, 1994, 1997, 2002, 2006, and 2009 are confirmed as eight El Niño cases and 1988, 1999, 2007, and 2010 also are considered as La Niña cases. Figure 1(a) presents BoB TC frequency number is significantly different with its climatology in October–November of IOD and the confidence level is above 90%. On the other hand, the difference of TC number exceeds 90% confidence level only in October of El Niño (Figure 1(b)). Furthermore, the difference of TC number between NIOD and PIOD exceeds 95% confidence level, respectively in October and November (Figure 1(c)). However, the difference significance between El Niño and La Niña is less than 75% (Figure 1(d)). In October–November, the difference of TC number between NIOD and PIOD is statistically significant at a confidence level exceeding 99%. Hence, it may be concluded that IOD mostly modulates the
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The paper examines the spatial distribution of GPI and TC genesis locations during October–November of both IOD phases over BoB, respectively. The horizontal distribution of the GPI anomaly well fit the TCs’ feature (figure not shown). Moreover, according to BoB TC genesis region, a box (5°–20°N, 80°–100°E) is defined as the TC formation area. We calculated the area averaged GPI values during October–November of NIOD and PIOI and found that these values match quite well the averaged October–November TC number (Figure 2(a)).

Because the GPI well reflects TC genesis frequency difference between NIOD and PIOI, we further use GPI to diagnose the relative contribution of each environmental factor. Figure 2(b) shows the calculation result (i.e. difference between NIOD and PIOI). As one can see, more frequent TC frequency during NIOD is primarily attributed to the interaction between the mid-tropospheric RH and the long-term mean states of absolute vorticity, VWS, and potential intensity. Because mean states of absolute vorticity, VWS, and potential intensity are constant and independent of year-to-year changes, we may approximately state the contribution is attributed to RH term. In the same way, the vorticity term has a weak positive contribution. The VWS term has a negative contribution. The effect of SST or PI is negligible.

The result indicates that IOD primarily modulates BoB TC frequency through the change of RH field in middle troposphere (600 hPa). Given that RH is a function of specific humidity (SH) and air temperature (AT), a natural question is whether RH change is primarily controlled by SH or AT changes. By the diagnosis, we confirmed the change of the 600 hPa RH associated with IOD is primarily attributed to the change of SH, not AT.

The examination of time evolution of 600 hPa SH shows that it has a maximum tendency in September–October and reaches its peak in October–November for both the PIOI and NIOD composites (Figure 3). Therefore, a question of why GPI has a maximum positive (minimum negative) value in October–November during NIOD (PIOI) may be converted to a question of why local SH at 600 hPa increases (decreases) rapidly during September–October. This motivates us to further investigate the change of 600 hPa SH.
Figure 5. (Top) Vertical profiles of area-average anomalous vertical (a) p-velocity ($\omega$), (b) vorticity and (c) SH over (5°–20°N, 80°–100°E) in September–October for NIOD (dashed) and PIOD (solid). (Bottom) Composite 850 hPa wind anomaly (vector) and SST anomaly (shading) in September–October during (d) PIOD and (e) NIOD.

Figure 4 shows the SH budget analysis result. Because the PIOD and NIOD results are approximately a mirror image, we discuss in the following only the NIOD result (Figure 4(b)). The increase of SH at 600 hPa is primarily attributed to the vertical advection, while the apparent moisture sink term has a negative contribution. Physical interpretation is given below. Note that SST anomalies associated with NIOD are quite strong in September–October. Although the SSTA is asymmetric about the Equator in the eastern IO during IOD events and its center generally occurs at nearby 8°S in the vicinity of Sumatra, the wind curl anomaly responding to the SSTA is quite symmetric to the Equator (Li et al., 2003; Yu et al., 2005; Schott et al., 2009). In response to a warm (cold) SSTA in the eastern (western) equatorial IO during NIOD, there are pronounced westerly anomalies at the equator and two cyclonic gyre circulations at both sides of the equator (Figure 5(e)). The cyclonic flow at top of atmospheric boundary layer induces anomalous ascending motion through the Ekman pumping effect. The anomalous ascending motion advects the mean moisture upward.
and causes the increase of SH in the lower and middle troposphere (Figure 5(c)). This promotes a convectively unstable stratification and favors the deepening of ascending motion and moist layer (Figure 5(a) and (b)). The increase of mid-tropospheric moisture favors more frequent TC genesis in October–November during NIOD.

In addition to the moisture effect, the low-level cyclonic (anticyclonic) wind anomaly at 850 hPa also contributes to TC genesis frequency change during NIOD (PIOD). The cyclonic and anticyclonic wind anomaly makes the vorticity term to provide secondary positive contribution for TC number difference between NIOD and PIOD.

Because climatologic wind vertical shear in BoB is easterly shear, an increase (decrease) of such a shear in the BoB TC genesis region during NIOD (PIOD) prohibits (favor) cyclogenesis. This is why VWS has a negative impact on BoB TC on the interannual timescale. On the other hand, according to previous studies (e.g. Li, 2006; Li, 2012), compared with same amount of westerly shear, easterly vertical shear is more favorable for the growth of synoptic-scale perturbations. From this point, NIOD may provide a favorable environmental shear condition for TC formation. SST plays a key role in PI term. The effect of the PI on the GPI during October–November of IOD is very weak since BoB region averaged SST is similar to its climatology during this period.

4. Summary

Through quantitative diagnosis of various environmental parameters associated with the GPI, we found that the most important parameter that determines the interannual variation of TC genesis frequency in BoB during the post-monsoon season is interaction between the mid-tropospheric RH and long-term mean states of absolute vorticity, VWS, and PI. The contribution is attributed to RH term because the mean states of the rest terms are constant and independent of interannual variation. In the same way, for BoB TC-number difference in October–November between NIOD and PIOD, we can state low-level environmental vorticity has a weaker positive contribution, VWS has a negative contribution, and PI’s contribution is almost negligible.

Due to the critical role of mid-tropospheric water vapor, a moisture budget analysis is further carried out to understand key processor responsible for the mid-tropospheric moisture change. It was found that the major process affecting the moisture change is anomalous vertical advection.

Based on the results above, a scenario through which a negative or positive IOD influences BoB TC genesis frequency in October–November is proposed. Because NIOD and PIOD scenarios are generally a mirror image, in the following we focus on the discussion of the NIOD case.

1. Although the SSTA is asymmetric about the equator, its impact on atmospheric circulation in the eastern IO is approximately symmetric. In response to a NIOD, there are pronounced low-level westerly anomalies at the equator and an anomalous cyclone over BoB. The cyclonic flow at top of boundary layer induces anomalous ascending motion, which transports mean moisture upward and moistens the lower troposphere. The so-induced convective instability further strengthens large-scale ascending motion. As a result, the moist layer deepens. Our moisture budget analysis confirms that the anomalous vertical advection, instead of anomalous horizontal advection, plays a critical role in increasing the specific humidity anomaly at 600 hPa. The change of SH mainly contributes to the change of RH in 600 hPa.

2. Meanwhile the low-level cyclonic vorticity anomaly at 850 hPa makes a weaker positive contribution to the GPI difference between NIOD and PIOD.

3. The VWS associated with IOD has a negative contribution to the GPI change. This is because the wind anomaly associated with NIOD has nearly same direction as the climatological wind over BoB TC genesis region.

4. The PI term does not significantly contribute to the GPI change, simply because the SSTA associated with IOD over BoB is very weak and the most significant SSTA appears south of the equator.

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