The structure and evolution characteristics of Rossby wave trains induced by tropical cyclone (TC) energy dispersion are revealed based on the Quick Scatterometer (QuikSCAT) and Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) data. Among 34 cyclogenesis cases analyzed in the western North Pacific during 2000–01 typhoon seasons, six cases are associated with the Rossby wave energy dispersion of a preexisting TC. The wave trains are oriented in a northwest–southeast direction, with alternating cyclonic and anticyclonic vorticity circulation. A typical wavelength of the wave train is about 2500 km. The TC genesis is observed in the cyclonic circulation region of the wave train, possibly through a scale contraction process.

The satellite data analyses reveal that not all TCs have a Rossby wave train in their wakes. The occurrence of the Rossby wave train depends to a certain extent on the TC intensity and the background flow. Whether or not a Rossby wave train can finally lead to cyclogenesis depends on large-scale dynamic and thermodynamic conditions related to both the change of the seasonal mean state and the phase of the tropical intraseasonal oscillation. Stronger low-level convergence and cyclonic vorticity, weaker vertical shear, and greater midtropospheric moisture are among the favorable large-scale conditions. The rebuilding process of a conditional unstable stratification is important in regulating the frequency of TC genesis.

1. Introduction

Tropical cyclone (TC) genesis is the transformation process of a tropical disturbance to a warm-core cyclonic system. Except for the southeastern Pacific and the southeastern Atlantic where cold SST tongues (and excessively large vertical shears) are pronounced, TC genesis is observed over all other tropical ocean basins. A statistical TC distribution map (Gray 1968) showed that the western North Pacific (WNP) experiences the most frequent TC activity among eight TC genesis regions. The average annual number of TCs in the WNP is 22, which contributes 36% of the global total TC numbers. Both the western North Atlantic and South Pacific have an average of seven TCs, while the Bay of Bengal, South Indian Ocean and eastern North Pacific have six TCs. The Arabian Sea and the sea off the northwest Australian coast have an average of three cyclogenesis events each year. To explain the geographic distribution of TC genesis, Gray (1975, 1977) investigated the environmental conditions around the cyclogenesis regions and found that six physical parameters can be used to evaluate the potential tendency for TC genesis. Based on these parameters, the favorable conditions for TC genesis include positive relative vorticity in lower troposphere, a location away from the equator, a warm SST above 26.1°C, a small vertical shear, a large equivalent potential temperature (θ_e) gradient between 500 hPa and the surface, and a relatively large value of relative humidity in the middle troposphere.

Ritchie and Holland (1999) documented environmental flow regimes that favor TC genesis in the WNP. They are the monsoon shear line, the monsoon confluence region, and the monsoon gyre. While these favorable large-scale flow patterns are important in providing background stages for tropical storm development, they are not sufficient to lead to cyclogenesis. Although these patterns are often seen in daily weather maps, cyclogenesis does not occur every day. This implies that...
there must be some triggering mechanisms for cyclogenesis. Previous studies pointed out various synoptic perturbation triggering scenarios such as the Rossby wave energy dispersion of a preexisting TC (Frank 1982; Davidson and Hendon 1989; Ritchie and Holland 1997; Briegel and Frank 1997), energy accumulation of easterly waves in a confluent mean zonal flow (Kuo et al. 2001), and a mixed Rossby–gravity wave packet (Dickinson and Molinari 2002). Tropical cyclone genesis may also involve wave–mean flow interactions (Ferreira and Schubert 1997; Zehrner et al. 1999; Molinari et al. 2000) or development of synoptic-scale wave trains (Lau and Lau 1990; Chang et al. 1996). Sobel and Bretherton (1999) analyzed the energy flux convergence in the WNP and found that wave accumulation may operate either on waves coming from outside of the region or in situ.

In this study we particularly focus on the scenario related to TC energy dispersion (TCED). As we know, a mature TC is subject to Rossby wave energy dispersion because of the change of the Coriolis force with latitude (e.g., Anthes 1982; Flierl 1984; Luo 1994; McDonald 1998). While a TC moves northwestward because of mean flow steering and the beta drift, it emits Rossby wave energy southeastward, forming a synoptic-scale wave train with alternating anticyclonic and cyclonic vorticity perturbations (Carr and Elsberry 1994, 1995). Using barotropic vorticity equation models, Chan and Williams (1987) and Fiorino and Elsberry (1989) showed that a monopole, cyclonic vortex on a beta plane experiences Rossby wave energy dispersion that generates asymmetric structure with a cyclonic gyre to the west and an anticyclonic gyre to the east of the vortex center. While the motion of the vortex associated with the beta drift is toward the northwest, the vortex is subject to energy dispersion due to Rossby wave radiation (Flierl 1984). Luo (1994) focused on the effect of Rossby energy dispersion on the motion and structure of a TC-scale vortex on a beta plane in a nondivergent barotropic model. A wavelike wake is formed in his model as a result of Rossby energy dispersion from the vortex. Because of the leaking of energy, the vortex weakens during its northwestward movement. Carr and Elsberry (1995) found the existence of TCED-induced Rossby wave trains in both a barotropic model and Naval Operational Global Atmospheric Prediction System (NOGAPS) forecast fields. Holland (1995) discussed how sequential TCs might be generated because of TCED in the WNP. Once a TC develops in the vicinity of the monsoon trough/ITCZ, it moves poleward and westward because of the planetary vorticity gradient and mean flow steering; group energy associated with the TC propagates eastward and equatorward, forming a synoptic-scale wave train of alternating regions of anticyclonic and cyclonic vorticity perturbations. The perturbation energy, being accumulated in the confluenze zone, helps release the barotropic/baroclinic instability in the ITCZ. Cumulus convection is organized in the region of cyclonic circulation, amplifying the disturbances through the convection–frictional convergence feedback (Wang and Li 1994). As the process proceeds, a TC-strength vortex may be generated. Once the newly generated vortex moves away from the region, it may trigger the new development of disturbances through the same energy dispersion/accumulation processes.

The role of TCED on cyclogenesis was previously suggested based on either simple numerical models or indirect observational data (e.g., global model analysis products) over open oceans. So far, no direct observational evidence shows the detailed structure and evolution characteristics of the Rossby wave train induced by TCED, nor its relationship with TC genesis. Recent satellite products [such as the Quick Scatterometer (QuikSCAT)] provide a great opportunity to reveal the detailed pattern and evolution of the Rossby wave train and associated cyclogenesis processes (e.g., Li et al. 2003). The objective of this paper is to document from direct satellite measurements the detailed structure and evolution characteristics of the Rossby wave train in association with TCED and its role in cyclogenesis. While Part I is focused on the observed aspects, Li et al. (2006, hereafter Part II) will simulate the TCED-induced cyclogenesis in a 3D model.

Part I is organized as follows. In section 2 we briefly describe the data and methodology. In section 3 we present the observed structure and evolution characteristics of the Rossby wave train. Then in section 4 we examine possible factors that control the generation of the Rossby wave train. The large-scale dynamic and thermodynamic control on TCED-induced cyclogenesis is discussed in section 5. Finally, conclusions are given in section 6.

2. Data and methodology

The primary data used in this study are the National Aeronautics and Space Administration (NASA) QuikSCAT level 3 daily surface wind data and the Tropical Rainfall Measurement Mission (TRMM) Microwave Imager (TMI) cloud liquid water data. The QuikSCAT data are retrieved by a microwave scatterometer named SeaWinds that was launched on the QuikBird satellite in 1999. The primary mission of QuikSCAT is to measure near-surface winds. By detecting the sea surface roughness, it can use some empirical relationship to calculate the 10-m surface wind
speed and direction. The products from this scatterometer include 10-m surface zonal and meridional wind speed daily data. It covers global oceans with a resolution of $0.25^\circ \times 0.25^\circ$. The TMI data are retrieved by a radiometer on board the TRMM satellite. It is a joint mission supported by NASA and the National Space Development Agency of Japan. The TMI data cover the tropical region between $40^\circ$S--$40^\circ$N and provide 10-m surface wind speed (11 and 37 GHz), sea surface temperature, atmospheric water vapor, cloud liquid water, and precipitation rates with a horizontal resolution of $0.25^\circ \times 0.25^\circ$.

In addition, National Centers for Environmental Prediction–National Center of Atmosphere Research (NCEP–NCAR) reanalysis data are used. Having a horizontal resolution of $2.5^\circ \times 2.5^\circ$, the reanalysis data are used to fill in gaps in satellite observations with a linear optimal interpolation scheme. The reanalysis data are also used to construct the vertical structure of the wave train in section 3.

We focus on the cyclogenesis events associated with TCED in the WNP during the 2000–01 typhoon seasons (from 1 June to 30 September). The TC best-track data issued by Joint Typhoon Warning Center (JTWC) are used to determine the cyclogenesis location. To clearly examine synoptic-scale wind structures prior to TC formation, a time filtering method is used to retain 3–8-day signals for both QuikSCAT and TMI data. The selection of the 3–8-day bandwidth is based on Lau and Lau (1990, 1992), who noted a significant synoptic-scale wave spectrum appearing in this frequency band in the WNP. The filtering scheme is that of Murakami (1979). The response function for this filter is a Gaussian function. In the 3–8-day band, the maximum response is at a period of about 5 days. A low-pass filter (with a period greater than 20 days) was used to compose large-scale background conditions associated with cyclogenesis and nuncyclogenesis cases in section 5.

The TCED-induced cyclogenesis events are identified based on the following analysis approach. The daily synoptic maps of the filtered surface wind fields from the QuikSCAT data are first examined. If a new TC formed in the cyclonic circulation embedded in a wave train produced by a preexisting TC (based on JTWC best-track data), then this new TC belongs to the TCED-induced cyclogenesis scenario. Note that this analysis approach neglects possible triggering processes from the upper troposphere.

3. Rossby wave train structure

During the summers of 2000–01, we analyzed 34 cyclogenesis cases in the WNP in which a tropical perturbation finally reached or exceeded tropical storm intensity. Among the 34 cases, 6 cyclogenesis cases are identified to be associated with the energy dispersion of a preexisting TC. Figure 1 illustrates such a case. It shows the evolution of the 3–8-day filtered QuikSCAT wind fields from 1 to 9 August 2000. The letter “A” in Fig. 1 represents the location of Typhoon Jelawat that formed on 1 August. During the first 2–3 days, because of its weak intensity, Jelawat did not generate a visible Rossby wave train in its wake. During its northwestward journey, the intensity of Jelawat increased. With its steady intensification, the Rossby wave train became more and more evident in its wake. On 6 August, a clear wave train was observed. This Rossby wave train has a zonal wavelength of about 2500 km, oriented in a northwest–southwest direction.

The most remarkable characteristic of this wave train is that it has a large meridional wavelength. A half meridional wavelength is about 4000 km on 6 August. The meridional wavelength of the wave train is greatly reduced in subsequent days, leading to generation of a new TC named Ewiniar on 9 August. Note that TC Ewiniar (represented by the letter “B” in Fig. 1) formed in the cyclonic vorticity center of the Rossby wave train.

The cloud liquid water fields show a similar evolution feature (Fig. 2). During the first 2–3 days, convective clouds (represented by the cloud liquid water field) were loosely organized. On 3 August, there was a northwest–southeast-oriented cloud band in the wake of Jelawat. On 6 August, when the wave train pattern was discerned in the wind field, convective clouds also started to be organized in such a way that the cloud liquid water concentrated in cyclonic circulation regions. With the further development of the wave train, one can see clearly the alternating cloudy–clear–cloudy pattern, corresponding well to the alternating cyclonic–anticyclonic–cyclonic circulation. Thus, the independent satellite products provide consistent wind and cloud patterns during the course of the wave train development.

To clearly demonstrate the role of the Rossby wave energy dispersion, we calculated an $\mathbf{E}$ vector (an energy propagation vector) for TC Jelawat (see Fig. 3), based on Trenberth (1986). Here, $\mathbf{E} = ([−u′u′ + v′v′]/2, [−u′v′])$, where $[\ ]$ represents time averaging, and $u′$ and $v′$ are zonal and meridional components of synoptic-scale wind perturbations, respectively. The $\mathbf{E}$ vector in Fig. 3 was calculated based on a 15-day period centered on 5 August 2000. It indicates the direction of energy dispersion around 5 August, assuming that the wave train is a Rossby wave type. Figure 3 illustrates that there is indeed southeastward propagation of...
Rossby wave energy. The $E$ vector is not sensitive to the averaged period, say, from 9 to 21 days.

The second example of the TCED scenario is the genesis of Typhoon Prapiroon on 26 August 2000. Figure 4 shows the evolution of Rossby wave trains associated with a preexisting Typhoon Bilis, which formed on 18 August. On 22 August, Bilis was located to the northeast of the Philippines. One day later, it made landfall on the southeastern coast of China and rapidly dissipated within 48 h. Nevertheless, its wave train still
existed. On 26 August, a new TC (Prapiroon) formed in the cyclonic vorticity region of the wave train.

The third example is the formation of Typhoon Utor. Figure 5 illustrates the time sequence of surface wind patterns associated with a preexisting Typhoon Durian that formed over the South China Sea on 29 June 2001. A clear Rossby wave train was discerned in the wake of Durian on 30 June. This wave train has a zonal wavelength of 3000 km, tilting toward the east-southeast. On 1 July, Durian made landfall on China, near Hong Kong. In the meantime, a new TC (Utor) formed in the cyclonic vorticity center of the wave train.

Fig. 2. Time sequences of the cloud liquid water field (mm) associated with the Rossby wave energy dispersion of Typhoon Jelawat. As in Fig. 1, “A” represents center location of Jelawat and “B” represents center location of Ewiniar.
To make sure that the wave trains derived above are not an artifact of the filtering, we subtracted the original QuikSCAT wind field from a 31-day sliding-window average (from 15 days prior to 15 days after). The results (figures not shown) show that the wave trains are clearly present even with the raw data. Thus, the satellite (QuikSCAT and TMI) measurements provide the direct observational evidence of the TCED-induced Rossby wave train, its structure and evolution characteristics, and its role in TC genesis. This confirms the previous TCED-induced cyclogenesis hypothesis based on either limited observational data (e.g., Frank 1982) or numerical model experiments (e.g., Holland 1995).

To reveal the vertical structure of the TCED-induced Rossby wave train, we examine the NCEP–NCAR reanalysis fields. The same synoptic-scale filtering technique was applied to the reanalysis data. By comparing the QuikSCAT-derived wave train patterns with those from the reanalysis data, we noted that the reanalysis data in general underestimate the strength of the wave train; in particular, it completely misses the wave train pattern induced by Typhoon Saomai (2000). However, it captures the wave train pattern of Typhoon Jelawat (2000) reasonably well. Thus, in the following we examine the vertical structure of the Rossby wave train for Jelawat. Figure 6 shows that the northwest–southeast-oriented wave trains are clearly evident from the surface to 500 hPa, and become less organized in the upper troposphere (200 hPa).

Figure 7 illustrates the vertical cross section of the Rossby wave train along a northwest–southeast-oriented axis (i.e., the solid black line at the bottom panel of Fig. 6). From Fig. 7, one can see that the TCED-induced Rossby wave train penetrates through the entire tropospheric layer. While the divergence in the wake region exhibits a dominant first baroclinic mode structure, with maximum divergence or convergence centers located at 150 hPa or the surface, the vorticity has an equivalent barotropic structure. In the cyclonic vorticity center of the wave train, the maximum vorticity perturbation appears at 850 hPa. The zonal and meridional wind perturbations are collocated and have the same sign, implying a northwest–southeast orientation for the wave train.

4. Factors that influence the wave train generation

The examination of all the 34 TC cases reveals that not all TCs have a Rossby wave train in their wake. This poses an interesting question: What determines the generation of the Rossby wave train? Previous theoretical and modeling studies suggested that TCED depends greatly on the size of a TC (Carr and Elsberry 1995) and its wind profile (e.g., Flierl et al. 1983). A TC with zero total relative angular momentum tends to conserve its energy (Shapiro and Ooyama 1990).

Because of a lack of reliable observational information on TC size and wind profiles, we examine relationships between the occurrence of the Rossby wave train and TC intensity, based on the fact that many intense TCs in the WNP are often large ones. For all the 34 TCs, we have a total of 233 (snapshot) samples. We separate the 233 samples into three groups based on TC minimum sea level pressure (MSLP). They represent strong (MSLP > 960 hPa), moderate (960 hPa ≥ MSLP ≥ 980 hPa), and weak (MSLP > 980 hPa) TC intensities. Figure 8 shows the percentages of occurrence of the Rossby wave train at each group. For the strong TCs, the percentage is quite high (about 87%). It drops quickly to 40% (30%) for moderate (weak) storms. The result above shows a great sensitivity of the occurrence of the Rossby wave train to the TC intensity.

In addition, the occurrence of the Rossby wave train may also depend on the mean background flow. Figure 9 illustrates the wave train generation and nongeneration cases for the strong TC group (i.e., MSLP < 960 hPa).
Here closed circles (squares) in the figure represent the cases with (without) the appearance of the wave train. For strong TCs, the southeast–northwest-oriented Rossby wave trains were observed west of 155°E. They are absent to the east of this longitude where the mean easterly trades are large (~4 m s⁻¹ or stronger).

Physical mechanisms that give rise to this mean flow dependence is unclear. It is speculated that it may result from the frequency modulation by the mean flow, as the strong easterly trades may prevent the southeastward propagation of the perturbation kinetic energy emitted from a TC.

5. Large-scale dynamic and thermodynamic control on cyclogenesis

Our synoptic analysis of the QuikSCAT wind field shows that for 15 TCs that generated a Rossby wave train in their wake, only 6 of them eventually lead to new TC formation. The fact that some of the Rossby wave trains lead to cyclogenesis but others do not raises an important question: What determines the cyclogenesis threshold in the Rossby wave train? In the following, we intend to address this issue from large-scale dynamics and thermodynamic points of view.

We hypothesize that the background flow and moisture conditions are critical for the wave train development. To identify the difference of the background flow between the cyclogenesis and noncyclogenesis cases, we conducted a composite analysis. Six genesis cases and nine nongenesis cases are composed. Figure 10 shows the comparisons of the composite large-scale background variables for the genesis and nongenesis cases. These variables are averaged in a 5° × 5° box centered at the cyclonic vorticity center of the wave train, and have been subjected to a low-pass filter to retain signals longer than 20 days so that they reasonably represent the large-scale environmental conditions including the seasonal cycle and intraseasonal oscillation (ISO) activities. Note that there are clear differences between the genesis and nongenesis cases in large-scale background conditions. For instance, the vertical motion profiles have a significant difference in
the midtroposphere. A stronger upward motion appears in the genesis cases. Consistent with the vertical motion, stronger low-level convergence, upper-level divergence, and low-level cyclonic vorticity occur in the cyclogenesis cases. Vertical wind shear is also relatively smaller and the relative humidity is larger throughout the troposphere in the cyclogenesis cases. Thus, all environmental variables exhibit a favorable condition in the cyclogenesis cases.

Figure 11a shows the horizontal patterns of the composite large-scale background wind fields at 500 and 850 hPa associated with the genesis and nongenesis cases. For the cyclogenesis cases, a background cyclonic wind shear appears at both 850 and 500 hPa. However, for the nongenesis cases, an anticyclonic shear appears in lower troposphere (850 hPa), while rather uniform winds cross the center at 500 hPa. These wind features are well reflected in the vorticity composites (see Fig. 11b), which exhibit a positive vorticity near the composite center at both 850 and 500 hPa in the cyclogenesis cases, compared to a near-zero or negative vorticity in the noncyclogenesis cases. The synoptic-scale wave train center nearly collocates with the background vorticity center in the genesis cases.

In addition to the large-scale flow pattern, back-

Fig. 5. Time sequences of synoptic-scale surface wind patterns associated with the Rossby wave energy dispersion of Typhoon Durian; “A” represents the center location of Durian that formed on 29 Jun 2001 and “B” represents the center location of a new TC named Utor that formed on 1 Jul 2001 in the wake of the Rossby wave train of Durian.

Fig. 6. Rossby wave train patterns at 200, 500, 850, and 1000 hPa produced by TCED from Typhoon Jelawat on 8 Aug 2000. The wind fields are 3–8-day filtered. The thick line in the bottom panel represents the axis of the wave train.
Ground thermodynamic conditions such as humidity distribution and moist static stability in lower troposphere may also play a role. It is noted that the specific humidity field has a more favorable large-scale background in the genesis cases, that is, large humidity centers appear near the cyclonic vorticity centers of the wave trains at both 850 and 500 hPa.

Figure 12 illustrates the thermodynamic control in the formation of TC Prapiroon. Both Prapiroon and the previous Typhoon Bilis formed in the same area (within 10° longitude distance). Strong winds and dense clouds associated with Bilis cooled the ocean surface through enhanced upwelling/evaporation and reduction of shortwave radiation. As a result, the local ocean surface became relatively cooler compared to its surroundings, even a few days after Bilis had moved away from the region. This led to relatively stable stratification in the region, which did not favor cyclogenesis even though the wave train had developed. Then, the atmospheric conditional instability started to build up gradually, which can be clearly seen from the increase of both surface moisture and air temperature at 2 m from day −4 to day 0 in Fig. 12 (here day 0 refers to the genesis date for Typhoon Prapiroon). During this building-up period, surface air temperature increased by 1°C, while the surface specific humidity increased by 2 g kg⁻¹. The increase of both the temperature and moisture led to the increase of the surface saturation.

Figure 7. Vertical cross section of (a) divergence (10⁻⁶ s⁻¹), (b) vorticity (10⁻⁵ s⁻¹), (c) zonal wind (m s⁻¹), and (d) meridional wind (m s⁻¹) along the Rossby wave train axis defined in Fig. 6. The y axis denotes pressure (hPa). The open and closed circles represent the relative position of cyclonic and anticyclonic centers of the wave train.

Figure 8. Percentage of occurrence of TCED-induced Rossby wave train for strong, moderate, and weak TCs.

Figure 9. Geographic location of the strong TCs (MSLP < 960 hPa) that do (closed circle) and do not (closed square) have a Rossby wave train in the wake. Superposed are background June–September-averaged wind vectors at 1000 hPa. The region where the mean easterly wind speed is greater than 4 m s⁻¹ is contoured and shaded.
equilibrium potential temperature. As a result, $\partial \theta_p/\partial z$ decreased in the lower troposphere (between 1000 and 850 hPa), leading to more unstable atmospheric stratification.

The slow thermodynamic building-up process might be responsible for the observed frequency of TC genesis in the WNP. In an observational study, Frank (1982) found that the typical separation distance and time period between the two subsequent TCs in the WNP were around 2000 km and 7–10 days. For the Typhoon Prapiroon case, it is likely that the timing of cyclogenesis was determined by both dynamic factors such as the strength of the wave train and background flow conditions and the thermodynamic accumulation of the surface moist static energy. Prapiroon formed in the same region as Bilis did 8 days prior. Strong winds and thick cloud cover associated with Bilis reduced the SST and the surface moist static energy. It required a "recovery" period to rebuild conditional unstable stratification in the region.

6. Conclusions

The structure and evolution characteristics of Rossby wave trains induced by TC energy dispersion are revealed with direct satellite (QuikSCAT and TMI) measurements. An $E$ vector calculation shows that the wave train is indeed associated with the TC Rossby wave energy dispersion. For 34 TCs analyzed in the WNP during 2000–01 typhoon seasons, there are 6 cyclogenesis cases that are associated with the Rossby wave energy dispersion of a preexisting TC. In all these cases, a new TC formed in the cyclonic vorticity region of a preexisting well-defined wave train. It seems that there is a scale (particularly meridional scale) contraction process associated with cyclogenesis. The cause of this
scale contraction is unknown at the moment, and it is speculated that it may result from the wave train–mean flow interaction and energy accumulation in a convergent mean background flow.

Our observational analyses indicate that not all TCs have Rossby wave trains in their wakes. The occurrence of the wave train seems dependent on TC intensity and the background mean flow. For stronger TCs, there are more chances to form a Rossby wave train in their wakes.

We also note that not all TCs that have a Rossby wave train finally lead to new TC formation. Whether or not the Rossby wave train results in cyclogenesis depends on background dynamic and thermodynamic conditions that are related to both the seasonal cycle and ISO activities. The stronger low-level convergence
and cyclonic vorticity, weaker vertical shear, and greater midtropospheric moisture are among the favorable background conditions for cyclogenesis. The rebuilding process of a conditional unstable stratification is important in regulating the frequency of TC genesis.

In Part I we focus on the observational aspects of the Rossby wave train. In reality, the TC energy dispersion might not be a sole factor, and other processes such as easterly waves (e.g., Li et al. 2003) and upper tropospheric forcing may also play a role. Thus it is necessary to conduct a composite analysis to obtain common features associated with the Rossby wave train scenario. In Part II, we will simulate the process associated with TCED-induced cyclogenesis in a 3D model to reveal fundamental dynamics that give rise to the TC genesis.

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