Tropical Cyclogenesis Associated with Premonsoon Climatological Dryline over the Bay of Bengal

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Research Article

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Abstract

Tropical cyclones of the Bay of Bengal (BoB) that formed near the synoptic-scale dryline usually intensified over a short distance ($\sim 600\text{--}800$ km) within 3 days and caused severe destruction after landfall. High-resolution simulations of very severe cyclonic storms in association with dryline indicate that the meridional shear aids in the development of a group of linear convective cells that mature as an east-west oriented quasi-linear convective system (QLCS) within the boundary between the dry-moist air masses. The leading edge deep convections are supported by low-level moist southwesterly inflow; however, the typical mid-level mesoscale convective vortex (MCV) associated with these QLCS is unremarkable due to a very narrow trailing stratiform region within the QLCS. Supercells are likely to be organized within the QLCS due to extremely unstable atmospheric conditions resulting from a strong vertical shear of $27\text{--}39\text{ m s}^{-1}$ between 0–6 km and large convective available potential energy of $>3000\text{ J kg}^{-1}$. The vertical shear veering with height causes several numbers of low-level mesovortices having diameters less than 10 km at the leading edge in the different convective stages of the QLCS. The dryline aloft in the BoB produces horizontal positive shear vorticity of the order 10^{-5} s^{-1} with higher values in the levels 850–600 hPa. The advection of intense cloud-scale cyclonic vortices ($\sim 10^{-3}\text{ s}^{-1}$) assists and enhances a cyclonic vortex to the rear side of the QLCS that performs as an MCV for cyclogenesis over the BoB.

1. Introduction

The genesis of a tropical cyclone (TC) follows a hierarchical environment. The appropriate climatology and proper synoptic conditions primarily favor initiating tropical disturbances or cloud clusters over oceans (Gray, 1998). Within these tropical disturbances, the triggering of mesoscale convective systems (MCSs) are the fundamental precursors of cyclogenesis in any ocean basin (e.g., Zehr 1992; Harr et al. 1996; Bister and Emanuel 1997; Ritchie and Holland 1997; Simpson et al. 1997; Gray 1998; Houze 2010). MCSs are organized cloud clusters containing both deep convective cells and stratiform clouds and precipitation (Houze 2004). Once, MCSs were usually classified in terms of either the organizational modes (Bluestein and Jain 1985) or the location of the stratiform precipitation (Parker and Johnson 2000). More recently, Gallus et al. (2008) divided MCSs into nine morphologies for warm-season convection: nonlinear systems, cellular convection (three types; isolated, clusters, and broken lines), and linear systems (five types; bow echoes, squall lines with trailing/leading/parallel/no stratiform rain). However, only a few studies (Bister and Emanuel 1997; Powers and Davis 2002; Akter 2015) have examined the types of MCSs that occur during the formation of a tropical cyclone (TC) that typically develops in warm summer seasons.

Many studies have observed that a mature or decaying MCS comprises a warm-core cyclonic circulation called mesoscale convective vortex (MCV) that develops in the stratiform precipitation, which is typically 50–300 km in diameter and several kilometers deep (e.g., Bartels and Maddox 1991; Davis and Weisman 1994; Davis et al. 2004; Davis and Trier 2007; James and Johnson 2010). An MCV may persist long after

the dissipation of the parent convective system and keep support new storms for several days (e.g., Bosart and Sanders 1981; Bartels and Maddox 1991; Fritsch et al. 1994; Trier and Davis 2007). According to the AMS glossary, an MCV is formed by the latent heat release in the stratiform region of the MCS (American Meteorological Society 2019). The lower-tropospheric evaporative cooling and stratiform condensational heating below the upper troposphere allow the growth of a mid-tropospheric potential vorticity anomaly within the squall line MCS (Hertenstein and Schubert 1991). However, different studies have provided different explanations for the mechanism of the individual MCV formation. The generally accepted processes include: vortex-stretching mechanism by the mesoscale vertical motion (Bartels and Maddox 1991); tilting of the ambient or baroclinically generated horizontal vorticity resulting in vertical vorticity and enhancement of the mid-tropospheric convergence due to rotation of the earth, especially, the cyclonic vortex is dominated on the northern end of the asymmetric squall line MCSs (Davis and Weisman 1994; Skamarock et al. 1994); and reduction of the local Rossby radius that induces mid to low-level convergence through localized warming by organized convection (Chen and Frank 1993). As a whole, the fundamental formation of an MCV depends on four atmospheric regimes: convergence-dominated, heating gradient-dominated, heating-dominated, and vertical advection-dominated regions working from the bottom to the top of the troposphere, and a variation in these atmospheric regimes affects the origin and intensification of the MCV in different developmental paths (Kirk 2003; 2007).

The MCVs associated with the MCSs that form over the tropical oceans serve as a nucleus for a TC and are correspondingly the essential precursors to its genesis (e.g., Zehr 1992; Harr et al. 1996; Bister and Emanuel 1997; Ritchie and Holland 1997; Simpson et al. 1997; Reasor et al. 2005; Houze 2010). The MCVs contribute to the development of the surface vortex for TCs either due to a downward vortex penetration through the mesoscale subsidence in association with the stratiform precipitation (the top-down process) (Emanuel 1993; Bister and Emanuel 1997; Ritchie and Holland 1997) or a convective tower-driven development known as a bottom-up process (Montgomery and Enagonio 1998; Hendricks et al. 2004; Reasor et al. 2005; Montgomery et al. 2006; Braun et al. 2010). In the latter case, aggregation of low-level vortices associated with rotating deep convection (vortical hot towers; VHT) intensifies the low-level cyclonic vortex within the embryonic mid-level vortex via a wave pouch (Dunkerton et al. 2009). According to the marsupial theory, a wave pouch is formed before the TC genesis, providing it a “sweet spot”, which is also supported by a recent study (Gjorgjievska and Raymond 2014).

The previous study by Akter (2015) revealed that MCSs, which are connected with cyclogenesis over the Bay of Bengal (BoB, a sub-region of the North Indian Ocean), portray a specific nature depending on the seasonal variability. The large-scale dynamics and thermodynamics conditions over the BoB permit bimodal cyclone seasons: before and after the monsoon season (June-September) (Akter and Tsuboki 2014). Due to the season location of the monsoon trough, the temporal pattern of TC formation over the BoB is unique compared with the other ocean basins where only the summer season is favorable for cyclone development (McBride 1995; Harr and Chan 2005; Camargo et al. 2007; Kikuchi and Wang 2010; Akter and Tsuboki 2014). The advection of seasonal winds toward the BoB favors the formation of distinctive MCSs in terms of structure, movement, and severity during cyclogenesis (Akter 2015). The development of a severe bow-echo-type MCS in an environment of high CAPE and strong clockwise shear

was reported in connection with the cyclone Akash (2007), which was formed during premonsoon (March-May) in the presence of a dry-moist coupled environment i.e. dryline in the BoB (Akter and Tsuboki 2017). According to the Joint Typhoon Warning Center (JTWC) best track data, cyclone Akash (2007) changed from a depression-level (Maximum surface wind speed, MSW < 25 kt) to a very severe cyclonic storm (MSW of 65 kt) in just one and a half days. This changeover within a very short period is unusual and a threat to the low-lying coastal countries like Bangladesh and India. To understand the fast and intense growth of TCs that form along the dryline in the BoB, it is essential to identify the behavior of MCSs for more cases and to study the associated MCVs related to the TC formation.

To the best of our knowledge, no studies have been conducted before for the climatology of cyclone formation associated with dryline that is critical for severe weather formation. Therefore, the main objective of the present study is to find out the favorable conditions that allow the MCS formation and to analyze the MCV development process responsible for the dryline-related cyclogenesis over the BoB. Due to a dearth of observed data, a high-resolution simulation is the only conceivable method for investigating the detailed structure and mechanism of the MCSs and MCVs during the formation of TCs in the BoB.

2. Synoptic-scale Dryline And Premonsoon Cyclones

A large-scale prominent dryline, the boundary between dry and moist air masses, occurs during premonsoon season (March-May) along the east coast of India and adjacent BoB for 1990 to 2009 (Akter and Tsuboki 2017). The dryline has a horizontal length of ~ 2000 km with an intense dew point gradient of 1°C per 10 km at the surface. Severe local convective storms (Nor'wester) and tornadoes over the land are most likely near the premonsoon dryline, which causes destruction, damage and a large number of deaths each year (Ono 1997; Yamane et al. 2009; Bikos et al. 2016). Moreover, the vertical structure of the dryline shows that a dry-moist air convergence zone aloft sloping from the east at a level of approximately 700 hPa over the BoB to the west at the surface of the eastern coast of India. As a result, the northern and northwestern BoB contains dry air, with the moist air kept in the southern and southeastern region of the BoB (Akter and Tsuboki 2017). In this study, the zero anomalies of specific humidity is displayed in Fig. 1a which is averaged from the surface to 700 hPa during the premonsoon from 2001–2020 for indicating the synoptic-scale dryline over the BoB. The vertical cross-section in Fig. 1b confirms that the dry (less than 6 g kg^{-1}) and hot (more than 26°C) air is coming from the north and northwest, whereas the moist (more than 14 g kg^{-1}) and warm (less than 20°C) air is entering from the southwest to the BoB. The north and northwesterly dry air masses are very deep and are stretched from the surface to 500 hPa; on the other hand, moist southwesterlies are shallow, i.e., up to 700 hPa pressure level from the surface. The dry-moist boundary forms a lifted and inclined dryline over the BoB.

In the last 21 years (2001–2020), a total of 18 cyclones occurred during premonsoon with the maximum TC frequency in May (61%) and the minimum in March (6%). The frequencies of premonsoon TCs is almost half that in the postmonsoon season (October–December) due to the presence of a stable layer over the northern BoB; this inhibits the development of TCs even though the monsoon trough is located in

the northernmost region of the Bay (Akter and Tsuboki 2014). The larger values ($> 50 \text{ J kg}^{-1}$) of convective inhibition (CIN) shown in Fig. 1a support the stable environment aloft in the northern and northwestern BoB due to deep hot, dry air.

Within the premonsoon TCs, four TCs out of the nine very severe TCs ($\text{MSW} \geq 65 \text{ kt}$) i.e., BoB_02 (2004), Akash (2007), Aila (2009), and Mora (2017) developed around the strong gradient where the dry-moist air masses congregated. Akash (2007) was simulated before by the author to examine the formation and characteristics of the MCS during its genesis period (Akter, 2015). Subsequently, the remaining three TCs have been simulated in this study to analyze the details of the MCS and identify the causes for MCVs (core vortex of TCs) in the presence of a dryline. All these three TCs have common basic criteria, i.e., all formed in May and appeared as “very severe cyclonic storms” of the IMD cyclone intensity scale. BoB_02 (2004) and Aila (2009) had MSWs of 65 kt, but Mora (2017) had attained a relatively higher MSW (80 kt) according to JTWC. The corresponding minimum surface pressures were 976 hPa, 974 hPa and 963 hPa for BoB_02 (2004), Aila (2009) and Mora (2017), respectively. They reached the maximum intensity within a very short distance ($\sim 600\text{--}800 \text{ km}$) from their depression locations within ~ 3 days. The tracks of these cyclones are shown in Fig. 1a. Brief information of these TCs is listed as follows:

- BoB_02 (2004): a depression formed at 14.8°N , 89.7°E at 1800 UTC 14 May 2004 and developed as a cyclonic storm (MSW of 34 kt) at 18.0°N , 89.8°E at 0000 UTC 17 May 2004 according to the JTWC data. At 0600 UTC 19 May 2004 the TC made landfall in northwestern Myanmar as a very severe cyclonic storm. The storm caused significant loss of life (236 deaths) and extensive damage to the infrastructure; approximately 25,000 people were affected by the storm surges and flooding (Myanmar Red Cross Society 2004).
- Aila (2009): JTWC reported the pre-Aila formation as a depression located at 16.9°N , 88.8°E at 1800 UTC 22 May 2009. It developed into a cyclonic storm at 17.7°N , 88.5°E at 0000 UTC 24 May 2009 and made landfall on the southwestern coast of Bangladesh and eastern West Bengal, India at 0600 UTC 25 May 2009 as a very severe cyclonic storm. The cyclone caused deadly storm surges of 3–4 m along the western coastlines of Bangladesh and the immediate death of about 325 people (Roy et al. 2009). It affected 3.9 million people and caused massive infrastructure damage; for example, it washed away nearly 350,000 acres of cropland (UN 2010).
- Mora (2017): According to the JTWC, a depression was located at 13.9°N , 87.4°E at 0600 UTC 27 May 2017. It strengthened into a cyclonic storm at 13.9°N , 89.2°E at 1800 UTC 27 May 2017 and made landfall at 0000 UTC 30 May 2017 over the southern coast of Bangladesh as a very severe cyclonic storm. High winds, heavy rain, and storm surges triggered severe floods and landslides, killing 7 people immediately and affecting 3.3 million (Bangladesh Red Crescent Society 2017).

3. Model Configurations And Data Used

Three premonsoon cyclones BoB_02 (2004), Aila (2009), and Mora (2017) that formed over the BoB (Fig. 1a), were selected for the simulations using the Advanced Hurricane Weather Research and

Forecasting (WRF) Model (AHW), version 3.7.1 (Davis et al. 2008; Skamarock et al. 2008). The simulations were performed during the genesis of each TC to improve the understanding of the formation process and the characteristics of the MCSs and MCV connected with cyclogenesis. To manage the surface fluxes within the TCs properly, a one-dimensional mixed-layer ocean model based on Pollard et al. (1973), surface drag coefficients described by Donelan et al. (2004), and enthalpy exchange coefficients described by Dudhia et al. (2008) were also selected for the AHW model. All the configurations in the simulation were preferentially retained from the previous study (Akter 2015). The selected physical parameters in this previous study were followed according to Raju et al. (2011) and Krishna et al. (2012), who adapted all the parameters for simulating cyclones over the BoB using the WRF model.

A two-way nested simulation was facilitated with a horizontal resolution of 12 km and 4 km, respectively, for the outer and inner domains. The outer domain (D1) with 861×595 grid points covers the area 0° – 30° N and 70° – 110° E (not shown here but same as Fig. 1 in Akter 2015); however, the inner domain (D2) with 1064×996 grid points includes 5° – 22° N and 80° – 97° E (area of Fig. 2). The vertical resolution was defined as 30 terrain-following levels with a top pressure level of 50 hPa. The six-hourly National Center for Environmental Prediction Final analysis (NCEP-FNL) data (<https://rda.ucar.edu/datasets/ds083.2/>) with a resolution of $1^{\circ} \times 1^{\circ}$ were applied as the atmospheric initial and boundary conditions. In addition, the sea surface temperatures were implemented daily using the NCEP high-resolution (0.083°) real-time global sea surface temperatures (RTG_SSTs) data (<https://polar.ncep.noaa.gov/sst/>); moreover, no bogussing was imposed for the TC initialization. In the model set-up, the map projection, convective parameterization, microphysics, planetary boundary layer (PBL), surface physics, land-surface model, and longwave and shortwave radiation processes were considered according to the Lambert projection, Kain–Fritsch scheme (Kain, 2004; only for the parent domain), Ferrier scheme (Ferrier et al. 2002), Yonsei University scheme (YSU, Hong et al. 2006), Monin–Obukhov scheme (Monin and Obukhov 1954), Noah land-surface model (Chen and Dudhia 2001), rapid radiative transfer model (Mlawer et al. 1997), and Dudhia scheme (Dudhia 1989), respectively. A brief description of all the processes has been given in Akter (2015).

The intensity and position data for the TCs were collected from the six-hourly best track data of the JTWC (<https://www.metoc.navy.mil/jtwc/jtwc.html?north-indian-ocean>). According to these data records, each TC needed less than two days to intensify from a tropical depression into a cyclonic storm, which is a system regulated by a self-sustaining mechanism with an MSW of at least 34 kt or $\sim 17.5 \text{ m s}^{-1}$ (Zehr 1992; McBride 1995). Therefore, four days excluding the 6-hour model spin-up time were allocated as the simulation period in each case: three for the period before the development of the TC ($\text{MSW} < 34 \text{ kt}$) and the remaining day for the time after the formation ($\text{MSW} \geq 34 \text{ kt}$). The TC simulations were assessed from 0000 UTC 14 May to 0000 UTC 18 May 2004 for BoB_02; 0000 UTC 21 May to 0000 UTC 25 May 2009 for Aila; and 0000 UTC 25 May to 0000 UTC 29 May 2017 for Mora. The model outputs were stored every 30 min.

The NCEP climate forecast system reanalysis (CFSR) six-hourly data (<https://rda.ucar.edu/datasets/ds094.0/>) with a horizontal resolution of $0.5^{\circ} \times 0.5^{\circ}$ were used to verify

the model outputs; its monthly mean data were also utilized to analyze the seasonal variability of the environment over the BoB. The data provide the vertical resolution of 64 hybrid sigma-pressure levels, with a top pressure of ~ 0.266 hPa (Saha et al. 2010). Though JTWC best track data was available and used in this study, however, the intensity scale of TCs was mentioned according to Indian Meteorological Department (IMD).

4. Result

4.1. Model verification

The simulated results at the resolution of 4 km were verified with the NCEP-CFSR six-hourly reanalysis data given in Fig. 2. The shaded area represents the surface convective available potential energy (CAPE, Glickman 2000), which is averaged from the depression-to-genesis period of each TC. The total period from depression-to-genesis varies for each TC mentioned in the previous section. In this research work, the primary objective was to understand the MCSs associated with pre-TCs. Thus, the distribution of the pre-cyclone environmental CAPE was verified as it is directly related to the initiation of deep convection (e.g., Williams and Renno 1993; Bhat et al. 1996; May and Rajopadhyaya 1999; Rapp et al. 2011), including that in the TC environment (Molinari and Vollaro 2009; Molinari et al. 2012; Lee and Frisius 2018). The simulated CAPE in all cases was found to be almost similar in distribution to that in the reanalysis data; the larger values of CAPE in the north and northwestern areas of the BoB are especially well projected. However, the distributed CAPE has larger values in model outputs due to higher resolution. The low-pressure center and corresponding rotation of the wind were achieved satisfactorily in the simulation result. The intensity and position of each simulated cyclone were compared with the JTWC data at its genesis. In the simulation, the minimum pressures of BoB_02 (2004) and Aila (2009) were 998 and 994 hPa, respectively, whereas the corresponding intensities reported by JTWC were 997 and 993 hPa, which are almost identical. The simulation of Mora (2017) was less intense (5 hPa greater than the observed center pressure of 996 hPa). At the genesis time, the positions of the TCs were close to the observed positions. The maximum errors in longitudinal position were found 0.5° , 1.6° and 0.8° for BoB_02 (2004), Aila (2009) and Mora (2017), respectively.

In this study, the simulated rainfall for Aila (2009) is compared in Fig. 3 with the available rainfall data collected from Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) 2A12 version-6 because Aila (2009) has been represented as a case study to understand the formation of MCV in depth. A strong northern rainband of Aila (2009) is observed approaching the Bangladesh coast in both cases; however, the time achieved in the simulation is 12 hours later than the actual time at 2118 UTC 23 May 2009.

4.2. Pre-cyclone environment and convection initiation

The reflectivity in Figs. 4a–c shows the early MCSs for all pre-cyclones. The initiation time of MCSs differs from cyclone to cyclone. In all cases, early MCSs are considered as a group of cells with a

maximum intensity of 35–40 dBZ. The moisture advection by a strong southwesterly wind triggers the convection along the convergence zone of dry-moist air masses encompassing minor temperature differences (~ 1 K). The temperature gradient in the BoB is not sufficient to decide upon a typical front that describes a moderate baroclinic zone (Sanders 1999) with a threshold value in temperature change being $8^{\circ}\text{C}/220\text{ km}^{-1}$. Here, the warm moist southwesterly flow having a high equivalent potential temperature θ_e of ~ 350 K (on average for all TCs), is lifted by the density current (Lin 2007) associated with low θ_e (~ 330 K) air advected from the hot dry north and northwesterly, which helps raise the low-level air parcel. The vertical cross-sections of θ_e along the low-level convergence zone at 13.5°N for pre-BoB_02 (2004), at 11°N for pre-Aila (2009) and at 12.8°N for pre-Mora (2017) are shown in Fig. 5 which indicate that the southwesterly air with θ_e of ~ 354 K are converged and elevated at the lower level by the north and northwesterly air having θ_e of ~ 340 K. The low-level warm moist air is potentially unstable with the gradual decreasing of θ_e , on the other hand, the increasing of θ_e with height supports a very stable layer for the hot dry air aloft.

The instability that aids in the convection development in each pre-cyclone period is calculated from a skew-T log-P diagram and hodograph at the location noted in Figs. 4a–c with the stars. The positions are chosen adjacent to the convection near the dryline at the zero anomalies of temperature so that the environment of the dry-moist boundary can be identified properly. The skew-T log-P diagrams are not displayed here. However, the values of different atmospheric indices for all cases are noted in Table 1, which show that the K index (George 1960) measuring the thunderstorm potential is approximately 40°C , the value of precipitable water (PW) is more than 6 cm, the lifted index (LI; Galway 1956) is greater than -6 , the total totals index (TT; Miller 1972) is $> 40^{\circ}\text{C}$, and θ_e is ~ 370 K. These values indicate very unstable air masses to be lifted to form strong or severe organized thunderstorms with great potential for heavy rain. For all cases, the vertical total wind shear from 0–6 km above ground level (AGL) was found to be $27\text{--}39\text{ m s}^{-1}$ and bulk shear between the surface and 6-km height is $12\text{--}14\text{ m s}^{-1}$; these are within or greater than the threshold value for a typical supercell formation (Bunkers 2002). CIN is found 0 J kg^{-1} , whereas, strong shear with clockwise veering and a high CAPE of more than 3000 J kg^{-1} support extreme unstable conditions conducive to supercell initiation (Weisman and Klemp 1982; Bluestein 1993). In addition, the moderate value of storm-relative environmental helicity (SREH, Davies-Jones et al. 1990) and the bulk Richardson number (BRN) of ~ 30 (Weisman and Klemp 1982) provide an indication of the development of a thunderstorm with a rotating updraft.

Table 1
The values of environmental parameters

Parameters	Pre-BoB_02 (2004)	Pre-Aila (2009)	Pre-Mora (2017)
0-6-km bulk shear (m s^{-1})	12.40	13.08	13.70
0-6-km total shear (m s^{-1})	30.12	27.19	38.54
with direction	Veer clockwise	Veer clockwise	Veer clockwise
CAPE (J kg^{-1})	3972	3421	3188
PW (cm)	6.67	6.44	6.35
K ($^{\circ}\text{C}$)	41	38	37
TT ($^{\circ}\text{C}$)	46	42	43
θ_e (K)	371	368	372
LI ($^{\circ}\text{C}$)	-7	-6	-6
0-3-km SREH ($\text{m}^2 \text{s}^{-2}$)	122	109	176
BRN	37	32	28
CIN (J kg^{-1})	0	0	0

4.3. Characteristics of mature MCSs

Over a period of several hours (6–12) from the formation, an early MCS matures as a quasi-linear convective system (QLCS) or comma shape convection with a series of embedded echoes having maximum reflectivity more than 50 dBZ, as shown in Figs. 4d-f for each case. Based on the potentiality of the environment and low-level SREH, supercell thunderstorms are likely embedded within the organized QLCS. In the figures the 500-hPa geopotential height accompanied by the wind indicates that mid-level air is rotating near or at the center of the QLCS and the low value of geopotential height indicates a mesovortex with a central low pressure.

The structural details of a pre-Aila (2009) QLCS in horizontal and vertical dimensions during the matured stage are incorporated in Fig. 6. The positions of the QLCS at 1030 UTC and at 2030 UTC 21 May 2009 are displayed with green and red hatched areas, respectively in Fig. 6a. The QLCS, oriented approximately in the east-west direction, has a horizontal length of ~ 500 km and remains quasi-stationary with a very little movement toward the southeast. The movement of the QLCS is approximately calculated by the perpendicular movement of the center of the convective line (i.e. horizontal line connecting with deep convective cells of QLCS). The rear inflow jet (RIJ, Weisman 1993) shown in the figure coming from the north and northwest assists the MCS to be a comma-shaped structure. Figure 6b shows the vertical cross-section of reflectivity, updraft, downdraft and system-relative motion along AB in Fig. 6a, which

indicates that the strong low-level southwesterly inflow converges and supplies moisture into the system and consequently deep convection containing maximum precipitation of 60 dBZ develops along the leading edge or south side of the system. The system height is extended up to the pressure level of 300 hPa where the RIJ is flowing between 850 and 500 hPa levels. Updrafts with a maximum speed of 6 m s^{-1} are observed, which are separated from the downdrafts descending at a maximum speed of 1.5 m s^{-1} . Moreover, after 10 hours the same vertical analysis along CD in Fig. 6c represents the dissipation of older convective cells that develops a narrow stratiform precipitation region extending toward the rear side and a new convective cell appears at the leading edge which also can be confirmed by the strong low-level convergence in Fig. 6d. During the mature stage, the interaction of low-level (surface to 3 km) shear of 20 m s^{-1} with a weak cold pool supports forced lifting of air, ensuring new cell development along the down shear in the organized systems of QLCS, similar to observations in previous studies (Rotunno et al. 1988; Weisman and Rotunno 2004). The potential temperature (θ) anomalies in Fig. 6d suggest that the 1.5-km-deep cold pool has the lowest temperatures of -4 K , which is sufficient for developing tropical convection (COMET Program/UCAR 1999). The characteristics of the MCSs for pre-Aila (2009) and the other two pre-cyclones (pre-BoB_02 (2004) and pre-Mora (2017)) are in agreement with the features of pre-Akash (2007) mentioned in Akter (2015), except for their movement speeds. However, most MCSs remain stationary from their early stage to mature, which can be confirmed by the position of the initially formed MCSs and their mature phases in Fig. 4. In all cases, the lifetime of MCSs are very long (~ 20 hours) and within this lifetime several mesovortices (vorticity at meso- γ -scale: 2–20 km; Orlanski 1975) are formed at low levels along the convective line of QLCSs (Trapp and Weisman 2003; Weisman and Trapp 2003; Atkins and Laurent 2009a). Figure 6a displays cloud-scale cyclonic vortices for pre-Aila (2009) with an approximate horizontal length of 10 km or less. Further additional features of the mesovortices are described in the next section.

4.4. Mesovortices in pre-Aila

4.4.1 Formation of mesovortices

The shade in Fig. 7a represents the meridional wind component, which indicates that when the wind coming from the south initially intrudes into the northerly wind, an early MCS (a group of convective cells) for pre-Aila (2009) is initiated in the east-west direction along the horizontal shear line. The convective cells are embedded in a strong vertical shear environment where the strong southwesterly flows beneath the weak north and northwesterly (Fig. 7b). Consequently, low-level mesovortices with horizontal diameters of $\sim 6\text{--}8 \text{ km}$ develop along with the individual convective cells under the effect of an environmental low-level (mid-level) vertical wind shear of $\sim 20 \text{ m s}^{-1}$ ($\sim 27 \text{ m s}^{-1}$) within 0–3 km (0–6 km) AGL. A similar mesovortex genesis in the mature phase of QLCS has been noticed in the previous studies (Trapp and Weisman 2003; Weisman and Trapp 2003). Moreover, Atkins and Laurent (2009b) observed mesovortices during different stages of the convective system including the early cellular stage. The mesovortices are extended vertically due to the tilting of the crosswise horizontal vorticity by the updraft, which results in a mesovortex pair. Figure 7b displays the vertical vorticity distribution showing

the mesovortex pair (named “X”) due to the growing updraft with a maximum velocity of 7 m s^{-1} accompanied by a thunderstorm displaying a weak hook-like structure in the vicinity of the updraft. The vortex-pair is aligned in the north-south direction with a cyclonic rotation in the north (to the right of the vertical shear vector) and an anticyclonic in the south (to the left of the vertical shear vector), similar to a supercell mesocyclone (Klemp 1987). Previous studies have found that mesovortices can produce short-lived tornadoes but are weak with respect to the damaging straight-line winds linked with mesovortices (Weisman and Trapp 2003). The highest positive (negative) vertical vorticity of $2.8 \cdot 10^{-3} \text{ s}^{-1}$ ($-1.7 \cdot 10^{-3} \text{ s}^{-1}$) develops approximately at the 700 hPa level because of the strongest vertical shear within this layer in association with dryline boundary aloft in the BoB.

4.4.2 Movement and life cycle of mesovortices

The vortex-couplet “X” located at 11.4°N , 83.7°E in Fig. 7a is found moving in the southeast direction. Subsequently, after one and a half hours, i.e., at 0600 UTC 21 May 2009, the vortex-couplet takes a position at 11.33°N , 84.12°E , as shown in Fig. 8a. In Fig. 8b, the vertical cross-section displays the strong updraft (maximum of 13 m s^{-1}) extending up to 200 hPa, which amplifies the vortex pair by the stretching of vortex tubes. During the mature stage of the vortex-pair, a maximum downdraft of -1.5 m s^{-1} is observed, which remains separated from the updraft, indicating the presence of supercell characteristics (Lemon and Doswell 1979; Davies-Jones 1984). The height of the vortices is calculated as 9 km approximately. The maximum vertical vorticity of $\sim 3 \cdot 10^{-3} \text{ s}^{-1}$ at the mid-level, connected with the intense updraft, indicates the possibility of the presence of a low-level mesocyclone in the tropical environment.

Moving of “X” toward the southeast direction or the rear side within the QLCS and the development of a new convective cell at the west side of “X” (Fig. 7a) inhibit the inflow of warm air to the old cell, which induces a reduction in the updraft as well as in the cell growth. Under these circumstances, the vortex pair is going to be weakened. Figures 8c-d display the horizontal and vertical structures of “X” that represent the dissipating stage at 0930 UTC 21 May 2009, where the updraft is diminished. The positive or cyclonic vortex of “X” is continuing its movement a little toward the rear side of the QLCS after dissipating the negative vortex. The vortex-pair reaches the average speed of approximately 7 m s^{-1} during its lifetime of ~ 6 hours.

Figure 9a exhibits the lifespan of three individual cyclonic mesovortices indicated by A, B, and C at 1-hour intervals for understanding their movement directions. All vortices are developed along the convective lines of QLCS represented by the solid lines. The vortex “A” appeared at 1000 UTC and dissipated at 1900 UTC 21 May 2009. Similarly, “B” and “C” both had lifetimes from 0800 UTC to 1700 UTC 21 May 2009. It is clear that the system remained almost stationary for a long period from 0900 to 2100 UTC 21 May 2009; however, cyclonic vortices that developed in the convective side gradually moved to the rear side of the QLCS. The arc-shaped movement tracks of the mesovortices and their speeds depend on the mean wind which is dominated by the southwesterly wind (Fig. 6a) and their speeds are influenced by the

intensity of the wind. The Hovmöller diagram of average vorticity advection from 1000 to 500 hPa for the area of 84–88°E during the period from 0330 UTC to 2230 UTC 21 May 2009 (Fig. 9b) also supports the advection of cloud-scale cyclonic values of vorticity to higher latitudes and intensification of the vorticity at the north as a consequence.

A similar development of vortices is frequent along the shear line or leading edge of the QLCS at different convective stages but the highest frequency is observed in the mature stage. The characteristics of all mesovortices are not similar to “X”. They are more or less intense in terms of value, height, and lifetime of the vertical vorticity. The maximum height of the vortices varies from 600 to 200 hPa above the surface and the lifetime ranges from 5 to 10 hours. Some vortices are found to be merged with others, whereas some split during their lifetime. In particular, the characteristics of the mesovortices depend on the associated convective cells embedded in the QLCS. In this study, all types of cloud-scale vortices are considered as mesovortices based on the horizontal diameter.

5. Discussion

5.1. Mesoscale convective vortex for pre-cyclones

By definition, MCV is a low-pressure center at mid-level that develops in the stratiform rain region of an MCS due to latent heat release. The origin of tropical cyclone circulations is suggested to be the MCV embedded within a low-level vorticity-rich environment (Houze 2004). Dunkerton et al. (2009) referred to it as a “pouch” where a closed cyclonic circulation is protected by a large-scale environment. In this study’s simulation, the MCSs for all pre-cyclones were convective for the most part, where the stratiform precipitation regions were observed as very narrow (Figs. 4d-f). The linear convective system containing narrow or no-stratiform precipitation region, which depends on storm-relative elevated front-to-rear flow, has also been noticed previously (Gallus et al. 2008; Lombardo and Colle 2013). Figure 3a also agrees with the existence of a very narrow stratiform precipitation region in the rainband of Aila (2009) imaged from the satellite data. Further, the weak convergence and smaller negative geopotential height anomalies in the stratiform region shown in Fig. 6d do not support any significant low pressure in the middle level, which can be representative of MCV for cyclogenesis.

The MCS of pre-Aila (2009) and associated low-level moisture advection by the synoptic-scale flow are displayed in the left panel in Fig. 10, whereas the right panel shows the zonal and vertical average of the wind and potential temperature (q). In the figures, the QLCSs that mature within the zone of convergence have emerged in an environment of potential temperature gradient of $\sim 2^\circ\text{C}$ and within the changing fluxes of $\sim 150\text{ g m}^{-2}\text{ s}^{-1}$ between the dry air in the north and moist air from the south. The streamline and also the average wind in the zonal direction indicates a meridional shear along the dry-moist boundary that causes shear vorticity in the synoptic scale. In addition, the streamline indicates a counterclockwise vortex growing at 1330 UTC due to the spinning dry and hot air in the northernmost part of the MCS, which further enhances, as shown after 6 hours at 1930 UTC 21 May 2009.

The variations of average vorticity (both relative and shear) for the two equal areas (A and B indicated in Fig. 10) with respect to time are calculated and displayed in Fig. 11a for the period of 0430 UTC to 2330 UTC 21 May 2009. During the majority of these 19 hours, the lower area “A” covers the entire convective parts of the MCS, whereas the upper area “B” does not include the MCS after the initial few hours. At the initial stage of the MCS, both areas have almost similar intensities of shear vorticity ($\sim 2 \times 10^{-5} \text{ s}^{-1}$) or relative vorticity ($\sim 4 \times 10^{-5} \text{ s}^{-1}$) that are supposed to come from the shear instability along the dryline. During the mature stage at approximately 1200 UTC, the cyclonic vorticity starts increasing in area B by the advection of strong cloud-scale positive vortices to the northern side of QLCS (Fig. 9b). In other words, the mesovortices that formed in the deep convection within the convergent boundary move to the rear side of the QLCS enhancing the relative vorticity to a maximum of $8.7 \times 10^{-5} \text{ s}^{-1}$ in area B. The linear trend in the figure clearly shows that the relative vorticity in area B is increasing significantly after 1100 UTC with the value of the coefficient of determination (R^2) of 0.62, which represents a good fit of the line to the data. On the other hand, in area A the vorticity remains constant or decreases very little even though the area contains the convection. Moreover, a similar pattern in the variations of vorticities confirms that the contribution of horizontal shear vorticity mainly influences the relative vorticity with the curvature vorticity remaining nearly constant ($\sim 2 \times 10^{-5} \text{ s}^{-1}$) through the entire period. The values of cyclonic vorticity are higher in the level 850-600 hPa and the dimension of the vortex is approximately $\sim 200 \text{ km}$, which is similar to a typical mid-tropospheric MCV in the rear side of the QLCS (Houze 2004; 2010).

Similar results are also observed for the pre-BoB_02 (2004) and pre-Mora (2017) cyclones shown in Figs. 10b and 10c, respectively. In both cases, the relative vorticity starts to increase during the mature convective stage and shows positive trends ($R^2=0.95$ for BoB_02 and $R^2=0.94$ for Mora) for the northern areas of the pre-cyclones but shows a negative or constant trend for the southern areas. The area coverages for both the pre-cyclones are mentioned in the figure and have been selected according to the position of MCSs in each case (Figs. 4d and 4f).

Therefore, during the premonsoon, the meridional shear connected with the dryline forms positive vorticity ($\sim 10^{-5} \text{ s}^{-1}$) from the lower to middle levels in the synoptic-scale and the vertical wind shear influences a number of cloud-scale mesovortices to be produced at the lower level. Fast spinning, intense mesovortices ($\sim 10^{-3} \text{ s}^{-1}$) and their positive advection trigger and enhance a mesoscale cyclonic vortex to the north of QLCS. However, for cyclogenesis, Hendricks et al. (2004) introduced a term VHT (horizontal diameter $\sim 10 \text{ km}$), which characterizes the highly positive vorticity centering within an extended updraft, and an early MCS begins with a set of one or more isolated deep VHTs (Houze 2010). Among the mesovortices, some may form into VHTs, which are not separately identified here, because all meso- γ -scale vortices are considered as mesovortices at the cloud scale. In this study, the TC formation process was observed to have similarity with the bottom-up paradigms (Hendricks et al. 2004; Reasor et al. 2005; Montgomery et al. 2006; Braun et al. 2010) rather than with top-down processes (Emanuel 1993; Bister and Emanuel 1997; Ritchie and Holland 1997). The vorticity along the dryline does not penetrate

downward to form a surface vortex, whereas, within the embryonic environment of shear vorticity, low-level mesovortices are formed, integrated, and finally turned into a cyclone vortex.

In brief, during the premonsoon, an MCS is initiated along the shear line with intense vertical low-level mesovortices, which are the fundamental constituents following the bottom-up paradigms for cyclogenesis over the BoB.

6. Summary

Bimodal cyclone distribution in the Bay of Bengal is one of its unique characteristics, which is primarily associated with seasonal monsoon trough locations. Another important climate feature is the presence of dryline aloft over the northern BoB during the premonsoon season. Cyclones formed near or along the dryline quickly evolve to severe TCs within a very short distance in a small time period and frequently have severe consequences after landfall. On that account, the formation of TCs that develop near the dryline is analyzed in this study during 2001-2020. As a representative of dryline-associated TCs, three very severe cyclonic storms, i.e., BoB_02 (2004), Aila (2009), and Mora (2017) are simulated during their genesis period using the WRF model with the highest resolution of 4 km in the inner domain. The simulated results are then validated by comparing them with datasets of NCEP_CFSR reanalysis and JTWC best track data. The events are well simulated by the model, which allows the analysis of mesoscale phenomena associated with cyclogenesis. As the MCS is one of the primary characteristics that describe the activities of cyclogenesis, the convective systems and related environments for pre-BoB_02 (2004), pre-Aila (2009), and pre-Mora (2017) are examined in detail. Finally, the formation process of an MCV, which is the nucleus of a cyclone vortex, is discussed. The important findings are summarized below:

Environment:

During premonsoon, a synoptic dryline, which incorporates a deep hot, dry air flow in the north and northwest and a shallow warm, moist air flow in the south and southeast of the BoB, causes temperature and moisture gradients over the BoB. The dryline over the BoB is basically an inclined moisture gradient aloft appearing when shallow air masses penetrate the deep layers. The elevated dryline not only causes a horizontal shear in the meridional direction but also produces a vertical wind shear due to the change in the strong southwesterly of the lower level to an upper level weak northwesterly wind. The boundary between the two air masses with different densities helps the high q_e flow from the southwesterly to be uplifted for initiating updrafts. As a result, a group of linear convective cells originates in an environment of deep-layer strong clockwise shear and intense CAPE, which favors the formation of supercells. The environmental conditions calculated from the indices are observed to be potent, which leads to the initiation of severe thunderstorms capable of strong wind, rotating updrafts, and heavy precipitation.

MCS:

Due to the influence of linear forcing along the boundary of the dryline, the early linear MCS remains nearly stationary for longer periods and matures as a comma shape MCS or QLCS within the zone of enhanced convergence. Accordingly, the QLCS appears as a large system with a horizontal length of ~500 km oriented approximately in the east-west direction. The system comprises leading-line deep convections including supercells and follows a very narrow trailing stratiform region. The southwesterly inflow facilitates low-level moisture to the QLCS supporting a convective line in the south of the system, whereas 850-500 hPa dry air coming from the north and northwest functioning as the RIJ assist the QLCS to turn into a comma-shaped convective system. Evaporative cooling accelerates the downdrafts that support a weak cold pool on the surface extending from the leading edge rearward. The weak cold pool and strong shear instability generate new cells in the front of the system in a tropical environment. Once developed, the individual mature deep convections move to the rear side extending the trailing stratiform precipitation region to the north of the QLCS. The stronger shears lead to a longer QLCS life, which is approximately 20 hours or more.

MCV:

According to the results of this study, the mid-tropospheric conventional MCV that is associated with the stratiform rain region within the quasi-linear MCS is not significant enough to be a part of the cyclogenesis. The uplifted dryline over the BoB produces horizontal positive shear vorticity in the synoptic-scale with high values (order 10^{-5}) in the level 850-600 hPa. Within this embryonic environment, the strong vertical shear veering with height initiates low-level mesovortices ($\sim 10^{-3} \text{ s}^{-1}$) with 6-8 km length in the individual thunderstorms of QLCS at the different convective stages. The tilting of ambient crosswise horizontal vorticity by the local updrafts supports the appearance of a mesovortex pair with intense cyclonic (less intense anti-cyclonic) rotation to the right (left) of the vertical shear vector. Some mesovortices have the characteristics of mesocyclones with stronger updrafts depending on the convective type embedded in the QLCS. After forming along the leading edge of the QLCS, the cloud-scale mesovortices mature and move to the rear side due to the influence of the mean wind flow. These fast-spinning cyclonic vortices and their positive advection to the rear side trigger a cyclonic vortex in mesoscale, which performs as a core vortex similar to a conventional MCV. The core vortex continues beyond the life cycle of the QLCS, which further allows the formation of new convective systems and develops the TC vortex subsequently.

Therefore, within the cyclonic vorticity-rich environment due to the synoptic-scale horizontal shear, cloud-scale cyclonic mesovortices are the essential prerequisites responsible for the formation of the MCV in the vicinity of MCS for cyclogenesis connected to the dryline over the BoB. Embedded supercells with intense mesovortices (or mesocyclones) in a favorable environment of quasi-linear MCS may cause a rapid intensification of the TC vortex within a short distance. Further study using a cloud-resolving model is necessary to examine and comprehend the contributions of different kinds of mesovortices associated with the TCs formed over the BoB.

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The data required to reproduce these findings cannot be shared at this time as the data also forms part of an ongoing study.

Conflict of interest

The authors declare that they have no conflict of interest.

Ethical approval

This article does not contain any studies with human participants or animals performed by the author. It also has not been published by another journal.

References

- Akter N (2015) Mesoscale Convection and Bimodal Cyclogenesis over the Bay of Bengal. *Mon Weather Rev* 143:3495-3517
- Akter N, Tsuboki K (2014) Role of synoptic-scale forcing in cyclogenesis over the Bay of Bengal. *Clim Dynam* 43: 2651-2662
- Akter N, Tsuboki K (2017) Climatology of the premonsoon Indian dryline. *Int J Climatol* 37:3991-3998
- American Meteorological Society (2019) Mesoscale convective vortex. Glossary of Meteorology. http://glossary.ametsoc.org/wiki/Mesoscale_convective_vortex.
- Atkins NT, Laurent MS (2009a) Bow echo mesovortices. Part I: Processes that influence their damaging potential. *Mon Weather Rev* 137:1497-1513.

Atkins NT, Laurent MS (2009b) Bow echo mesovortices. *Part II: Their Genesis*. Mon Weather Rev 137:1514-1532

Bangladesh Red Crescent Society (2017) Bangladesh: Cyclone Mora. Emergency appeal no. MDRBD019, International federation of Red Cross and Red Crescent societies. [Available online at <https://reliefweb.int/report/bangladesh/bangladesh-cyclone-mora-emergency-appeal-operations-update-n-1-mdrbd019/>]

Bartels DL, Maddox RA (1991) Midlevel cyclonic vortices generated by mesoscale convective systems. Mon Weather Rev 119:104-118

Bhat GS, Srinivasan J, Gadgil S (1996) Tropical deep convection, convective available potential energy and sea surface temperature. J Meteorolog Soc Jpn 74:155-166

Bikos D, Finch J, Case JL (2016) The environment associated with significant tornadoes in Bangladesh. Atmos Res 167: 183–195

Bister M, Emanuel K (1997) The genesis of hurricane Guillermo: TEXMEX analyses and a modeling study. Mon Weather Rev 125:2662-2682

Bluestein HB, Jain MH (1985) Formation of mesoscale lines of precipitation: Severe squall lines in Oklahoma during the spring. J Atmos Sci 42:1711-1732

Bluestein HB (1993) Synoptic-Dynamic Meteorology in Midlatitudes: Volume II: Observations and Theory of Weather Systems, Oxford University Press, New York.

Bosart LF, Sanders F (1981) The Johnstown flood of July 1977: A long-lived convective system. J Atmos Sci 38: 1616-1642

Braun SA, Montgomery MT, Mallen KJ, Reasor PD (2010) Simulation and interpretation of the genesis of tropical storm Gert (2005) as part of the NASA tropical cloud systems and processes experiment. J Atmos Sci 67: 999-1025

Bunkers MJ (2002) Vertical wind shear associated with left-moving supercells. Weather Forecasting 17:845-855

Camargo SJ, Emanuel KA, Sobel AH (2007) Use of a genesis potential index to diagnose ENSO effects on tropical cyclone genesis. J Clim 20:4819-4834

Chen F, Dudhia J (2001) Coupling an advanced land-surface/hydrology model with the Penn State/NCAR MM5 modeling system. Part I: Model implementation and sensitivity. Mon Weather Rev 129:569-585

Chen SS, Frank WM (1993) A numerical study of the genesis of extratropical convective mesovortices. Part I: Evolution and dynamics. J Atmos Sci 50:2401-2426

The COMET Program/UCAR (1999) Mesoscale convective systems: squall lines and bow echoes, University Corporation for Atmospheric Research, Cooperative Program for Operational Meteorology, Education, and Training (COMET), Boulder, CO. [Available online at <http://meted.ucar.edu/convectn/mcs/>]

Davies-Jones R (1984) Streamwise vorticity: The origin of updraft rotation in supercell storms. *J Atmos Sci* 41:2991-3006

Davies-Jones R, Burgess D, Foster M (1990) Test of helicity as a forecast parameter. Preprints, 16th Conf. on severe local storms, Kananaskis Park, AB, Canada, Am Meteorol Soc, pp 588-592.

Davis C, Wang W, Chen SS et al (2008) Prediction of landfalling hurricanes with the Advanced Hurricane WRF Model. *Mon Weather Rev* 136:1990-2005

Davis CA, Weisman ML (1994) Balanced dynamics of mesoscale vortices produced in simulated convective systems. *J Atmos Sci* 51:2005-2030

Davis CA, Atkins N, Bartels D et al (2004) The Bow Echo and MCV Experiment: Observations and opportunities. *Bull Am Meteorol Soc* 85:1075-1093

Davis CA, Trier SB (2007) Mesoscale convective vortices observed during BAMEX. Part I: Kinematic and thermodynamic structure. *Mon Weather Rev* 135:2029-2049

Donelan MA, Haus BK, Reul N, Plant WJ, Stiassnie M, Graber HC, Brown OB, Saltzman ES (2004) On the limiting aerodynamic roughness of the ocean in very strong winds. *Geophys Res Lett* 31:L18306, <https://doi.org/10.1029/2004GL019460>.

Dudhia J (1989) Numerical study of convection observed during the winter monsoon experiment using a mesoscale two-dimensional model. *J Atmos Sci* 46:3077–3107

Dudhia J, Done J, Wang W, Chen Y, Xiao Q, Davis C, Holland G, Rotunno R, Torn R (2008) Prediction of Atlantic tropical cyclones with the Advanced Hurricane WRF (AHW) model, 28th AMS Conference on hurricanes and tropical meteorology, Paper 18A2, Orlando. [Available online at http://ams.confex.com/ams/28Hurricanes/techprogram/paper_138004.htm/]

Dunkerton TJ, Montgomery MT, Wang Z (2009) Tropical cyclogenesis in a tropical wave critical layer: easterly waves. *Atmos Chem Phys* 9:5587–5646

Emanuel KA (1993) The physics of tropical cyclogenesis over the eastern Pacific. In: Lighthill J et al (eds) *Tropical Cyclone Disasters*, Peking University Press, pp 136-142

Ferrier BS, Jin Y, Lin Y, Black T, Rogers E, DiMego G (2002) Implementation of a new gridscale cloud and precipitation scheme in the NCEP Eta model. Preprints, 15th Conf. on numerical weather prediction, San Antonio, TX, Am Meteorol Soc, pp 280-283

- Fritsch JM, Murphy JD, Kain JS (1994) Warm core vortex amplification over land. *J Atmos Sci* 51:1780-1807
- Gallus Jr. WA, Snook NA, Johnson EV (2008) Spring and summer severe weather reports over the midwest as a function of convective mode: A preliminary study. *Weather Forecasting* 23:101-113
- Galway JG (1956) The lifted index as a predictor of latent instability. *Bull Am Meteorol Soc* 37:528-529
- George JJ (1960) *Weather Forecasting for Aeronautics*. Academic Press, Waltham, MA
- Gjorgjievska S, Raymond DJ (2014) Interaction between dynamics and thermodynamics during tropical cyclogenesis. *Atmos Chem Phys* 14:3065-3082
- Glickman TS (ed) (2000) *Glossary of Meteorology*. 2nd edn. Am Meteorol Soc, Boston [Available online at <http://glossary.ametsoc.org/>.]
- Gray WM (1998) The *formation of tropical cyclones*. *Meteorol Atmos Phys* 67:37–69
- Harr PA, Chan JCL (2005) Monsoon impacts on tropical cyclone variability. In: Chang C-P, Wang B, Lau N-CG (eds) *The Global Monsoon System: Research and Forecast*, WMO, pp 512–542
- Harr PA, Elsberry RL, Chan JCL (1996) Transformation of a large monsoon depression to a tropical storm during TCM-93. *Mon Weather Rev* 124:2625-2643
- Hendricks EA, Montgomery MT, Davis CA (2004) On the role of “vortical” hot towers in formation of tropical cyclone Diana (1984). *J Atmos Sci* 61:1209-1232
- Hertenstein RFA, Schubert WH (1991) Potential vorticity anomalies associated with squall lines. *Mon Weather Rev* 119:1663-1672
- Hong S-Y, Noh Y, Dudhia J (2006) A new vertical diffusion package with an explicit treatment of entrainment processes. *Mon Weather Rev* 134:2318-2341
- Houze Jr. RA (2004) *Mesoscale convective systems*. *RevGeophys*42:RG4003, <https://doi.org/10.1029/2004RG000150>
- Houze Jr. RA (2010) Clouds in tropical cyclones. *Mon Weather Rev* 138:293-344
- James EP, Johnson RH (2010) A climatology of midlatitude mesoscale convective vortices in the rapid update cycle. *Mon Weather Rev* 138:1940-1956
- Kain JS (2004) The Kain-Fritsch convective parameterization scheme: An update. *J Appl Meteorol* 43:170-181

- Kikuchi K, Wang B (2010) Formation of tropical cyclones in the Northern Indian Ocean associated with two types of tropical intraseasonal oscillation modes. *J Meteorol Soc Jpn* 88:475-496
- Kirk JR (2003) Comparing the dynamical development of two mesoscale convective vortices. *Mon Weather Rev* 131:862-890
- Kirk JR (2007) A phase-plot method for diagnosing vorticity concentration mechanisms in mesoscale convective vortices. *Mon Weather Rev* 135:801-820
- Klemp JB (1987) Dynamics of tornadic thunderstorms. *Ann Rev Fluid Mech* 19:369-402
- Krishna KO, Mohanty UC, Routray A, Kulkarni MA, Mohapatra M (2012) Customization of WRF-ARW model with physical parameterization schemes for the simulation of tropical cyclones over North Indian Ocean. *Nat Hazards* 63:1337–1359
- Lee M, Frisius T (2018) On the role of convective available potential energy (CAPE) in tropical cyclone intensification, *Tellus A* 70:1, 1433433, <https://doi.org/10.1080/16000870.2018.1433433>
- Lemon LR, Doswell III CA (1979) Severe thunderstorm evolution and mesocyclone structure as related to tornadogenesis. *Mon Weather Rev* 107:1184-1197
- Lin Y-L (2007) *Mesoscale Dynamics*. Cambridge University Press, <https://doi.org/10.1017/CBO9780511619649>
- Lombardo K, Colle BA (2013) Processes controlling the structure and longevity of two quasi-linear convective systems crossing the southern New England coast. *Mon Weather Rev* 141:3710-3734
- May PT, Rajopadhyaya DK (1999) Vertical velocity characteristics of deep convection over Darwin, Australia. *Mon Weather Rev* 127:1056-1071
- McBride JL (1995) Tropical cyclone formation. In: Elsberry RL (ed) *Global perspectives on tropical cyclones*, WMO/TD-No. 693, World Meteorological Organization, Geneva, Switzerland, pp 63-105
- Miller RC (1972) *Notes on Analysis and Severe-Storm Forecasting Procedures of the Air Force Global Weather Central*. Weather Service Tech. Rep. 200 (Rev.), Air Weather Service, Scott AirForce Base, IL.
- Mlawer EJ, Taubman SJ, Brown PD, Iacono MJ, Clough SA (1997) Radiative transfer for inhomogeneous atmosphere: RRTM, a validated correlated-k model for the long-wave. *J Geophys Res* 102:16663-16682
- Monin AS, Obukhov AM (1954) Basic laws of turbulent mixing in the ground layer of the atmosphere. *Trudy Geofiz. Inst. Akad. Nauk SSSR*, 151:163–187 (Translated by Am Meteorol Soc in 1959) [Available online at <http://handle.dtic.mil/100.2/AD672723>]
- Molinari J, Romps DM, Vollaro D, Nguyen L (2012) CAPE in Tropical Cyclones. *J Atmos Sci* 69:2452-2463

Molinari J, Vollaro D (2009) Distribution of Helicity, CAPE, and Shear in Tropical Cyclones. *J Atmos Sci* 67:274-284.

Montgomery MT, Enagonio J (1998) Tropical cyclogenesis via convectively forced vortex Rossby waves in a three-dimensional quasi geostrophic model. *J Atmos Sci* 55:3176-3207

Montgomery MT, Nicholls ME, Cram TA, Saunders AB (2006) A vortical hot tower route to tropical cyclogenesis. *J Atmos Sci* 63:355-386

Myanmar Red Cross Society (2004) Myanmar: Cyclone Rakhine. Appeal no. 14/04, International Federation of Red Cross and Red Crescent Societies. [Available online at <https://reliefweb.int/report/myanmar/myanmar-cyclone-rakhine-appeal-no-142004-final-report/>]

Ono Y (1997) Climatology of tornadoes in Bangladesh, 1990–1994. *J Meteorol* 22: 325–340

Orlanski I (1975) A rational subdivision of scales for atmospheric processes. *Bull Am Meteorol Soc* 56:527-530

Parker MD, Johnson RH (2000) Organizational modes of midlatitude mesoscale convective systems. *Mon Weather Rev* 128:3413-3436

Pollard RT, Rhines PB, Thompson RORY (1973) The deepening of the wind-mixed layer. *Geophys Fluid Dyn* 3:381-404.

Powers JG, Davis CA (2002) A cloud-resolving regional simulation of tropical cyclone formation, *Atmos Sci Lett* <https://doi.org/10.1006/asle.2002.0054>.

Raju PVS, Potty J, Mohanty UC (2011) Sensitivity of physical parameterizations on prediction of tropical cyclone Nargis over the Bay of Bengal using WRF model. *Meteorol Atmos Phys* 113:125-137

Rapp AD, Kummerow CD, Fowler L (2011) Interactions between warm rain clouds and atmospheric preconditioning for deep convection in the tropics. *J Geophys Res* 116:D23210, <https://doi.org/10.1029/2011JD016143>.

Reasor PD, Montgomery MT, Bosart L (2005) Mesoscale observations of the genesis of Hurricane Dolly (1996). *J Atmos Sci* 62:3151-3171

Ritchie EA, Holland GJ (1997) Scale interaction during the formation of typhoon Irving. *Mon Weather Rev* 125: 1377-1396

Rotunno R, Klemp JB, Weisman ML (1988) A theory for strong, long lived squall lines. *J Atmos Sci* 45:463-485

Roy K, Kumar U, Mehedi H, Sultana T, Ershad DM (2009) Initial Damage Assessment Report of Cyclone AILA with focus on Khulna District, Unnayan Onneshan-Humanity Watch- Nijera Kori, Khulna, Bangladesh,

pp 31 [Available online at

https://www.researchgate.net/publication/273771123_Cyclone_Aila_2009_Initial_Assessment_Report_with_focus_on_Khulna_District]

Saha S, Moorthi S, Pan H-L, Wu X, Wang J, Nadiga S et al (2010) The NCEP Climate Forecast System reanalysis. *Bull Am Meteorol Soc* 91:1015-1057

Sanders F (1999) A proposed method of surface map analysis. *Mon Weather Rev* 127:945-955

Skamarock WC, Weisman ML, Klemp JB (1994) Three dimensional evolution of simulated long-lived squall lines. *J Atmos Sci* 51:2563-2584

Skamarock WC, Klemp JB, Dudhia J et al (2008) A description of the Advanced Research WRF version 3. NCAR Tech. Note NCAR/TN-4751STR, pp 113 [Available online at http://www.mmm.ucar.edu/wrf/users/docs/arw_v3_bw.pdf]

Simpson J, Ritchie EA, Holland GJ, Halverson J, Stewart S (1997) Mesoscale interactions in tropical cyclone genesis. *Mon Weather Rev* 125:2643-2661

Trapp RJ, Weisman ML (2003) Low-level mesovortices within squall lines and bow echoes. Part II: Their genesis and implications. *Mon Weather Rev* 131:2804-2817

Trier SB, Davis CA (2007) Mesoscale convective vortices observed during BAMEX. Part II: Influences on secondary deep convection. *Mon Weather Rev* 135:2051-2075

United Nations (2010) Cyclone Aila: Joint UN Multi-Sector Assessment & Response Framework. United Nations, New York. [Available online at <https://www.scribd.com/document/79662690/Aila-UN-Assessment-Framework-FINAL/>]

Weisman ML (1993) The genesis of severe, long-lived bow-echoes. *J Atmos Sci* 50:645-670

Weisman ML, Klemp JB (1982) The dependence of numerically simulated convective storms on vertical shear and buoyancy. *Mon Weather Rev* 110:504-520

Weisman ML, Rotunno R (2004) "A theory for strong, long-lived squall lines" Revisited. *J Atmos Sci* 61:361-382

Weisman ML, Trapp RJ (2003) Low-level mesovortices within squall lines and bow echoes. Part I: Overview and sensitivity to environmental vertical wind shear. *Mon Weather Rev* 131:2779-2803

Williams ER, Renno NO (1993) An analysis of the conditional instability of the tropical atmosphere. *Mon Weather Rev* 121:21-36

Yamane Y, Hayashi T, Dewan AM, Akter F (2009) Severe local convective storms in Bangladesh: part I. *Climatology. Atmos Res* 95:400–406

Figures

Figure 1

a Surface CIN (shaded; $J kg^{-1}$) and the zero anomalies of average specific humidity (red dashed line) from surface to 700 hPa for premonsoon (March-May) during 2001-2020 using NCEP-CFSR reanalysis. The anomalies are calculated from the average values of specific humidity for the area in the figure. The dots represent the positions of very severe cyclonic storm (MSW ≥ 65 kt) at their initial depression levels (MSW < 25 kt). The track of BoB_02 (2004), Aila (2009), and Mora (2017) are represented by violet, green and black lines, respectively. b The vertical cross-section of streamline with specific humidity along AB in Fig. 1a averaged for premonsoon from 2001-2020. Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

Figure 2

a-c Simulated surface-based CAPE (shaded; J kg^{-1}) averaged from TC depression to formation period, surface pressure (contours; 2-hPa interval), wind velocity (vector; m s^{-1}) at 850 hPa at the genesis of each cyclone. d-f The same parameters as Fig. 2a-c using NCEP-CFSR reanalysis data for each TC. The white dots are the centers of each TC at genesis using JTWC data

Figure 3

a Rainfall rate in mm hr⁻¹ from (a) TRMM satellite data at 2118 UTC 23 May 2009 (available in https://sharaku.eorc.jaxa.jp/cgi-bin/typ_db/typ_db.cgi). b Simulation at 0900 UTC 24 May 2009 for Aila (2009)

Figure 4

Reflectivity (shaded; dBZ) at 925 hPa, temperature anomalies (contours; 1-K interval) calculated from the average area of D2 and wind velocity (vector; m s⁻¹) with moisture flux (color shaded; g m⁻² s⁻¹) at 850 hPa for pre-cyclones of a BoB_02 (2004) b Aila (2009) and c Mora (2017). d-f Mature MCS (shaded, dBZ) at 925 hPa and 500-hPa wind velocity (vector; m s⁻¹) with geopotential height (color shaded; m) for the same pre-cyclones, respectively. The locations of star in the left panel are the sounding points

Figure 5

Vertical cross-section of equivalent potential temperature (shaded; K), vertical velocity (green contours; 0.2 m s^{-1} interval) and horizontal wind velocity (vector; m s^{-1}) for a pre-BoB_02 at 13.5°N for 0230 UTC 14 May 2004 b pre-Aila at 11°N for 0430 UTC 21 May 2009 c pre-Mora at 12.8°N for 1100 UTC 25 May 2017

Figure 6

a Reflectivity of 10-dBZ at 925 hPa for 1030 UTC 21 May 2009 (green hatched area) and for 2030 UTC 21 May 2009 (red hatched area). The 850 hPa horizontal wind (vectors; m s^{-1}) at 1030 UTC 21 May 2009 and vorticity of $1 \times 10^{-3} \text{ s}^{-1}$ (black contours) at 2030 UTC 21 May 2009. Vertical reflectivity (shaded; dBZ), system-relative wind (vectors; m s^{-1}), updraft (white solid contours; 2 m s^{-1} interval), and downdraft (white dashed contours; -0.5 m s^{-1} intervals) along the line b AB at 1030 UTC 21 May 2009 and c CD at 2030 UTC 21 May 2009 indicated in figure a. d Vertical cross-section along CD for potential temperature anomalies (shaded; K) and geopotential height anomalies (black contours; 10 m interval) calculated from background area in Fig. a and convergence (green contours; $-0.5 \times 10^{-3} \text{ s}^{-1}$ interval)

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Figure 7

a Meridional wind (shaded; m s^{-1}), horizontal relative vorticity of $1 \times 10^{-3} \text{ s}^{-1}$ (red contours for cyclonic and black contours for anti-cyclonic vorticity) at 700 hPa and 20-dBZ reflectivity (blue contours). b The vertical cross-section of vorticity "X" (shaded; m s^{-1}) indicated in Fig. a, reflectivity (hatched area; 10-dBZ), vertical velocity (red contours; 2 m s^{-1} interval), and wind velocity (vector; m s^{-1}) at 83.7°E for 0430 UTC 21 May 2009

Figure 8

a Horizontal relative vorticity at 875 hPa (shaded; s^{-1}), b Vertical cross-section of the vorticity "X" (shaded; s^{-1}), updraft (solid contours; 2 m s^{-1} interval), and downdraft (dash contours; 0.4 m s^{-1} interval) along AB at 0600 UTC 21 May 2009. c and d are the same as figures a and b, respectively at 0930 UTC 21 May 2009

Figure 9

a Solid lines represent the convective lines of QLCS at 0900 UTC, 1300 UTC, 1700 UTC, and 2100 UTC 21 May 2009. The locations and life cycles of the three cyclonic vortices (A, B, and C) at 650 hPa are shown by the shade with 1-hour intervals. b Hovmöller diagram of vorticity advection ($\times 10^{-8} \text{ s}^{-2}$) averaged between 84–88°E from 1000 to 500 hPa during the period from 0330 UTC to 2230 UTC 21 May 2009

Figure 10

Mature-stage MCS (shaded; dBZ) and streamline with moisture flux (color shades; $\text{g m}^{-2} \text{s}^{-1}$) at 925 hPa in the left panel. Zonal (82–89°E) and vertical (1000–500 hPa) averaged wind (red line; m s^{-1}) and potential temperature (blue line; $^{\circ}\text{C}$) in the right panel at a 1330 UTC 21 May 2009 b 1930 UTC 21 May 2009

Figure 11

a Average relative vorticity (s^{-1}) and shear vorticity (s^{-1}) from 1000 to 500 hPa for two areas (A: 84–88°E, 9–11°N and B: 84–88°E, 11–13°N) indicated in Fig. 10 from 0430 to 2330 UTC 21 May 2009 for Aila. The same average relative vorticity (s^{-1}) for b BoB_02 from 0230 to 2130 UTC 14 May 2004 and c Mora from 1100 UTC 25 May to 0600 UTC 26 May 2017