Initiation and Intensification of Tropical Depressions over the Southern Indian Ocean: Influence of the MJO

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ABSTRACT

Using ERA-Interim global atmospheric reanalysis, an original tracking approach is developed to follow tropical low pressure systems from the early tropical depression (TD) stage up to possible intensification into developed tropical cyclones (TCs). The different TC stages are identified using the IBTrACS dataset. This approach detects many more TD initiations compared to IBTrACS alone and thus gives a more comprehensive dataset to study the cyclogenesis by considering separately TD initiations and the probability of intensification.

In the south Indian Ocean (SIO), the MJO modulation of the number of TCs is primarily due to the modulation of the number of TD initiations and secondarily to the probability of their intensification. The TD initiations are more probable at 55°S, 75°E and 95°E and can be primarily attributed to the development of an unstable cyclonic meridional shear of the zonal wind at low levels. The reinforcement of this shear results from (i) a heat low, related to a precipitation anomaly, which triggers westerly winds equatorward of the initiation region and (ii) an easterly wind strengthening south of the initiation regions due either to a reinforcement of the subtropical high (for western and central SIO) or to a large-scale depression over the western Maritime Continent (for eastern SIO). Over the western and central SIO, the concomitance of precipitation and subtropical high anomalies at the origin of the shear reinforcement could be partly stochastic, giving a weaker relation with MJO and ENSO. Over the eastern SIO, the large-scale MJO (and ENSO) perturbation pattern alone can reinforce the shear, giving a larger modulation of the number of TD initiations.

1. Introduction

The convectively active phase of a MJO event is a large-scale region covered by numerous synoptic-scale perturbations and convective systems of different sizes and durations. There are many examples showing how meso- to synoptic-scale convective systems may be organized in the active MJO phase (e.g., Nakazawa 1988; Chen et al. 1996; Katsumata et al. 2009). In particular, convective systems may be organized in large clusters possibly related to equatorial waves with periods of a few days. Many studies also show the more specific relation between the MJO and tropical cyclones (TCs) in different parts of the globe (e.g., Liebmann et al. 1994; Maloney and Hartmann 2000a,b; Hall et al. 2001; Bessafi and Wheeler 2006; Ho et al. 2006; Camargo et al. 2009; Ramsay et al. 2012; Klotzbach 2014). The modulation of the cyclogenesis by the MJO is an important issue for forecasting the TC activity during a cyclone season. It is also important for understanding the physical source of the cyclogenesis and possibly the physical source of the MJO itself since a TC will also impact the large-scale dynamics in the active MJO phase. The present study focuses on the relation between the MJO and the initiation of tropical depressions (TDs) in the south Indian Ocean and on the chance for these TDs to reach the tropical storm (TS) stage (10-min sustained wind larger than 17 m s$^{-1}$).

Previous studies show that the modulation of the cyclical activity by the MJO may be attributed mostly to large-scale perturbations of the vorticity and/or of the vertical and horizontal wind shear. An interesting result reported in Liebmann et al. (1994) is that MJO perturbations do not specifically modulate the intensification process (i.e., the chance for a TD to reach the TS strength), but rather the number of TDs, with more initiation in the convective phase of the MJO. The ratio

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of TS to TD initiations was found to be around 0.8 for both active and suppressed MJO phases. This fact is not specific to the MJO since Liebmann et al. (1994) show that the relation between a large-scale convective perturbation and the ratio of TS to TD initiations is similar for another shorter time frequency domain (i.e., 18–27-day band instead of the 35–95-day band for the MJO). The relation between the MJO and the maximum TC intensity is also not well established. Such a tendency is noted in Maloney and Hartmann (2000a) for the Gulf of Mexico and Klotzbach (2014) shows that the likelihood of rapid intensification increases during the active MJO phase. However, Hall et al. (2001) show no significant relation between the MJO and the TC intensity in the Australian region.

The modification of the background state (dynamic and thermodynamic) in an active MJO phase could thus mostly promote initiation of more TDs. In this case, the analysis of the MJO modulation of developed TCs (i.e., depressions reaching the TS stage or above) should focus more on TD initiations than on the intensification process. It is likely that the MJO modulation of the background flow has a major role in the initiation of more numerous TDs. One can find an analogy with the background African easterly jet (AEJ) that produces African easterly waves (AEW) that later generate TDs in the North Atlantic (e.g., Landsea 1993; Dunkerton et al. 2009). The perturbation of the background flow by the MJO, and especially the westerly low-level jet developing in and after the active phase, may thus be conducive for the generation of more numerous TDs. These TDs will move mostly westward and poleward due to the β drift and encounter an environment (dynamic and thermodynamic) typical of the western side of the most active MJO phase. This environment, corresponding to decreasing large-scale convective instability, is not necessarily favorable to TD intensification compared to other MJO phases or to “neutral” MJO conditions.

Westward-propagating disturbances like easterly waves over the Atlantic Ocean and West Africa (Carlson 1969; Burpee 1972, 1975; Reed et al. 1977), or mixed Rossby–gravity waves (MRG) and TD-type disturbances (similar to easterly waves) over the North Pacific (e.g., Reed and Recker 1971; Lau and Lau 1990; Takayabu and Nitta 1993) are known sources of tropical depressions. The origin of these waves can be somewhat attributed to instabilities due to a meridional shear of the low-level zonal wind with reversal in the meridional PV gradient. Two examples are (i) the shear instability of the AEJ (e.g., Thornicroft and Hoskins 1994) and (ii) the “ITCZ breakdown” related to a meridional cyclonic shear poleward of the ITCZ (e.g., Ferreira and Schubert 1997). Other mechanisms such as the “wave accumulation” process (e.g., Sobel and Bretherton 1999; Maloney and Hartmann 2001) can be invoked for the initiation of tropical depressions. This mechanism is a reduction of the wavelength and an increase of amplitude of an existing wave due to the latitudinal stretch of a zonal jet. This may explain the transformation from MRG into easterly “TD type” wave disturbances as they propagate westward in the central Pacific. The wave accumulation process requests that an easterly wave is generated farther east. Sobel and Bretherton (1999) also noted that part of this process may operate on waves generated in situ in the vicinity of mature tropical cyclones. In fact, these TD-type wave disturbances are propagating northwestward over the northwestern Pacific in a trajectory very close to the trajectory of tropical cyclones. The energy budgets showing that the maintenance of these waves is mostly due to diabatic heating associated with convection (e.g., Lau and Lau 1990, 1992) could thus be partly distorted by the presence of tropical cyclones in the wave signature (Sobel and Bretherton 1999).

The relation between synoptic waves, TCs, and the MJO received more attention for the North Pacific basin compared to other ocean basins. Over the eastern part of the North Pacific basin, Maloney and Hartmann (2000b) found that TCs are 4 times more numerous during the MJO westerly phase of the low-level zonal wind compared to the MJO easterly phase. This was attributed to enhanced cyclonic meridional shear of the low-level zonal wind and to a weak vertical zonal shear favoring the cyclogenesis. Aiyyer and Molinari (2008) note, however, that for the eastern Pacific the vertical shear is in fact stronger in the MJO westerly phase if one considers the total shear instead the zonal shear only. This result is also compatible with the ITCZ breakdown process mentioned above that predicts initiation of tropical depressions by a meridional shear of the low-level zonal wind. For the western part of the North Pacific basin, Maloney and Hartmann (2001) found that the MJO westerly phase is also favoring the generation and the growth of tropical depressions by barotropic energy conversion from the low-level mean flow, together with the wave accumulation process. They also noticed that the westerly phase is associated with strong low-level convergence and high SSTs that can promote the intensification of these tropical depressions. The MJO modulation of the cyclogenesis over a given region can also be related to a shift of the mean state. For example Aiyyer and Molinari (2008) show that the number of cyclones over the Gulf of Mexico is augmented in a particular MJO phase because of a latitudinal shift of the easterly wave train by
the MJO large-scale dynamical perturbation. It is also worth mentioning the particular case of twin cyclones (Nitta 1989) that straddle the equator in the Indian and Pacific Oceans. Numerical simulations show that these twin cyclones develop preferentially when the equatorial mass sink associated to the MJO is stationary (Ferreira et al. 1996). These results suggest that the augmentation of the TC activity in particular MJO phases could be related primarily to favorable conditions for the initiation of more TDs. As mentioned above, one may thus consider that large-scale cyclonic vorticity perturbation related to the MJO does not promote the intensification of existing TDs. Instead, the cyclonic shear of the zonal low-level jet generated by the MJO could be unstable and multiply the number of TD initiations.

During the Cooperative Indian Ocean Experiment on Intraseasonal Variability in the Year 2011 (CINDY2011)–DYNAMO experiment (Yoneyama et al. 2013), many tropical depressions were observed during and after actives MJO phases, and only some of them intensified into developed TCs. The early stage of these TDs was well depicted in meteorological analyses of operational centers like Météo-France or ECMWF as well as in the ERA-Interim (ERA-I) global atmospheric reanalysis dataset (Dee et al. 2011). The purpose of the present paper is to explore the relation between the MJO and TD initiations in the south Indian Ocean on the basis of more than 30 years of ERA-I fields combined with the International Best Track Archive for Climate Stewardship (IBTrACS) dataset (Knapp et al. 2010) on tropical cyclone observations. Using ERA-I makes it possible to construct an alternative climatology of TD initiations using an objective approach. The aim is to detect more TD initiations at an earlier stage compared to those reported in IBTrACS on the basis of information given by Regional Specialized Meteorological Centers (RSMC). In IBTrACS, there are indeed only a few TDs that do not reach the TS strength; it is thus difficult to study exhaustively the physical environment of TD initiations and the probability of their intensification into TSs. The RSMC tropical cyclone dataset is used here only to identify those TDs that reach the TS intensity. In this paper, TDs will refer to low pressure systems preceding intensification into TSs and then possibly into more developed TCs. Here TDs will also refer to those low pressure systems that never reach the TS strength. A TD is defined as a negative anomaly (with some empirical thresholds) associated with a local minimum in the geopotential height at 850 hPa (this is nearly equivalent to a closed circulation for a geostrophic circulation, i.e., a minimum in the streamfunction).

Section 2 presents the datasets and the analysis methods. Section 3 is devoted to statistics on TD initiations and intensification characteristics as a function of latitude, year, and MJO phase. Section 4 details the possible physical sources of TD initiations for different regions by looking at the corresponding atmospheric state. The results are summarized and discussed in section 5.

2. Datasets and analysis

a. Datasets

The detection of the tropical low pressure systems is done using ERA-I (Dee et al. 2011) between 1979 and 2012. This reanalysis dataset has a horizontal resolution of $0.75^\circ \times 0.75^\circ$ and a 6-h time step. Tropical cyclone positions and intensities are retrieved from the IBTrACS dataset (Knapp et al. 2010) between 1979 and 2012. We use the data from the World Meteorological Organization (WMO) RSMCs that correspond to La Réunion (Météo-France) and Darwin (Australian Bureau of Meteorology). The Global Precipitation Climatology Project (GPCP) at 1° daily resolution (Huffman et al. 2001) is used to examine the precipitation field close to TD initiations. We also use the OLR dataset (Liebmann and Smith 1996) to define the MJO phases.

b. Tracking of the low pressure systems in ERA-I

The local depression associated with the passage of a tropical low pressure system is associated with a synoptic perturbation of both vorticity and geopotential in the lower troposphere. The vorticity field is noisier compared to the geopotential height and is used here to characterize the low pressure system but not to define its position and its horizontal extension. The tracking approach used in the present study is based on the overlap between the “depression areas” for two successive time steps [see also Satake et al. (2013)]. The depression area is defined by a closed contour in a geopotential height $(\phi)$ anomaly at 850 hPa. A geopotential height anomaly is used, instead of just the geopotential height, since the geopotential varies at the large scale at many time frequencies from semidiurnal to seasonal. A simple threshold in $\phi$ is thus not appropriate. To retain only perturbations related to synoptic-scale depressions, the anomaly field $\Delta \phi$ is constructed as the difference between the gridpoint value and the average over a region of $\pm 7.5^\circ$ (i.e., $\pm 10$ grid points) around this grid point. The depression area is then defined as continuous grid points with $\Delta \phi$ lower than a given threshold.

A critical point is the choice of the criteria for the selection of the low pressure systems. The aim is to
detect TCs for their whole life cycle (TD, TS, developed TC), including in particular TDs at an early stage in order to analyze the physical environment just before and during their initiation (section 4). The aim is also to detect those TDs that never reach the TS stage. Obviously, depending on these criteria, the number of systems detected, their strength, and duration will vary. The most critical criterion is the geopotential anomaly threshold that fixes the detectability of a low pressure system. This threshold has been set empirically in order to identify the early stage of all depressions appearing during the CINDY–DYNAMO campaign period (October–December 2011). After a series of tests, an empirical threshold $\Delta \phi_0 = -80 \text{m}^2\text{s}^{-2}$ was set as the minimal (in absolute value) geopotential perturbation. This threshold is sufficiently small to allow early TD detection. The depression area is defined as the area encompassed in a closed isoline at $\Delta \phi_0$. For a developed TC, however, a contour at this threshold may encompass a very large area with an ill-defined center and ill-defined characteristics. A maximum depression area of 50 grid points (equivalent radius of 3°) is thus imposed and the threshold is adjusted accordingly. For deep depressions the adjusted threshold can be as large as $-700 \text{m}^2\text{s}^{-2}$. Interestingly, the minimal threshold $\Delta \phi_0 = -80 \text{m}^2\text{s}^{-2}$ gives the larger difference between the TD initiation probability in phase 4 and 1 of the MJO (see analysis of Fig. 8 below) compared to estimates done by the RSMC at La Réunion. There is not a perfect correspondence between the positions given by IBTrACS and ERA-I, notably around 22°S, 72°E. The distribution of distances between ERA-I and IBTrACS systems in the south Indian Ocean is given in Fig. 2. There is a clear improvement in the adequacy between the two datasets between the first and the last decade. The most probable distance is around 1° for the first decade and 0.5° for the last decade. The average distance tends also to decrease regularly with time, from 1.5° in the 1980s to less than 1° in the 2000s.

Both data sources, ERA-I and IBTrACS, have their own strengths and weaknesses and the IBTrACS dataset, based on RSMC reports, is not necessarily as complete (many points have no intensity estimate) or as accurate as expected for the estimate of the intensity of the low pressure systems. The Dvorak technique used to determine this intensity can be basin dependent (Kossin and Velden 2004) and is indeed not yet fully calibrated and validated for the Indian Ocean. Also, because of the evolution of the number and/or the type and quality of satellite measurements (and other data sources) between 1979 and 2012, neither ERA-I nor IBTrACS can determine this intensity can be basin dependent (Kossin and Velden 2004) and is indeed not yet fully calibrated and validated for the Indian Ocean. Also, because of the evolution of the number and/or the type and quality of satellite measurements (and other data sources) between 1979 and 2012, neither ERA-I nor IBTrACS can be considered as perfectly homogeneous datasets. In particular, the lack of geosynchronous satellite information for the Indian Ocean prior to 1998 may have a
large impact on the IBTrACS dataset (Kossin et al. 2007). It is difficult to precisely estimate the impact of these uncertainties, but we will keep them in mind in the interpretation of the results. Here, TD initiations are detected with a more objective approach and in an earlier stage compared to RSMC reports (cf. Fig. 1). The more subjective information given by the RSMC is used here mostly to determine if and when a TD reaches the TS stage (17 m s$^{-1}$ for the 10-min sustained winds or 995 hPa for the minimum surface pressure, depending on the available information).

Over the SIO, a given low pressure system is thus first a TD that could possibly intensify to the TS stage (wind $>34$ kt) before possibly reaching the developed TC stage (wind $>64$ kt). In the following, those TDs that reach at least the TS stage according to IBTrACS are named “cyclogenetic TDs.” The “noncyclogenetic TDs” are thus the TDs detected in ERA-I but that are not referenced or that do not reach the TS strength in IBTrACS. There are thus three types of TD initiation: (i) initiation of all TDs (cyclogenetic or not) detected with ERA-I; (ii) initiation of cyclogenetic TDs detected with ERA-I and that intensify up to the TS stage according to IBTrACS; and (iii) TD initiation using IBTrACS only with some that never reach the TS stage, possibly because there is no reported wind information {because of this ambiguity, the last type [(iii)] is analyzed only on Figs. 3 and 5}. For simplicity, we call “TS initiation” the point where a TD reaches the TS strength for the first time. Basically, the number of cyclogenetic TD initiations must be equal to the number of TS initiations in the SIO area. However, there are small differences due either to TDs initiated in the SIO, but that reach the TS stage outside the SIO, or to TDs initiated outside the SIO that reach the TS stage in the SIO.

![Fig. 1. Successive positions of a low pressure system identified using ERA-I (gray line and circles) during the CINDY–DYNAMO experiment. This low pressure system corresponds to Tropical Cyclone (TC) Benilde reported in the IBTrACS dataset. According to IBTrACS, the TC Benilde reached moderate TS intensity (34 kt) on 29 Dec 2011.](image1)

![Fig. 2. Distribution of distances between the centers of the low pressure system estimated using ERA-I and reported in the IBTrACS dataset for the Indian Ocean region (0°–30°S, 50°–110°E) and for three decades.](image2)
d. The MJO

There is still no consensus on the exact definition of the MJO and on the precise manner to define its amplitude and its phase over a particular region. Here, for simplicity, we consider local intraseasonal events, even if the corresponding planetary pattern is not similar to the winter canonical MJO pattern; in other words, we are only interested in the large-scale organization of the convection at the intraseasonal time scale over the Indian Ocean basin, without concern for the propagation of the perturbation toward the Maritime Continent and the Pacific Ocean. As shown in previous publications (e.g., Bellenger and Duvel 2012), there is in fact no gap between canonical MJO events and other intraseasonal events, but rather a sort of continuum of patterns more or less organized at the planetary scale (i.e., propagation from one ocean basin to another). At the basin scale, the structure of the MJO is quite reproducible from one event to another (Xavier et al. 2008). The physical origin of most intraseasonal events over the Indian Ocean is thus probably similar and their relation with the TD and TC activity will certainly depend more on their local characteristics than on their planetary-scale characteristics.

Here the MJO phase is thus simply given by the filtered OLR signal averaged over the eastern equatorial Indian Ocean (7.5°–7.5°S, 75°–95°E). The OLR signal is filtered using a Kaiser–Bessel filter (20–120 days with an attenuation of 0.5 at 108 and 22 days) and six phases are defined with phase 4 being centered on the most convectively active phase (minimum filtered OLR) and phase 1 on the suppressed phase. To discard periods with weak or no MJO activity, the minimum of the filtered OLR must be smaller than $-10\text{ W m}^{-2}$ and the two surrounding maxima must be larger than $5\text{ W m}^{-2}$, the amplitude (average of the two maxima minus the minimum) of each oscillation must be larger than $30\text{ W m}^{-2}$. With these parameters, there are around 700 days for each MJO phase and there are around 2000 days classified as “no-MJO” for the Indian Ocean for the October–March 1979–2012 period.

e. Low pressure system characteristics and composites

Knowing spatial and temporal coordinates of each selected low pressure system in ERA-I, additional characteristics may be extracted from ERA-I or from other data such as GPCP daily precipitation. Composites are computed to extract average perturbation patterns corresponding, for example, to TD initiations. Distributions of TD or TS initiations for different time periods (MJO categories, years, etc.) are also computed. As in Hall et al. (2001), the statistical significance of these distributions is tested assuming a null hypothesis of a fixed initiation probability $p$ (i.e., the daily probability to obtain a TD initiation over a given region). According to the Moivre–Laplace theorem, the number of initiations $N$ for $n$ realizations (i.e., $n$ days) follows a binomial law with a normal distribution of mean $np$ and standard deviation $\sqrt{np(1-p)}$. The variable, 

$$Z = \frac{N - np}{\sqrt{np(1-p)}}$$

thus follows a standard normal distribution and has a probability of 0.95 (0.99) to be in the interval $[-1.96, 1.96]$ ([$-2.576, 2.576$]). We will use this rule to test if the population of a category can be considered as resulting from a random process with a fixed initiation probability, or if this population may more likely result from a change in the initiation probability, due to, for example, a change in the average mean conditions.


This section reports some statistics on TD initiations for austral summers (October–March) 1979–2012. We consider only TDs initiated inside a large southern Indian Ocean domain (SIO; 0°–30°S, 50°–110°E). The region is limited to the west at 50°E in order to avoid Madagascar and Mozambique channel regions that certainly have particular initiation processes related to orography and to the African continent. There are around 700 TDs, as defined in the previous section, and 240 TSs according to IBTrACS initiated in this SIO domain. Thus, there are around 35% of the TDs (the cyclogenetic TD) that reaches the TS intensity compared to 85% if one considers only the TDs in the IBTrACS dataset. This shows that many TDs with durations longer than 2 days are not identified in the IBTrACS dataset, as expected since RSMC are focusing on developed tropical cyclone activity. This is obviously an important point for the estimate of the chance for a TD to reach the TS intensity and more generally for the study of the potential physical source of tropical cyclones.

a. Geographical distribution

The TD initiations are most frequent around 10°S (Fig. 3). The latitudinal distribution of cyclogenetic TDs is very similar and their ratio with respect to all TDs (not shown) is nearly constant (around 0.4) between 6° and 12°S and smaller north and south of this band. The TS initiation distribution is shifted to the south compared to cyclogenetic TD initiations, as expected because of the
dominant southwestward propagation of the disturbances. The distribution of all TD initiations given by IBTrACS is far weaker and also shifted to the south compared to that given by ERA-I. This confirms that the ERA-I approach generally detects more TDs and at an earlier stage compared to IBTrACS. Distributions of IBTrACS TD and TS initiations are close with only a weak northward extension for the TDs.

It is also possible to inspect the regional distribution of TD initiations in small subregions (1.5° × 3°) sufficiently large to contain a “reasonable” number of initiations. For each subregion \(x\), the TD initiation probability \(p_x\) is defined by \(N_x/n\), where \(N_x\) is the total number of initiations for this subregion (so that \(\sum N_x\) is the total number of initiations for the whole SIO domain), and \(n\) is the total number of considered days. The regional distribution of this probability \(p_x\) (Fig. 4) is quite similar for TDs and cyclogenetic TDs with a larger number of initiations in a band slightly oriented southwest–northeast between roughly 15°S, 50°E and 7.5°S, 100°E. There are three regions with relatively larger TD and cyclogenetic TD initiations around 10°S, 55°E (region A in Fig. 4); 10°S, 75°E (region B); and 7.5°S, 95°E (region C). This could be related to the three “clusters” of TC tracks found by Ramsay et al. (2012), but the link between these three TD initiation regions and the three TC tracks clusters requires further analyses (in particular, one of the cluster encompasses the Madagascar area that is not considered in the present study).

Figure 4c shows the longitudinal distribution of the average number of initiations between the equator and 20°S. This highlights even more clearly the three preferred regions of TD initiation. The regular shape of the longitudinal distribution reveals its statistical nature with three bell-shaped distributions centered at 55°, 75°, and 95°E. In addition, this longitudinal distribution exists considering each decade separately, showing its robustness. The origin of this distribution is discussed in sections 4 and 5. For cyclogenetic TDs (Fig. 4d), two of the three preferred locations are also clearly visible, reinforcing the idea that there is no major distinction between cyclogenetic and noncyclogenetic TDs at the initiation stage. The ratio of these distributions shows, however, that TDs generated east of the domain have more chance to be cyclogenetic (0.55 at 100°E versus around 0.33 west of 90°E). The TS initiation locations (the location where a TD reaches TS strength) given by IBTrACS are more evenly distributed and slightly more likely to the west of the domain. This is related to the variable duration between the TD initiation and its intensification up to the TS strength (Fig. 5b).

b. Duration of the low pressure systems

The duration of the low pressure systems (including possibly different TC strength) is computed for their full life cycle, even if they go outside the SIO domain and possibly until 40°S. The duration distribution for low pressure systems that do not reach the TS strength (Fig. 5a, the gray part of the bars) shows a relatively sharp decrease around 5 days and then a slow decrease until 20 days. For low pressure systems that reach the TS stage, the duration distribution is quite different with a maximum around 13 days. Most systems with duration larger than 13 days are reaching the TS strength, but there is still about half of the systems with duration around 9 days that never reach that strength according to IBTrACS.

The most probable duration of the TD stage for cyclogenetic TDs (Fig. 5b) is around 3 days for ERA-I and 2 days for IBTrACS. The average TD duration for the SIO domain is around 4 days for ERA-I and 2 days for IBTrACS. This duration can be larger than 10 days. Considering a discretization of the MJO in six phases, the duration of one phase is around 6–7 days. Most TDs
Fig. 4. Average daily TD initiation probability (%) of the austral summer season (October–March) for (a) all TDs and (b) cyclogenetic TDs. (c) Longitudinal distribution of the number of TD initiations between the equator and 20°S for the whole period and for each of the three decades. (d) As in (c), but for cyclogenetic TD initiations and TS initiations. Curves in (c) and (d) are smoothed with a binomial (0.25, 0.5, 0.25) filter.
initiated in a given MJO phase will thus reach the TS strength in the following MJO phase.

c. Interannual variations of TD and TS initiations

Interannual variations of the number of TD and TS initiations (Fig. 6) are analyzed with a null hypothesis of a constant initiation probability. The 95% and the 99% significance levels are reported in Fig. 6 according to Eq. (1) with an average probability \( p = 11\% \) for a daily TD initiation (there is an average of 21 TDs initiated each season). The number of TDs is close to or reaches these significance levels for very few years. A potentially interesting point is the very small TD initiation number for austral summer 1997/98, which is characterized by a strong El Niño over the Pacific and a warm SST anomaly over most of the Indian Ocean. This lack of TDs in 1997/98 is not apparent in TS initiations (Fig. 6b). Considering an average probability \( p = 3.9\% \) for the daily TS initiation, only a few years also have a significantly different TS initiation probability, with no apparent relation with the modulation of TD initiations. Because of the small number of systems, especially TS, there is quite a large interannual fluctuation of the ratio TS/TD. Because of these small numbers, this result is also quite sensitive to the definition of TD (criteria defined in section 2b) and does not justify a detailed analysis. The main finding is that most of the interannual variations of the number of TD and TS initiations over the SIO may be attributed to a binomial law with a fixed initiation probability (this does not preclude correlation of the number of TD initiations with large-scale perturbations like ENSO for some part of the SIO, as shown below).

There is no significant long-term trend in the number of TD and TS initiation for the 33 austral summers considered here. For TS initiation, which relies on IBTrACS, the positive trend between 1979 and 1996 is probably due to the lack of intensity reports at the beginning of the series (regular intensity records begin in 1982). The trend is strongly reduced if one considers all IBTrACS systems (i.e., also those with no intensity estimate) instead of TS systems only [see Kossin et al. (2007) for a discussion on this trend]. We may also notice the larger year-to-year TS number variability prior to 1996 and the slightly larger number of ERA-I TDs after this date, but there is no robust and evident explanation for this feature.

Considering the whole region between 50° and 110°E, there is no significant correlation between the seasonal numbers of TD or TS initiations and the seasonal Niño-3.4 index (a simple average of the Niño-3.4 index between October and March). However, considering only the eastern part of the region between 80° and 110°E, this correlation is significant (99% confidence) for both

![Fig. 5. (a) Distribution of low pressure system durations (days) according to ERA-I. (b) Distribution of the duration (days) of the TD stage for cyclogenetic TDs identified with ERA-I and IBTrACS.](image-url)
TD and TS initiations (Fig. 7) and better for TDs (around −0.7) than for TSs (around −0.5). This correlation indicates fewer TD and TS initiation during El Niño year in the eastern Indian Ocean (80°–110°E). This is somewhat in agreement with Ho et al. (2006) who reported a longitudinal westward shift of the cyclogenesis during El Niño years. However, there is no correlation found here between TD or TS initiations and the seasonal Niño-3.4 index for the western Indian Ocean (50°–80°E). This is in agreement with Ramsay et al. (2012), which showed a significant modulation of the TC number for the eastern SIO only.

d. Modulation of TD initiations and intensification by the MJO

The MJO modulation of TS initiations (i.e., the point at which the TD reaches the TS stage) is shown in Fig. 8a for the six MJO phases and for no-MJO periods. In

![Fig. 6. Interannual distribution of the number of (a) ERA-I TD and (b) IBTrACS TS initiations in the southern Indian Ocean during austral summer (December–March). Years on the abscissa correspond to the year of December for the October–March season. The horizontal lines give the number of initiations for a statistical significance of 95% and 99%.

![Fig. 7. Total number of initiations between 80° and 110°E as a function of the average Niño-3.4 index for each austral summer (October–March). (left) ERA-I TD and (right) IBTrACS TS initiations. The regression lines and the linear correlation coefficients are also shown.](image-url)
agreement with previous results (Liebmann et al. 1994; Hall et al. 2001; Bessafi and Wheeler 2006), there are more TSs initiated after the most convectively active MJO phase and less TSs initiated during suppressed MJO phases (99% confidence for phases 1 and 5). The TD and cyclogenetic TD initiations are significantly more numerous during phase 4 and less during phases 1 and 2. The duration of the TD stage (Fig. 5b) thus tends to lag TS initiations toward phase 5. For either TD or TS, the initiation probability for no-MJO periods is not significantly different from the average probability. This shows that a MJO event modulates the initiation of the storms but that there is no significant nonlinear effect giving more or fewer storms on the average for large MJO events compared to periods of weak MJO signals. The ratio of cyclogenetic TDs (Fig. 8b) is slightly modulated by the MJO with a larger value during phase 4 and a smaller value during phase 1. The modulation of this ratio is too weak to be significant at the 95% level, but the coherence of this modulation (regular increase between phases 1 and 4) suggests that the MJO indeed tends to modulate slightly the chance for a TD to reach the TS strength over the SIO domain. These results are similar when considering years 1999–2012 for which both ERA-I and RSMC evaluations are expected to be more reliable for this region thanks to additional satellite data (e.g., relocation of Meteosat-5 over the Indian Ocean in July 1998).

The MJO thus modulates primarily the number of TD initiations. The chance for a TD to reach the TS strength is weakly modulated by the MJO. The TS initiations are lagged by about one MJO phase (i.e., about 6 days) compared to TD. These results, based on 34 years of the ERA-I dataset and an objective TD detection approach, refine the conclusions of previous papers by attributing the MJO modulation of developed cyclones to either the TD initiation probability or to the probability of a given TD to reach at least the TS strength. Because of the relatively small number of systems, the precise value of these probabilities are subject to some variations (TD detection criteria, attribution of IBTrACS systems, etc.) and only the order of magnitude has to be considered. For the whole SIO domain, there are around 3 times more cyclogenetic TDs initiated in phase 4 compared to phase 1. Considering the eastern part of the SIO domain only, this ratio is around 5. For both regions, a larger part of the modulation is due to the number of TD initiations, and a smaller part is due to intensification. The part due to the intensification (around 1.5) is similar for both regions. The modulation of the cyclogenesis by the MJO and the difference between the eastern and the western part of the basin are thus primarily due to MJO modulation of the TD initiation probability.

As expected, the MJO modulation of TD initiations is larger over region C compared to regions A and B (not shown). An intriguing point is the relatively small
probability (0.7%) for TD initiation in the no-MJO category over region A compared to regions B (19.9%) and C (1.6%), the probability being around 1.25% on the average for the other categories and nearly identical for the three regions. This suggests a decrease of TD initiation in no-MJO periods for region A and an increase for regions B and C (and thus possible nonlinear effect of the MJO regionally, while this was not significant for the whole SIO domain). Another interesting point is the seasonal variation of the TD initiation with more initiation during January–March (JFM) in the western and central SIO and more initiation during October–December (OND) in the eastern SIO. There are thus more TDs initiated over the eastern SIO (region C) when the ITCZ is closer to the equator.

4. Physical environment of TD initiations

The advantage of having an early TD detection is that the environment is not yet perturbed by the depression itself. This allows studying the main physical processes at the origin of TD initiations over the SIO domain and understanding how these processes are modulated by the MJO. We consider all TD initiations in order to increase the statistics since no significant difference was found between initiation conditions for cyclogenetic and noncyclogenetic TDs (not shown). This section analyzes composites of key atmospheric variables just prior to TD initiations in each of the three regions highlighted in Fig. 4. These three regions are selected because they have higher initiation probability all along the 33 austral summers (Fig. 4c). The choice of these three regions, instead of a composite over all TD initiations, conserves the geographical specificities of the composited fields and thus gives a basis of comparison between TDs initiated close to the Maritime Continent and those initiated in the middle or in the western part of the Indian Ocean. Average anomalies are computed for all days corresponding to a TD initiation inside one of the three regions and for lags up to 6 days. For a selected day, the anomaly is computed with respect to the corresponding monthly mean. All realizations are supposed to be independent and the significance level is computed considering that the average follows a normal law with a standard deviation of $\sigma/\sqrt{N}$, where $N$ is the number of realizations and $\sigma$ is the average among monthly mean standard deviation weighted by the monthly number of TD initiations.

a. Composites for the three initialization regions

As mentioned in the introduction, the large-scale vorticity and especially the meridional cyclonic shear can play a major role in TD initiations by barotropic energy conversion from the low-level mean flow (Maloney and Hartmann 2001). The average meridional shear of the zonal wind ($\partial U/\partial y$) at 850 hPa (Fig. 9a) shows a zonally elongated region of positive (cyclonic) values for the whole Indian Ocean between $5^\circ$ and $15^\circ$S with a notable maximum value around $10^\circ$S, $75^\circ$E corresponding to region B that is the region of maximum TD initiation probability (Fig. 4). This shear results from average westerlies to the north and easterly trade winds to the south. The average shear anomaly for days corresponding to a TD initiation in region B (this is similar for regions A and C) shows a strong local reinforcement of the cyclonic shear (Fig. 9b). Note that the shear anomaly has a zonally elongated shape showing that the TD initiation is related to a reinforcement of both the westerlies and the trade winds at a larger scale.

The origin of the reinforcement of the shear is investigated for each of the three regions by considering the meteorological fields 3 days before TD initiations. For these dates, there is no contamination by the depression itself and the average fields reveal the meteorological conditions favorable to TD initiations. Three days before TD initiation, there is already a significant reinforcement of the positive meridional shear of the zonal wind ($\partial U/\partial y$) northeast of the initiation region (Fig. 10). This is due to reinforcement of both westerlies to the north (stronger for region C) and easterlies to the south (stronger for region A). While this reinforcement is larger close to the TD initiation region, the positive shear anomalies cover most of the Indian Ocean with a slight southwest–northeast orientation. This orientation and the zonal extension of the shear anomaly are more evident for the composite corresponding to TD initiations over region C.

To establish if these anomalies of the low-level shear are sufficient to initiate barotropic instabilities, the average meridional gradient of PV is computed for longitudinal bands of $20^\circ$ around each initiation region. Two days before TD initiation, the shear already reduces the meridional PV gradient in the initiation region (Fig. 11). Just prior to or at the TD initiation date, the shear anomaly is large enough to give an inversion of the meridional gradient of PV suggesting that this TD initiation indeed results mostly from barotropic instability of the low-level wind. The inversion persists a few days after the TD initiation as the TD drifts southwestward. South of $10^\circ$S, the average PV gradient is weaker in the eastern part of the Indian Ocean basin (regions B and C), but it is not clear if this gives more favorable average conditions for the gradient inversion since this inversion occurs around $10^\circ$S where the gradient is still steep. In fact, this weak PV gradient south of $10^\circ$S could be
more a consequence of the more numerous TDs because TDs tend to steepen (reduce) the PV gradient north (south) of the TD initiation latitude (as seen in Fig. 11 for the +2-day curves).

Figure 11 also displays the MJO modulation of the meridional gradient of PV. Around region A and B, the meridional PV gradient is indeed reduced in phases 4 and 5 and steepens in phases 1 and 2 around 10°S. For region C, the PV profile is closer to the average profile in phase 4, reduced in phases 5 and 6, and steepened in phases 1–3. These MJO modulations of the PV gradient are thus consistent with the MJO modulation of TD initiations with a reinforcement of the “potential” barotropic instability (i.e., a reduction of the PV gradient giving a sort of “dynamical preconditioning”) in active MJO phases. At this stage, however, it is difficult to consider the MJO modulation of the PV gradient as a pure forcing. The presence of more TDs in active phases may also influence the PV gradient and may explain part of the gradient reduction, especially for phases 5 and 6. Other aspects of the MJO influence are discussed below.

As noted above, the strengthening of the meridional shear of the zonal wind results from a reinforcement of both westerlies to the north and easterlies to the south. Westerlies near the equator are typically linked to the presence of a low pressure anomaly (e.g., Gill 1980). Since these low pressure anomalies are likely to be related to deep convection, composites of GPCP precipitation fields are constructed. These composites (Fig. 12) indeed show that the reinforcement of the westerly wind is related to a precipitation anomaly located north of the maximum cyclonic shear band. The low-level flow is not geostrophic around the heat low associated with this precipitation anomaly, but westerly under the heat low. This is due to two factors related to Coriolis. First, as noted in Ferreira and Schubert (1997), the decrease of inertial stability toward the equator makes it easier for the heating low to draw mass from the equatorward than from the poleward side. Second, the flux converging toward the heat low stays more meridional on the equatorward side (because of the weak Coriolis force) and penetrates far into the heating low before being oriented westerly. This explains why the
westerly wind anomaly is located quite south under the heat low instead of resulting from cyclonic vortex as expected farther from the equator.

On the southern side of the shear anomalies, the reinforcement of the easterly trades appears to be related primarily to large-scale anomalies of the pressure fields. Three days prior to TD initiation over regions A and B, there is an extension (or reinforcement) of the Mascarene high (Fig. 13). The perturbation for region B is less significant with no region passing the 95% level. This region is nevertheless the region with maximum shear on the average (Fig. 9) and thus also needs less shear anomaly to be unstable. For region C, the easterly wind reinforcement south of the initiation region is related to a large-scale cyclonic geostrophic circulation around a depression located over the western Maritime Continent. As seen in Fig. 13, the west branch of this large depression exhibits a southwestward extension of

**FIG. 10.** Composite anomaly of the meridional shear ($\text{color; } s^{-1}$) of the zonal wind 3 days prior to TD initiation in one of the three selected regions (a) A, (b) B, and (c) C with 65, 91, and 83, respectively, TD initiations between 1979 and 2012 for regions A, B, and C. The regions are represented by the black boxes. The vectors represent the corresponding anomalies of the 850-hPa wind. The red contours represent statistical significance of the anomaly at the 99% (bold) and 95% (thin) levels. An arrow length of 1° of latitude–longitude represents a wind speed of 0.7 m s$^{-1}$. 

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the depression that could be related to the Rossby gyres associated with the stationary Matsuno–Gill pattern. The precise origin of the shape of this extension is, however, unclear, but it is also associated with a precipitation anomaly (Fig. 12). The westerly wind anomaly just south of the equator also results certainly in part from a Matsuno–Gill circulation associated with this large-scale Maritime Continent depression and is reinforced by the additional heat low related to the local precipitation (Fig. 12).

b. The MJO Perturbations

This section establishes a link between the large-scale perturbations related to the MJO and the TD initiation over the three regions. The perturbation of the meridional shear of the zonal wind in phase 4 of the MJO (maximum convective activity) is a large band of enhanced cyclonic shear around 10°S (Fig. 14a). This perturbation is hardly significant at the 95% level, possibly because of variations of the shear pattern during one MJO phase (6 days) and/or because of the variability of this pattern from one MJO event to another. The zonal wind perturbation is nevertheless significant for regions around the equator between 60° and 80°E (not shown). Note that the significance level is computed here by dividing the number of days in a given MJO phase by 7 (i.e., 7 consecutive days in a given MJO phase are supposed to give the same information, this minimizes the degree of freedom and the significance level). The average wind perturbation pattern for phase 4
shows a westerly wind perturbation north of a southwest–northeast-oriented line and an easterly wind perturbation south of this line. This wind pattern is clearly related to a large and statistically significant geopotential perturbation (Fig. 14b) that resembles the perturbation observed for TD initiations over region C (Fig. 13c). This confirms the larger link existing between the MJO and TD initiations in the eastern SIO (region C) compared to the western SIO (regions A and B). The MJO also has a statistically significant signature in thermodynamic variables such as precipitation (Fig. 14c). Phase 4 of the MJO is, by definition, related to a larger convective instability and precipitation over the eastern SIO and to a larger humidity of the midtroposphere. The precipitation pattern for phase 4 appears to lead (shifted to the west) somewhat the conditions for TD initiations over region C (Fig. 12c) for which the TD initiation is maximal in phases 4 and 5.

It is worth noting that, as for the composites of TD initiations, there is a low-level westerly wind anomaly under the MJO precipitation perturbation. This is because physical processes given above are certainly also
valid for the many convective cloud clusters developing close to the equator in the MJO envelop. However, the composite wind anomaly corresponding to TD initiation for region A is local and does not correspond to the large-scale dynamical MJO perturbation. For region B, it has only some resemblance with the MJO signature and there is no such resemblance for the geopotential anomaly. In addition, the reinforcement of the easterly trades south of regions A and B appears to be related mostly to a reinforcement of the Mascarene high with no evident link with the MJO anomaly. The large-scale dynamical and thermodynamical impacts of the MJO is thus of secondary importance for TD initiation over the western and central SIO. The enhanced TD initiation in MJO phase 4 (Fig. 8) is mostly due to the eastern SOI.

5. Summary and discussion

The tracking of tropical depressions using meteorological reanalyses at relatively high spatial and temporal resolution is useful to complement the cycogenesis information given by RSMC. Using reanalyses makes it possible to use objective approaches to detect and characterize TDs at an early stage, while RSMC information reported in databases such as IBTrACS are required to identify precisely the cyclonic nature of
developed systems. This is the approach adopted here where the IBTrACS dataset is used only to identify intensifying tropical depressions (i.e., TDs reaching the TS stage). This study essentially focuses on TD initiations that are probably more homogeneous in ERA-I than in IBTrACS because the detection is based on a more objective approach.

The IBTrACS dataset is still necessary to identify systems reaching tropical storm intensity. Some attempts to link the TD intensities (maximum surface wind, minimum surface pressure) in ERA-I and IBTrACS gave indeed very noisy and insignificant results (not shown). This may be obviously due in part to inaccuracy of intensity estimates by both ERA-I and IBTrACS. Over the SIO, there are indeed some discrepancies between TC intensity estimates from different RSMC sources (Schreck et al. 2014) and there is still no recurrent in situ measurement allowing validating or adapting the Dvorak technique. The inconsistency between TD intensities in ERA-I and IBTrACS is, however, too large to be due only to these inaccuracies. In fact, the spatial resolution of the reanalysis cannot...
resolve the dynamics of the developed systems and this inconsistency rather suggests that complex and sporadic relations exist between the strength of the synoptic-scale depression (resolved by ERA-I) and the strength of the surface wind related to the mesoscale dynamics of the cyclone.

The paper first focused on the climatology of TD initiations over the south Indian Ocean (SIO). For the SIO, around 34% of the considered ERA-I TDs (i.e., with the criteria given in section 2) attain the TS strength compared to around 85% for IBTrACS. This is expected since only the strongest systems are reported in IBTrACS. It is hypothesized that all TDs must be considered in the analysis of the origin of developed cyclone if one considers that all TDs are susceptible to intensification. Many noncyclogegetic TDs of the SIO have durations between 5 and 10 days and some have durations larger than 10 days. For cyclogegetic TDs, the TD stage duration is between 2 and 4 days, but this duration is longer for many TDs and can even exceed 10 days. This has consequences on the geographical distribution of TS initiations. While TD and cyclogegetic TD initiations clearly occur preferentially at three longitudes located near 55°E, 7°E, and 95°E, TS initiations are more evenly distributed in longitude. This suggests that TD initiations are more reproducible and linked to the average dynamics of the SIO basin than the intensification to the TS stage.

The physical source of TD initiations over the SIO is analyzed for these three longitudes. The sources of intensification between TD and TS appear to be more complex (probably with some stochastic aspects) and are not analyzed in the present study. For the three longitudes, there is a remarkable similarity in precipitation and 850-hPa wind anomalies 3 days before TD initiation with a positive precipitation and a westerly wind anomaly north of the initiation region. This is associated with an easterly wind anomaly south of the precipitation region that can be related mostly to large-scale geopotential anomalies. This generates a cyclonic meridional shear of the zonal wind that is quite elongated in longitude and results in an inversion of the meridional PV gradient that favors the development of barotropic instabilities. It is hypothesized that this entire process is the main source for TD initiation over the SIO.

Concerning the MJO modulation of cyclonic and noncyclonic TDs, the present study reinforces but also mitigates the Liebmann et al. (1994) conclusion that the MJO modulation of the number of TCs is due to the modulation of TD initiation and not to the intensifying processes. This implies that the probability of TD intensification is poorly related to the perturbation of the large-scale atmospheric state by the MJO. The present study reinforces these conclusions by showing similar results obtained with a significant temporal extension of the analysis (by a factor of 3) and with an extended TD ensemble given by ERA-I reanalyses compared to RSMC reports (by a factor 2.5). However, this study also mitigates this hypothesis by showing that the TD intensification probability is also slightly modulated by the MJO with a ratio of around 1.5 between the active and the suppressed phase. This is true for the entire SIO and for regions east of 80°E considered separately. The difference in the MJO modulation of the number of TCs between these two longitude bands is also mostly due to differences in the modulation of TD initiations. The MJO thus modulates significantly the number of TD initiations over the SIO with a maximum initiation in the active phase defined here for a maximum convection around 85°E. There is no significant nonlinear effect at the scale of the SIO since there is as much TD initiation for MJO or for no-MJO periods (or for MJO events with large or weak amplitude). These characteristics are, however, not homogeneous over the SIO domain and some differences exist between western and eastern SIO.

For the three main initiation regions, TD initiation will be favored when a precipitation anomaly near the equator is concomitant with a reinforcement of easterly wind to the south. The nature of this reinforcement is, however, different in the western and the eastern SIO. In the western SIO (region A), this easterly wind anomaly is related to a strong and significant reinforcement of the Mascarene high. In the eastern part of the SIO (region C), this easterly wind anomaly is related to a large and significant low pressure over the western Maritime Continent. Over the central SIO (region B) the pressure anomaly is weaker and not significant at the 95% level. This weaker average pressure anomaly over region B could result from a competition between the two types of forcing. This requires further analysis.

In the western SIO, the concomitance of the precipitation anomaly south of the equator and of the reinforcement of the Mascarene high that leads to an increase TD initiation could be mostly random and explains why the modulation by the MJO, or the impact of ENSO is weak. However, the lower TD initiation probability over region A for the low MJO signal is intriguing. This suggests that conditions favorable to TD initiation in the western SIO are also conditions favorable to the triggering of larger MJO perturbations. For this region, the TD initiation is in fact slightly more likely for MJO phases 1–3 corresponding to suppressed conditions preceding the MJO active phase over the eastern SIO. This intriguing point could thus be related to the fact that enhanced organized convection in the
western SIO and larger Mascarene high exist mostly prior to the triggering of the large MJO event. This also needs further analysis, especially concerning the definition of the MJO signal (e.g., the choice of the threshold between MJO and no-MJO cases).

In the eastern SIO, the precipitation anomaly and the reinforcement of easterly wind to the south are both related to the same large-scale low pressure over the western Maritime Continent. This large-scale low pressure is very close to the geopotential anomaly obtained for MJO phase 4 and corresponds well to a Gill-type response to an equatorially centered low pressure. The circulation is cyclonic in both hemispheres and gives a low-level convergence near the equator that is finally associated with a precipitation and a westerly wind anomaly near the equator. The cyclonic meridional shear of the zonal wind and the associated inversion of the meridional PV gradient are thus primarily related to this large-scale low pressure. This explained the tightest relation between the MJO and TD initiations for region C. The largest TD initiation numbers for La Niña years over the eastern SIO may be also attributable to the corresponding reinforcement of the westerly wind near the equator.

Two other points are probably worth being further studied. The first is the large and significant decrease of TD initiations for the El Niño winter 1997/98 with no counterpart in TS initiations. This suggests a compensation between the number of TD initiations, probably reduced in El Niño conditions because of the reduced equatorial low-level westerly wind, and an increase of the probability of TD to TS transitions, possibly related to other large-scale perturbations (e.g., the SST) favoring the TD intensification. The second point is the longitudinal distribution of the TD initiation with maxima over the three regions A, B, and C. Region B is the region of maximum average meridional shear of the zonal wind (Fig. 9a) and this may explain the maximum TD initiation probability over this region. The TD initiations over region C are tightly related to the presence of large-scale low pressure over the western Maritime Continent (Fig. 13c) that reinforces the cyclonic shear in the eastern SIO. This also generates a band meridional shear over most of the SIO (Fig. 10c) that may promote a large-scale instability initiating several cyclones, as for ITCZ breakdown episodes studied in Ferreira and Schubert (1997). The average effect of multiple vortex initiations is visible for the composite 1 day before TD initiation over region C (Fig. 15a). The
secondary vortex corresponds to a significant perturbation of the geopotential height and appears as a consequence of the tongue of low pressure already visible 2 days before (Fig. 13c) and corresponding to the zone of stronger meridional shear (Fig. 10c). This secondary vortex is located over region B and may give TD initiations over this region. The distance between the primary and the secondary vortex corresponds to a wavelength of roughly 20° and, according to Ferreira and Schubert (1997), this should correspond to the most unstable mode for an idealized PV strip of around 3.5°. This seems small compared to the PV profiles reported in Fig. 11 and requires further investigations. Another explanation for the three regions of enhanced TD initiation is the possible formation of a wave train associated with the reinforcement of a depression. An illustration is given in Fig. 15b with such a wave train existing on the average 2 days after TD initiation over region A with a secondary cyclonic circulation close to regions B and C. This is not significant on the composite, but this signature suggests that a deepening TD over one region may generate other depressions east and west of the initial TD with a wavelength consistent with the longitudinal separation between the three regions.

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