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Occurrence of monsoon depressions in the Southwest Indian Ocean: Synoptic descriptions and stratosphere to troposphere exchange investigations

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[1] This study focuses on Monsoon Depressions (MD) in the Southwest Indian Ocean (SWIO) basin. A MD is a subcategory of tropical depressions. Some aspects of MD dynamical characteristics differ from those of the classical tropical depressions. A MD displays a broader horizontal extension (>1000 km), and its structure is marked by deep convection clusters that are generally poorly organized. These convective clusters often develop as a wide convection belt in the eastern semicircle of the system. The MD nucleus is characterized by an area of weak winds about 200 km wide and by a hazy central cloud system, surrounded by a larger belt of strong winds. In the SWIO basin, between 2000 and 2008, 5 systems, out of the 100 tropical cyclones that were followed by RSMC La Réunion, comply with these criteria and then have been identified as MD. Using potential vorticity from European Centre for Medium-Range Weather Forecasts analyses as stratospheric tracer, stratospheric signatures have been identified in the troposphere for the five events, involving different dynamical mechanisms: tropopause fold between an anticyclone and MD, tropopause fold between MD and a jet front system, and cut-off low. The dynamical structure of MD, especially the external belt of strong winds, seems to favor tropopause disturbances and leads to stratosphere–troposphere exchange.

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1. Introduction

[2] The Earth's climate is governed by the solar flux entering the atmosphere. Marine and atmospheric flows contribute to balancing the received solar flux. Tropical and equatorial regions receive the most solar flux annually [*Sellers*, 1965; *Stephens et al.*, 1981]. In these convection-prone regions, the formation of tropical depressions or cyclones occur, with the tops of these storm potentially reaching the tropopause. Under specific conditions, the altitude dynamics and convection mechanisms can impact the stratospheric water vapor burden [*Potter and Holton*, 1995] or lead to stratosphere–troposphere exchange [*Baray et al.*, 1999; *Zachariasse et al.*, 2000, *Leclair de Bellevue et al.*, 2006].

[3] Monsoon areas are tropical convective regions where the wind direction differs by at least 120° between the wet and the dry seasons [*Ramage*, 1971]. The primary cause of monsoon is the much greater annual variation over large land areas compared with neighboring ocean surface. The low-tropospheric flow inside the monsoon is called the monsoon flow, and its intensity reaches a significant value (more than 2.5 m/s). Some monsoon areas, in particular the Indian monsoon, favor the development of Monsoon Depression (MD). These are a subcategory of tropical depressions, which conform to some precise dynamical criteria (Elsberry, Tropical Meteorology class notes, 2002) based on the horizontal extension, the structure of deep convection clusters or of winds, and described in detail in section 2.

[4] Mostly covered by an oceanic surface, the Southwest Indian Ocean (SWIO) is not affected by monsoon. However, it happens that some tropical cyclones in the SWIO, which are observed and predicted by the Regional Specialized Meteorological Centre (RSMC) of Reunion Island, do conform to the dynamical definition of a monsoon depression between November and April (wet season). In these circumstances, something similar to a monsoon flux flows through the Intertropical Convergence Zone (ITCZ) and gives birth to MD. Some aspects of MD dynamical characteristics differ from those of the classical tropical depressions. The aim of this paper is to study the dynamical aspects of these systems developing in the SWIO and precisely

[5] • To characterize the dynamical mechanisms of MD

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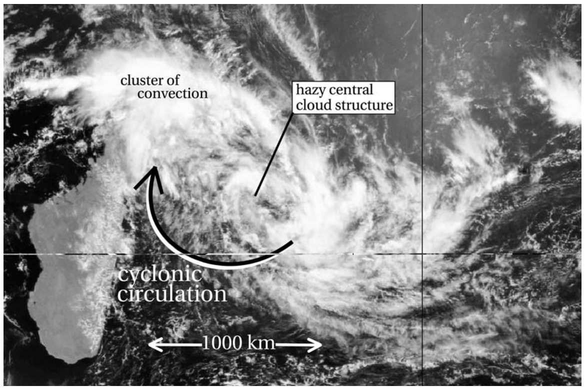

Figure 1. Meteosat Image in the visible channel on 8 March 2006 at 08:00 UTC: MD DIWA.

[6] • To describe the frequency of appearance of MD in the region

[7] • To establish whether MD can infer stratosphere–troposphere exchange and according to which processes

2. Definition of a MD

[8] The formation and development mechanisms of MD can exhibit some differences at the initial stage compared to other tropical depressions. MD originates inside the monsoon flux, from its convergence with trade, around a large hazy area of low pressures.

[9] The monsoon flux is characterized by the inversion of the wind direction between the wet season and the dry season. The official definition of monsoon fluxes by *Ramage* [1971] is based on the four following criteria:

[10] • A minimum change of 120° of the surface wind direction must be observed between January and July.

[11] • These winds must be observed at least 40% of the observed month.

[12] • The mean surface wind speed must be greater than 3 m/s.

[13] • The pressure spatial structure must not be controlled by frontal systems.

[14] One of the specific formation modes of MD occurs when convective belts are associated to the two fluxes, the latter being wrapped around the convergence area. A MD has different dynamical characteristics compared to the other types of tropical depressions. Its development leads to a wide closed circulation of weak winds in the middle and stronger winds along the edge of the system (Elsberry, Tropical Meteorology class notes, 2002). The MD structure can be compared to a large cyclonic tropical circulation depicted by *Raman et al.* [1978] and Elsberry (Tropical Meteorology class notes, 2002): It has a wide size ranging about 1000 km, convective elements forming deep convection clusters with a hazy structure but frequently arranged as a wide convection belt along the eastern semicircle of the cloud circulation, a nucleus of weak winds ranging about 200 km of diameter surrounded by a belt of asymmetric and stronger winds with large radius, and a lack of distinct central cloud structure. An example of MD is visible on the Meteosat image (Figure 1). This is the MD DIWA which approached Reunion Island in March 2006.

[15] In the tropics, convection mechanisms have an impact on meteorology (surface winds, rainfall) and on the transport of trace gases such as ozone [*Chatfield and Delany*, 1990; *Lelieveld and Crutzen*, 1994; *Jonquières and Marenco*, 1998] or water vapor [*Lohmann et al.*, 1995; *Liao and Rind*, 1997; *Soden*, 2000]. Convection mechanisms can disturb the tropospheric dynamics [*Chen and Houze*, 1997] and especially the TTL dynamics [*Dessler*, 2002; *Gettelman et al.*, 2002].

[16] RSMC La Réunion follows every organized cyclonic convective system in the SWIO whose maximum sustained winds over 10 minutes are above 45 km/h. These cyclones are classified according to their intensity, as tropical disturbance, tropical depression, tropical storm or tropical cyclone. A monsoon depression that conforms to the intensity criterion is followed by RSMC La Réunion, even if its structure and its dynamics do not conform to the classical mechanisms of a tropical cyclone.

3. MDs in the World

[17] Relying on the definition given in section 2, monsoon regions are central Africa, India and the north of the Indian Ocean, Australia, Indonesia, and Southeast Asia. Monsoon depressions may be found in this monsoon area, but they can also be found in other tropical regions, such as in the Northwest Pacific Ocean [*Harr et al.*, 1996].

[18] Still, the region where MDs are the most common is the North Indian Ocean (NIO), in the area of the Indian Monsoon [*Ramage*, 1971]. Thus, MDs in this region have been the most extensively studied, and the MD definition (section 2) comes from observations in this area. Oppositely to tropical storms and cyclones, MDs can remain over lands for many days with moderate fading thanks to a slow motion speed (500 km/day) and a stronger circulation above the surface. These characteristics lead to heavy and long-lasting precipitation, mainly located in the southwestern part of the depression [*Ramage*, 1971].

[19] The impact of MDs on monsoon precipitation has been estimated to 45% to 55% of the total Indian rainfall [*Yoon and Chen*, 2005]. The lifetime of a MD in the Bay of Bengal is 3 to 4 days during the Indian monsoon [*Mahajan et al.*, 2004].

[20] *Daggupaty and Sikka* [1977] described precisely the synoptic features of the MD in the NIO. MDs develop in the Indian monsoon flux, in association with the tropical easterly jet (TEJ), an altitude flow propagating from the east toward the west whose intensity may reach values above 110 km/h. The vertical wind shear is thus significant in the monsoon area. Most MDs develop from a weak easterly wave coming from Southeast Asia and intensify during westward propagation westward through India against the low tropospheric monsoon flows [*Krishnamurti et al.*, 1977], possibly a residual wave coming from the Western Pacific [*Chen and Yoon*, 2005]. Winds are strong at the surface with a maximum intensity at 800 hPa. Above this level, the winds are rapidly decreasing due to vertical wind shear, which also prevents MD from developing as a tropical cyclone. Vertical motions of deep convection occur in the western area of the depression, which accelerates the westward propagation of the system [*Daggupaty and Sikka*, 1977]. The temperature anomaly associated with a MD is quite unusual for a depression, since it is negative in the low levels, reaching its

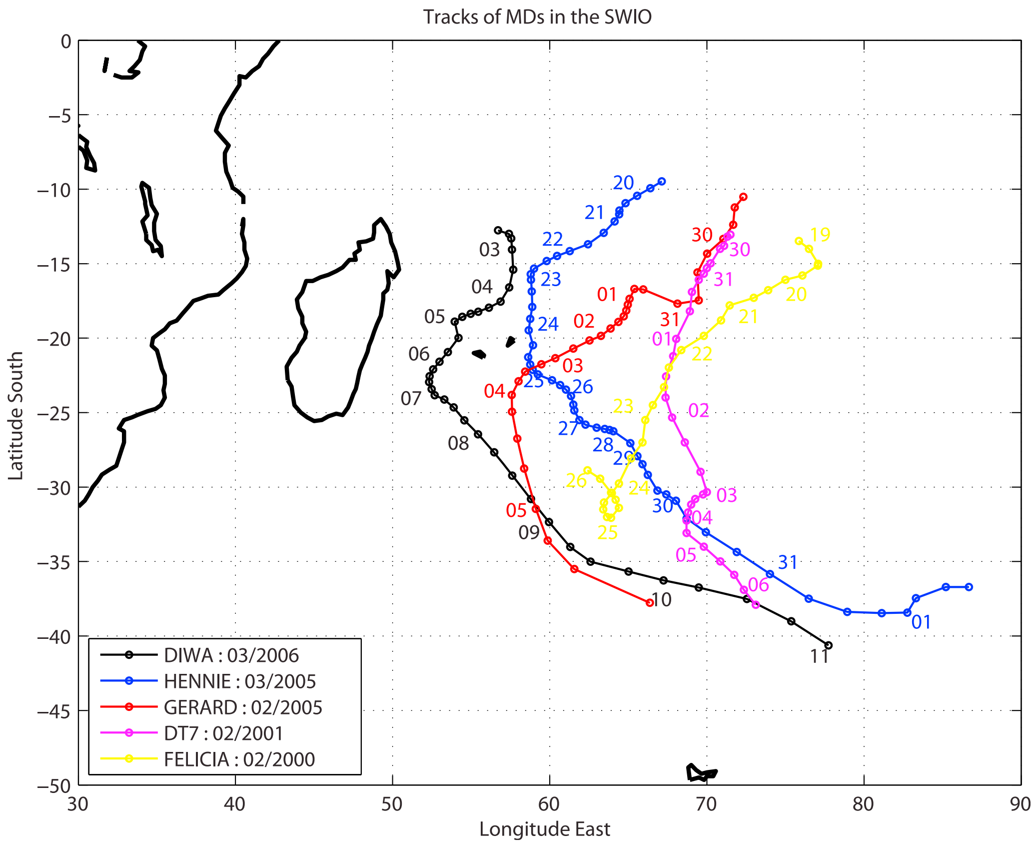

Figure 2. Tracks of five MDs in the SWIO.

coldest values around 900 hPa, and slightly positive in the upper levels.

4. MDs in the SWIO

[21] Although the SWIO is mostly covered by the ocean and the monsoon cannot develop in this basin, some MDs as defined as in section 2 may be encountered in the SWIO. A climatological study of tropical cyclones in the SWIO has been reported by *Leclair de Bellevue et al.* [2006]. This study states that the number of systems in the SWIO ranges from 8 to 15 cyclones per season, including all stages of intensity, from tropical disturbances to very intense tropical cyclones. Between 2000 and 2008, the number of cyclones in the SWIO that were followed by RSMC La Réunion (i.e., cyclonic vortex with maximum sustained wind over 10 minutes above 45 km/h) is estimated to a hundred. Five of these systems fulfilled the criteria described in section 2 and have been identified as MDs. Therefore, a raw estimate of 5% of the cyclones in the SWIO may be classified as MDs. The tracks of these five systems are displayed in Figure 2. In sections 4.1 through 4.5, a synoptic description of their development and evolution is reported. The evolution is examined with ERA-Interim analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) [*Simmons et al.*, 2007]. The dynamics of MDs in the SWIO are also compared with what is known from MDs in the NIO (section 3). ERA-Interim analyses are computed by a 4D-Var algorithm at truncation T255 and 60 vertical levels. Many observations are assimilated, including satellite radiances, which leads to a good confidence in the fields in the SWIO. The extracted fields are on a global grid at 1.5° resolution. The following specific questions will be addressed: (1) In what part of the system does convection mainly occur? (2) Is there any upper-tropospheric jet and precursor associated with the formation of MD? Some figures will be shown, at some relevant instants for the dynamics of the depressions (genesis and mature stage generally).

4.1. FELICIA: From 18 to 26 February 2000

[22] At the beginning of FELICIA in February 2000, a large area of clouds and convective activity is observed at the whole eastern part of the Southwest Indian Ocean cyclonic basin. The ITCZ has a particularly large extension up to 20°S, characterized by a strong monsoon flux (Figure 3). Within this low pressure area, a large isolated low pressure circulation with a diameter of about 1000 km is developing. It has the characteristics of a typical MD: weak winds on a fairly large extent around the center, stronger winds (40 km/h) in the periphery, with a maximum north of the MD, in the monsoon flow (50 km/h). On 19 February 2000, the system moves to the southeast. The diameter of the system is 1500 km, and its structure is circular, in agreement with an intensification of the hurricane. Using ascending vertical velocity as an indicator of large-scale convective activity, unorganized and quite volatile convective clusters are observed around the

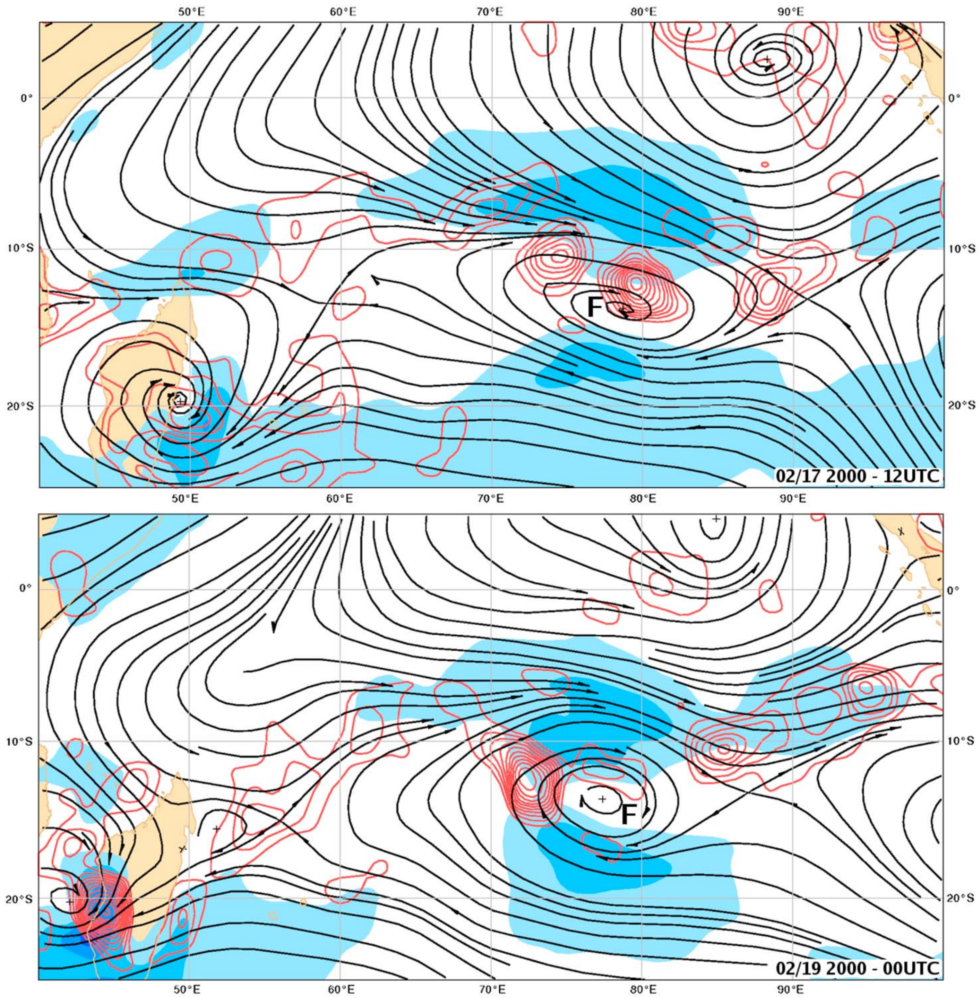

Figure 3. Synoptic environment and convective structure of FELICIA on 17 February 2000 at 12:00 UTC and 19 February 2000 at 00:00 UTC: 900-hPa streamlines in black, 900-hPa wind speed above 10 $m \cdot s^{-1}$ in blue shading (isoline 5 $m \cdot s^{-1}$), 500-hPa vertical velocity in red (only ascending motion, isoline 0.1 $Pa \cdot s^{-1}$), from the ERA-Interim reanalysis. Ascending vertical velocity indicates the areas of large-scale convective activity. The monsoon depression is marked by the first letter of its name ("F" for FELICIA).

center of the system, sometimes exploding in its western part (Figure 3). An upper-level divergence area is forming. On 20 February 2000, despite the unstable shifting and circular aspect of the system, it keeps a MD-type structure with a central area of approximately 150 km in diameter, smaller than previously, where the winds remain weak and are surrounded by a stronger winds area, 40–55 km/h in the northern semicircle and 45–60 km/h in the southern semicircle. The speed of the system is greater than 20 km/h. Its track deflects toward the southwest direction. The system is classified as a tropical depression by RSMC La Réunion, despite the absence of strong winds near the center. On 21 February 2000, the meridian extension is smaller, and the clouds are concentrated near the center. Its west-southwest-ward track and the low pressure structure of the system displays a significant asymmetry. In the afternoon, FELICIA is classified moderate tropical storm. On 22 February 2000, FELICIA gets over 20°S with a speed close to 30 km/h. At the end of the day, a very large eye of about 100 km diameter forms. The maximum of intensity is reached during the 22 to 23 February night, with the presence of a warm heart. FELICIA is classified as a strong tropical storm. Maximum sustained winds are estimated at 115 km/h. On 23 February 2000, the eye disappears. On 24 February 2000, the track takes a south-southwest-ward direction; the system is in phase of transition to an extratropical storm.

4.2. DT7: From 30 January to 6 February 2001

[23] The synoptic situation of the last week of January 2001 was affected by the presence of a large-scale low-pressure circulation south of Chagos Islands. After 24 January, a depression was already present with a large area of moderate, fluctuating, and highly fragmented convective activity located north of the 82°E, 12°S point. A well-organized monsoon

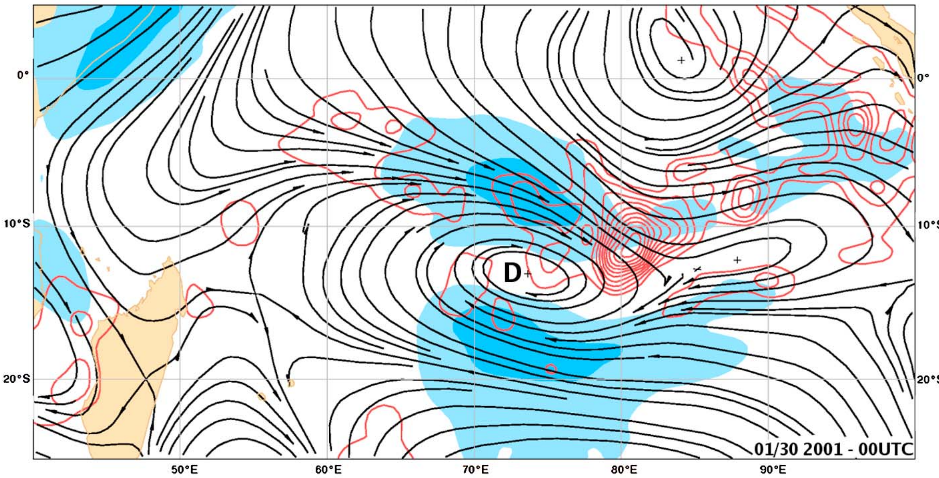

Figure 4. Same as Figure 3 on 30 January 2001 at 00:00 UTC (DT7).

flux (Figure 4) favors the intensification of this depression that moves toward the west-southwest. On 29 January 2001, the monsoon and trade wind flows increase around the depression. Despite of this, the convection remains unorganized, mostly present in the eastern and the northern parts of DT7. On 31 January 2001, the depression is growing with a depression center of 997 hPa at the end of the day. The depression is very wide with a horizontal extension of more than 1500 km and a relatively wide central part (200–250 km diameter). This central part, where the winds are weak, is surrounded by a ring 350–400 km wide, where winds have an average speed of 40–55 km/h. This ring is fed by the convergence of trade winds and monsoon flows. These observations show that this system fulfills all the characteristics of a MD. On 1 February 2001, the track is southwestward, and the system does not show a significant change except its speed, which ranges between 20 and 30 km/h. On 2 February 2001, the depression has a southward direction and crosses 20°S. On 3 February 2001, the track is toward the southeast and then toward the south, and the system is dissipating.

4.3. GERARD: From 29 January to 5 February 2005

[24] The system GERARD occurred from 29 January to 5 February 2005. This tropical storm formed from a very large low-pressure circulation at the beginning, combined with a monsoon trough whose characteristics are close to what is usually observed in the North Pacific (section 3). Initially, GERARD has a classical form of a tropical depression formed within the convergence of trade winds and monsoon fluxes. On 30 January 2005, the vertical wind shear increases and disrupts the system which finally implodes. It is unusual, since normally the wind shear tends to isolate the vortex but not to dissipate the convection. In this MD development phase, ascending motion is preferably located in the eastern part of GERARD. On 31 January 2005, a new pressure minimum is formed with weak winds near the center (Figure 5). The existence of an upper-level anomaly from extratropics triggers convection in the western part of the system. On 3 February 2005, convection has clustered at the center of the depression, but the system remains very large. On 4 February 2005, GERARD reaches the stage of tropical

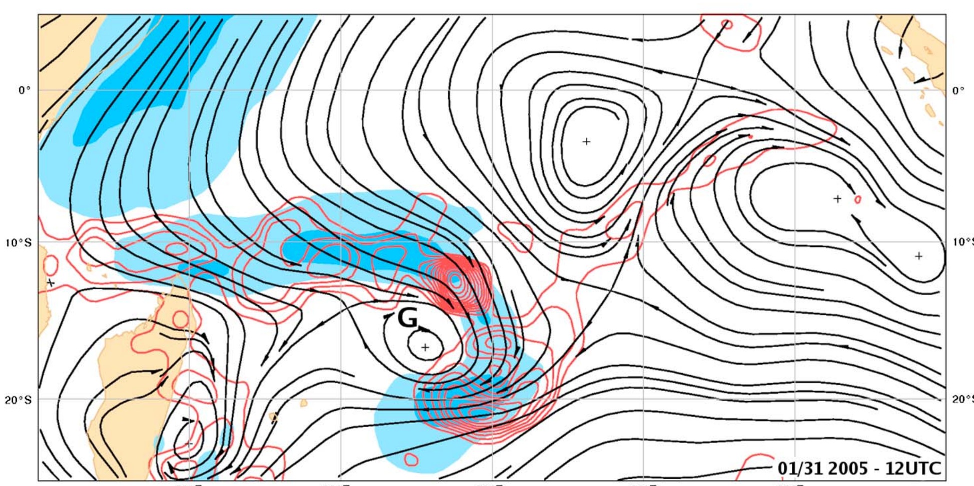

Figure 5. Same as Figure 3 on 31 January 2005 at 12:00 UTC (GERARD).

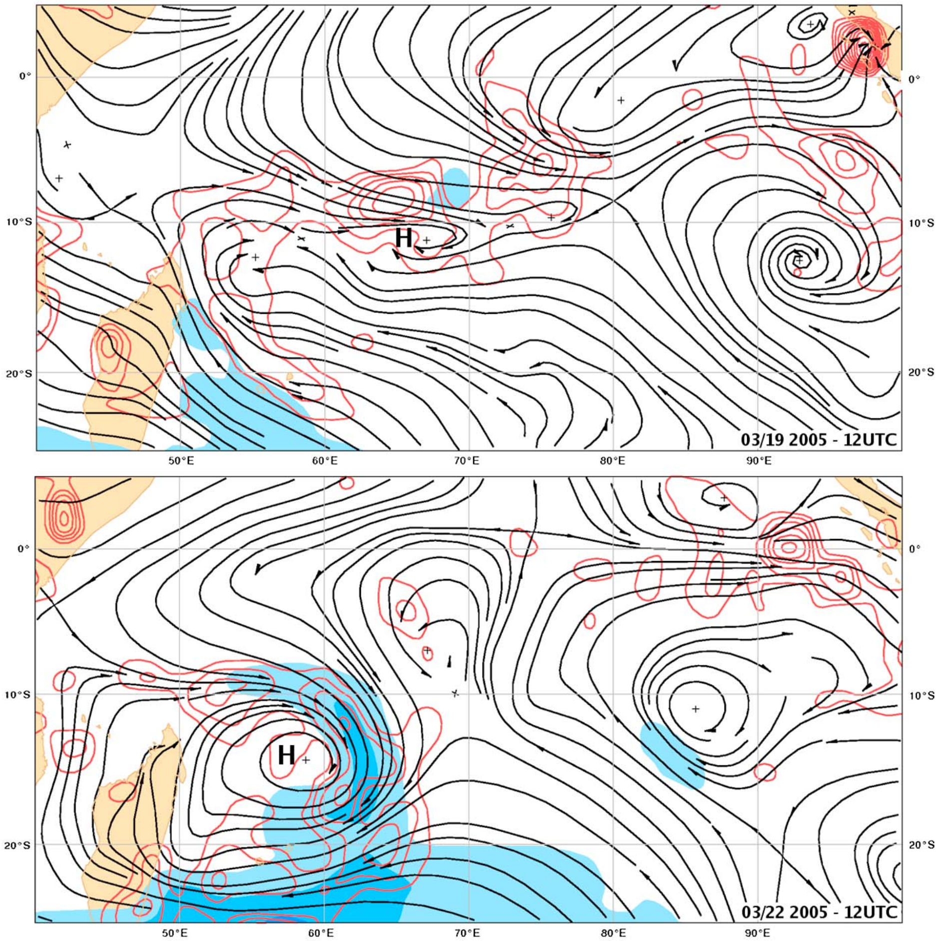

Figure 6. Same as Figure 3 on 19 and 22 March 2005 at 12:00 UTC (HENNIE).

cyclone at 30°S, which will be its maximum intensity before extratropical transition.

4.4. HENNIE: From 19 March 2005 to 1 April 2005

[25] HENNIE formed around 19 March 2005 with the classical configuration of a tropical depression. The monsoon flux is weak and not well organized (Figure 6). After 22 March 2005, HENNIE expands by a factor of 4 to 5 with a very chaotic structure and with ascending motion localized all around the MD. The system is then developing rapidly since there is convection near the center. This evolution is rather characteristic of a classical tropical cyclone, with organized cloud bands. It reaches the stage of a moderate tropical storm on 23 March 2005. During this stage, convection is rather localized in the eastern side of HENNIE. This system could therefore be described as a hybrid between a MD and a tropical cyclone, depending on its stage of development.

4.5. DIWA: From 2 to 11 March 2006

[26] DIWA occurred during one of the less active cyclone seasons for the last 30 years over the SWIO. However, DIWA was classified as a severe tropical storm, and it strongly impacted Reunion Island with strong winds and torrential rain for 4 days. A sudden activation of the ITCZ occurs at the beginning of March 2006 between the north of Madagascar and the northeast of Reunion Island, generating a well-structured monsoon flux (Figure 7), which creates a large cyclonic system. Around the longitude 75°E, the pre-existent tropical cyclone CARINA vanishes progressively as it goes to the South. On 2 March 2006, the winds in DIWA are asymmetric, with stronger winds in the eastern semicircle (55–75 km/h). The ring of strong winds is located about 200 km from the center, while weak winds are observed in the central area of 150 km diameter. The area of influence of the system is very large; its horizontal extension is greater than 2000 km on the northwest to southeast axis. All these elements are characteristics of a MD. Ascending motion is preferably localized in the eastern area of the MD (Figure 7). The following days, the configuration remains highly asymmetrical with the persistence of a large area of strong winds at the east and at the south of the depression, around a wide range of weak winds. During the night from 4 to 5 March 2006, the track direction of DIWA changes toward the south-southeast, consistent with the direction of the monsoon flux and with the localization of ascending motion. On 5 March 2006, the system remains very poorly organized. After 6 March 2006, the system located southwest of Reunion Island is evolving to a more conventional tropical depression system. The maximum intensity is reached on 8 March 2006, almost at the level

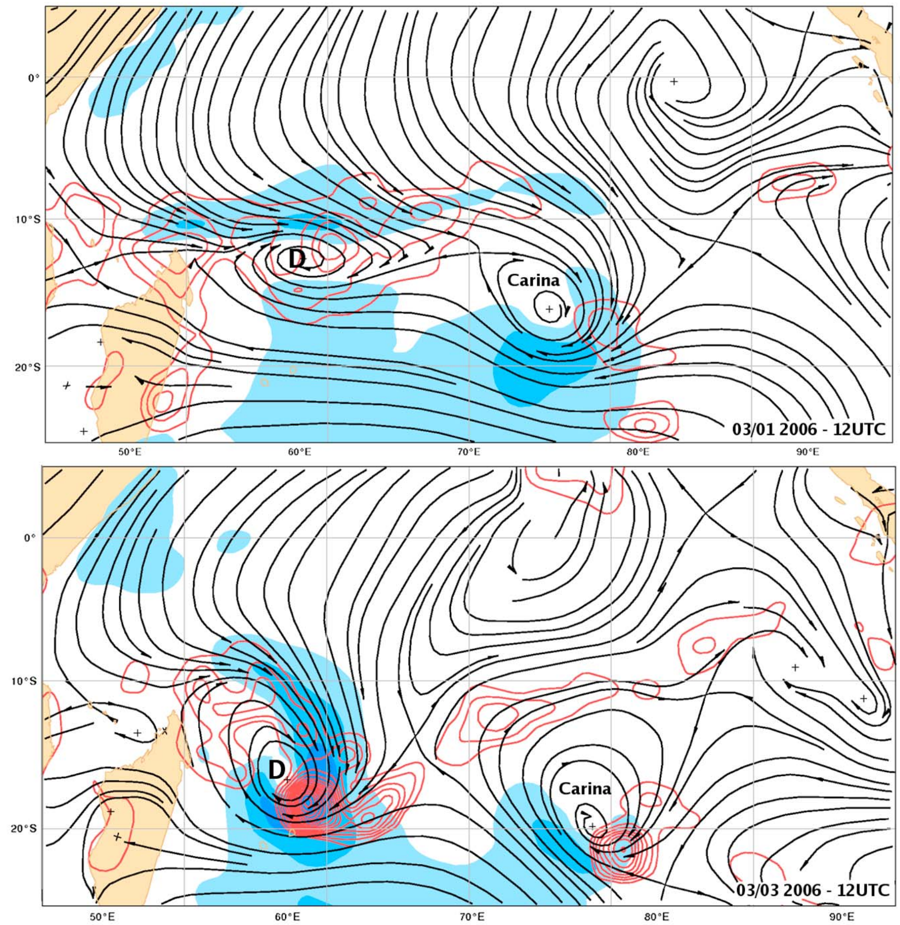

Figure 7. Same as Figure 3 on 1 and 3 March 2006 at 12:00 UTC (DIWA). The remnants of tropical cyclone Carina appear at the East of DIWA.

of tropical cyclone, before extratropical transition with a southeastward track.

4.6. Comparison of MD in the SWIO With Asian MD

[27] The tropical depressions in the SWIO that correspond to the definition of MD exhibit various synoptic descriptions and evolutions. During their formation and mature stages, their tracks (Figure 2) bend roughly toward the southwest. At the same time, convection is preferably localized in the eastern part of the MD, mostly in the southeast, except most ascending for FELICIA was at north. The localization of convection and direction of propagation is inconsistent with the observation MD in the NIO, and also with the vorticity budget by *Daggupaty and Sikka* [1977]. As a consequence, the production of vorticity by divergence is not the predominant term that explains the vortex displacement. In the SWIO, trade winds are assumed to significantly contribute to the vorticity advection term. Moreover, some phases of development may concentrate convection at the center of or all around the MD, similarly to a classical tropical cyclone (HENNIE, FELICIA). Interaction with an extratropical upper-level trough may also induce convection in the western side of the MD (GERARD).

[28] The possible role of an upper-level feature for the formation of MD is investigated through Figure 8. A well-defined upper-level easterly jet may be seen on the cases of DT7 (reaching 55 km/h) and FELICIA. In the jet corresponding to FELICIA, an upper-level trough may be seen, but its amplitude is quite weak. For DIWA, a jet is established in the eastern part of the SWIO, but DIWA generates a stronger divergent upper-level flow. For HENNIE, the upper-level field looks like the anticyclonic divergent flow that may be seen for classical tropical cyclones. The flow aloft GERARD does not have an easterly jet. Except for the case of FELICIA, no upper-level structure such as an easterly wave appears to trigger the formation of the MD. However, for DIWA (Figure 7), the role of the pre-existent cyclone CARINA to the east may be questioned: It could trigger waves in the easterly upper-level jet for instance. Low-level temperature anomalies do not show to be negative at the center of the MD, contrary to what *Daggupaty and Sikka* [1977] diagnosed in the MD of NIO. Although not shown in the figures, this has been checked on the whole evolution of the systems. It follows that MD in the SWIO do not show many of the specific features of MD in the NIO. MD in the SWIO share features with them but also with classical tropical cyclones. The upper-level jet does not seem to play the significant role that the tropical easterly jet plays for MD in the NIO. The temperature anomaly of the MD is quite classical for a depression (warm in the low levels). Contrary to tropical

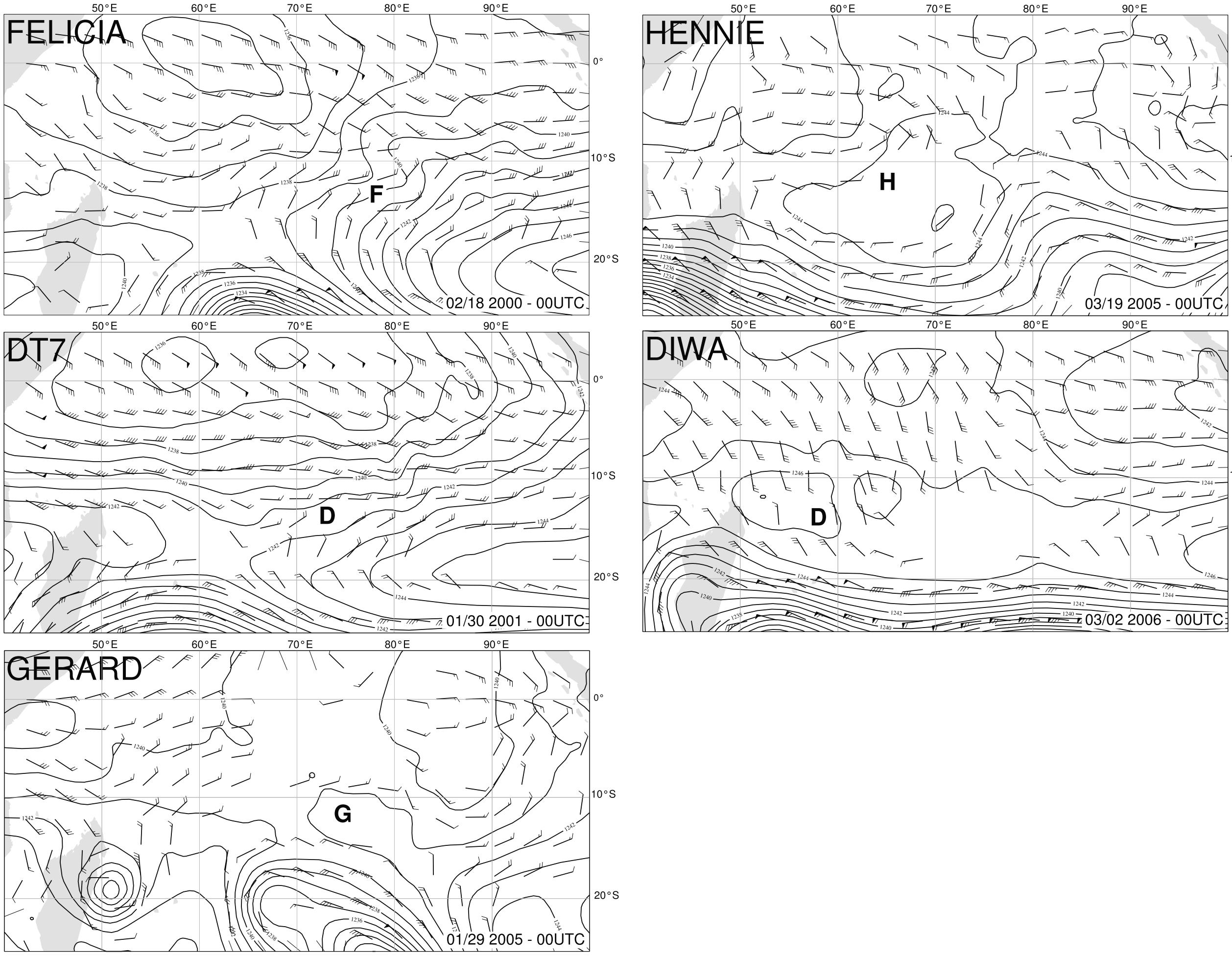

Figure 8. Upper-level wind structure: wind barbs and geopotential (isoline 1 damgp) at 200 hPa at the beginning date of the depressions: FELICIA (18 February 2000), DT7 (30 January 2001), GERARD (29 January 2005), HENNIE (19 March 2005), and DIWA (2 March 2006) at 00:00 UTC. In every case, the center of the low-level depression is indicated by the first letter of its name.

cyclones, convection is not localized near the center of the system. It does not seem to play a predominant role on vortex displacement.

5. Stratosphere-to-Troposphere Exchanges: Several Different Mechanisms

[29] Section 4 showed some different convective features which can be considered as MDs in the SWIO. Previous works have demonstrated that the tropical convection can induce stratospheric filaments in the upper troposphere by wind shear in the peripheral parts of the convection zones. For example, *Leclair de Bellevue et al.* [2006] have shown some signatures with large values of potential vorticity and ozone coming from the stratosphere near tropical cyclones and depressions. These stratospheric intrusions appear frequently by interaction between convection-induced upper-level circulation, jet-front systems and Rossby wave breaking [*Leclair de Bellevue et al.*, 2006]. To evaluate if a MD can induce such vertical exchanges, we have examined some profiles of ozone obtained close to these events by ozonesondes and tropospheric ozone lidar [*Baray et al.*, 2006], and the distributions of potential vorticity in the upper troposphere (250 hPa isobaric level), using reanalysis with 1 degree of horizontal resolution (ERA-40 before 2001 and operational ECMWF model after 2001). This resolution is sometimes too smoothed to highlight mesoscale structures; therefore, we also calculated vertical cross sections of potential vorticity with 48 h advection using the Reverse Domain Filling (RDF) trajectory mapping technique. This approach consists in running backtrajectories with a regular gridded array, then in interpolating the tracer to backtrajectory points, and finally in copying values forward to regular grid. This approach allows us to improve the vertical resolution and to retrieve subgrid information which was not in the initial field [*Dritschel*, 1988]. We applied this technique using the code LACYTRAJ previously used to evaluate the climatological influence of stratosphere–troposphere exchange on ozone profiles database of Reunion Island [*Clain et al.*, 2010].

[30] Regarding the FELICIA case (Figure 9), a filament of stratospheric air is also formed around an upper-level anticyclone located around the point 55°E, 40°S on 19 February 2000. The cross section of advected potential vorticity shows

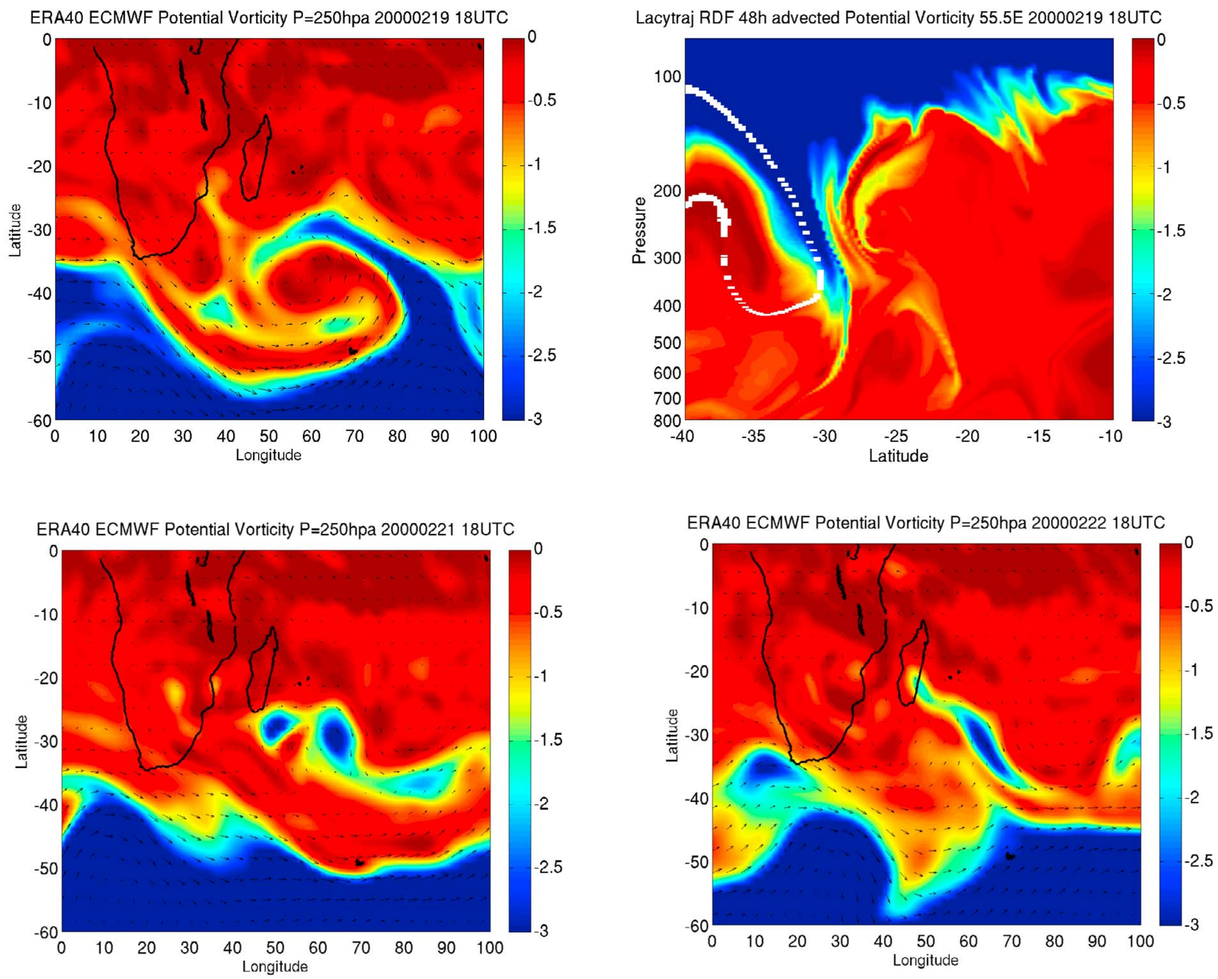

Figure 9. Dynamical fields corresponding to FELICIA. (top left) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 19 February 2000 18:00 UTC. (top right) Vertical cross section of potential vorticity at 55.5°E advected during 48 hours with LACYTRAJ on 19 February 2000 18:00 UTC. (bottom left) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 21 February 2000 18:00 UTC. (bottom right) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 22 February 2000 18:00 UTC.

that the filament reaches the 500 hPa pressure level. During the next days, the edge of the filament forms a stratospheric air mass moving northwestward. This air mass is close to Reunion Island on 22 February 2000. Although located close enough to the subtropical compartment, the synoptic configuration of this intrusion resembles those of a cutoff system.

[31] In the DT7 case (Figure 10), the interaction of two upper-level anticyclones will induce a stratospheric intrusion into the troposphere. This intrusion becomes isolated during the evolution of DT7 and will also form a cutoff low that seems quite similar to another case previously studied in the region [*Baray et al.*, 2003]. At the end of the DT7 lifecycle, the cutoff low stretches and dissipates into the troposphere. The cross section of advected potential vorticity indicates that this event has reached the 500 hPa pressure level (8 km). The ozone profile obtained by lidar on 4 February 2001 shows two ozone peaks at 7 and 12 km. Reunion Island was in the surrounding area of the cutoff low. It is possible that these peaks (especially at 12 km) have been induced by diffusion of the cutoff low. If an ozone measurement could have been performed inside the cutoff low, the amplitude of the ozone peak would have been larger, as in the case reported by *Baray et al.* [2003].

[32] Regarding the GERARD case (Figure 11), a stratospheric filament is also present in the upper troposphere from the early stages of its development. This filament is located between two upper-level anticyclones, one located over Africa near 20°East and another associated to GERARD near 60°East. The interaction of these two anticyclones results in an important wind shear area causing the intrusion of stratospheric air into the troposphere. The intensification of the system in a moderate tropical storm, and then in a strong tropical storm increase the intrusion which is very deep on 4 February 2005 at 18:00 UTC, between 30° and 35°South. Reunion Island is located northeast of the intrusion, the ozone profile obtained by lidar on 2 February 2005 is not affected by this event.

[33] In the HENNIE case, a stratospheric filament is visible on the 250 hPa pressure level at 30°South on 21 March 2005 (Figure 12) and a Rossby wave formed by the subtropical

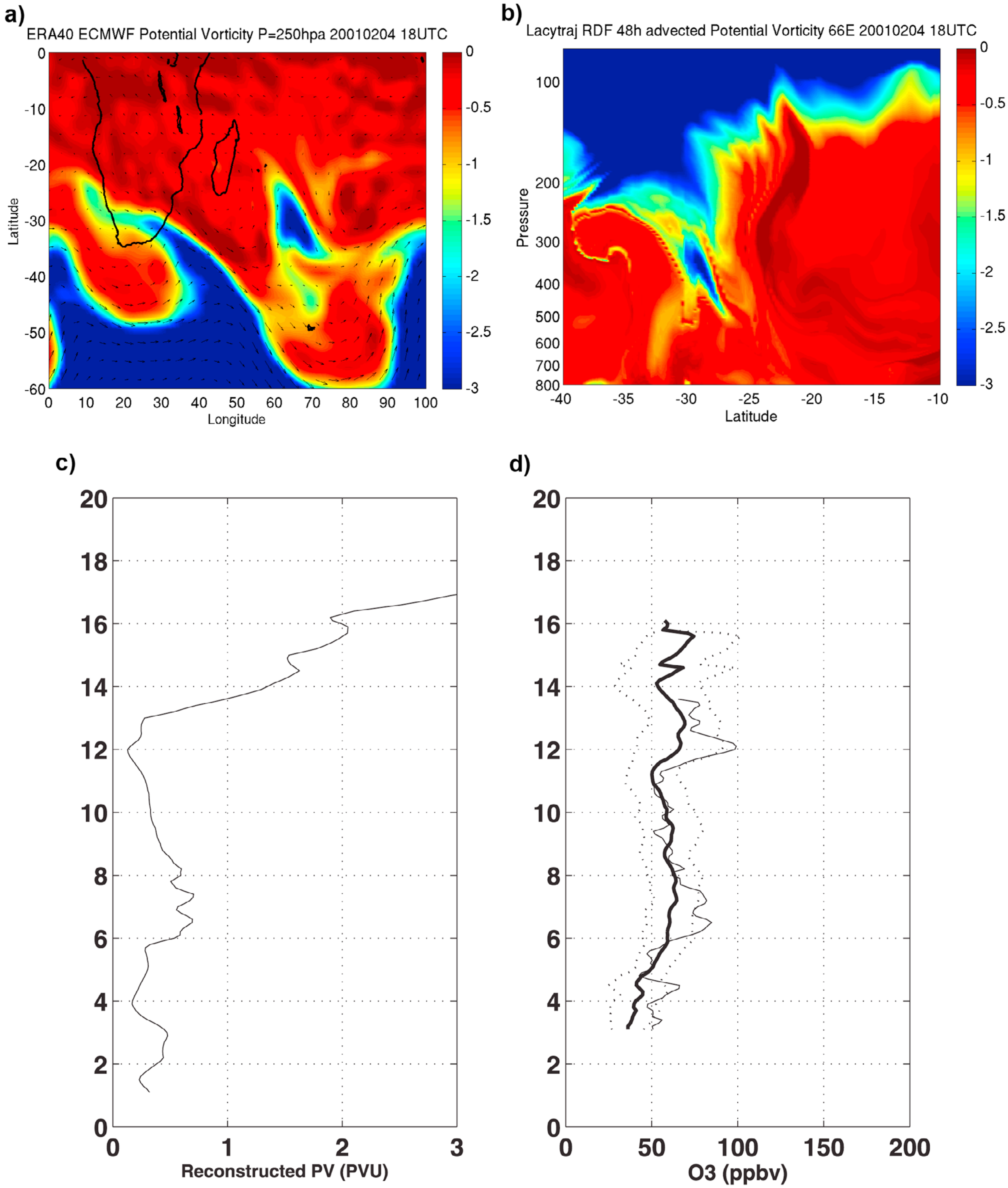

Figure 10. Dynamical fields and measurements corresponding to DT7. (a) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 4 February 2001 18:00 UTC. (b) Vertical cross section of potential vorticity at 66°E advected during 48 hours with LACYTRAJ on 4 February 2001 18:00 UTC. (c) Profile of absolute value of potential vorticity over Reunion Island (21°S, 55.5°E) advected during 48 hours with LACYTRAJ on 4 February 2001 06:00 UTC. (d) Ozone profile obtained by lidar on 4 February 2001 at Reunion Island, compared with the seasonal climatological profile for this site.

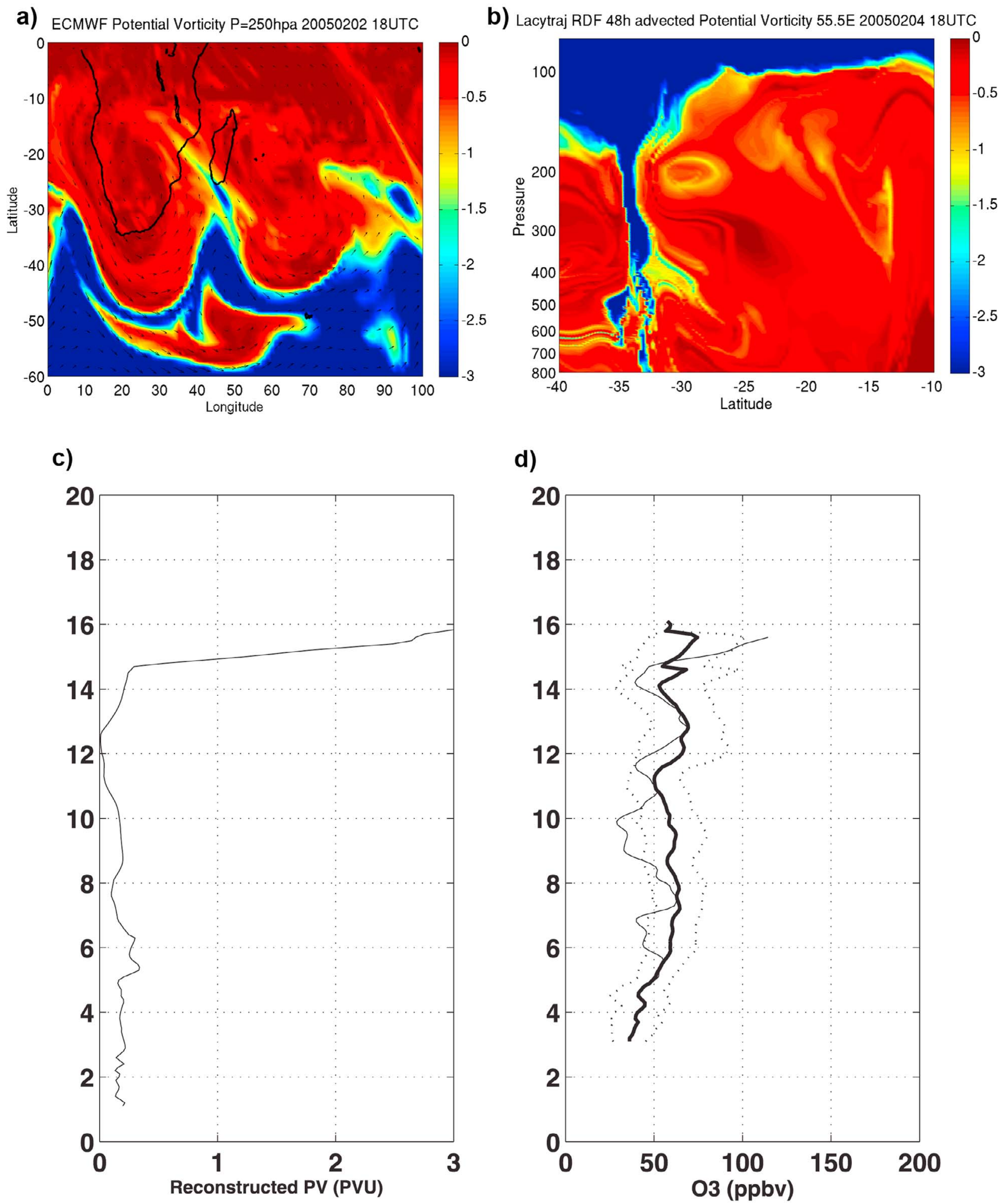

Figure 11. Dynamical fields and measurements corresponding to GERARD. (a) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 2 February 2005 18:00 UTC. (b) Vertical cross section of potential vorticity at 55.5°E advected during 48 hours with LACYTRAJ on 4 February 2005 18:00 UTC (c) Profile of absolute value of potential vorticity over Reunion Island (21°S, 55.5°E) advected during 48 hours with LACYTRAJ on 2 February 2005 06:00 UTC. (d) Ozone profile obtained by lidar on 2 February 2005 at Reunion Island (thin solid lines), compared with the seasonal climatological profile (thick solid lines) and standard deviation (dashed lines) for this site.

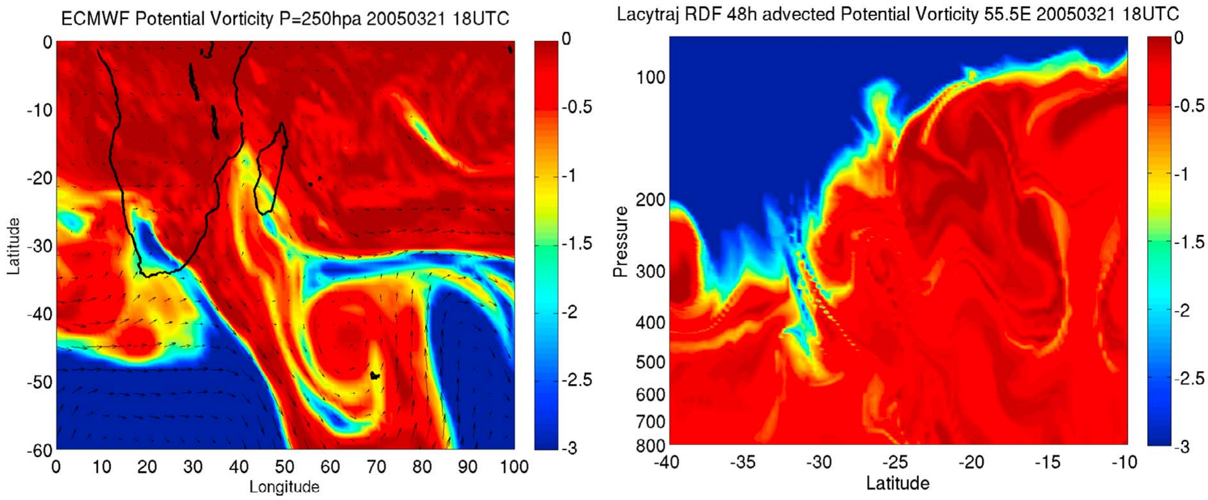

Figure 12. Dynamical fields corresponding to HENNIE. (left) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 21 March 2005 18:00 UTC. (right) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 21 March 2005 18:00 UTC.

jet stream appeared. This filament wraps around the low-pressure area of the Rossby wave centered around the point 65°East, 45°South. Vertically, the filament reached the 500 hPa pressure level (8 km).

[34] The DIWA case (Figure 13), from the upper tropospheric point of view, is quite similar to the HENNIE case. A stratospheric filament is formed in the upper troposphere and is visible at the 250 hPa pressure level on 3 March 2006. The interaction between the upper-level anticyclone centered at 65°East, 40°South and the upper-level divergent circulation associated to DIWA has induced an intrusion of stratospheric air mass into the troposphere. On the vertical cross section of advected potential vorticity on 6 March 2006, the filament reaches the 600 hPa pressure level (6 km). An ozone-sonde measurement has been performed on 8 March 2006, but the filament was too far from Reunion Island, and the signature is not visible on the tropospheric ozone profile (data not shown).

6. Conclusion

[35] In this paper, we have presented a study focused on MDs, a subcategory of tropical depressions, in the Southwest Indian Ocean basin. After having established the definitions and detection criteria of MD, the period 2000–2008 has been investigated and in this region, five events of MD have been identified amongst a total of about one hundred classical tropical depressions. The study of the synoptic situations of these five cases showed that this subcategory of tropical

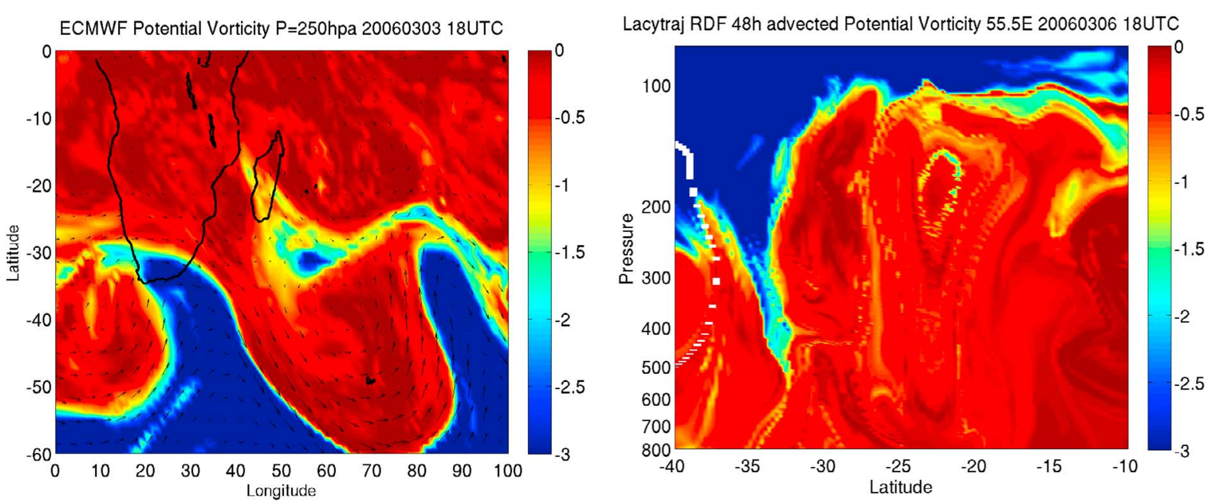

Figure 13. Dynamical fields and measurements corresponding to DIWA. (left) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 2 March 2006 18:00 UTC. (right) Horizontal potential vorticity and wind fields from ECMWF on the 250 hPa pressure level on 6 March 2006 18:00 UTC.

depressions presents some specificity with regards to the MD in the NIO that are well described in the literature. Large-scale ascending is mostly concentrated in the eastern part of MD, when the general direction is toward the Southwest, suggesting that vorticity budget is not the same as in the NIO. Moreover, some phases of development are quite similar to the ones of classical tropical storms in the SWIO. The coverage of SWIO by oceanic surface (and absence of large-scale orography) and the existence of trade winds may explain the differences between MDs in the SWIO and in the NIO.

[36] Potential vorticity distributions have revealed some stratospheric air masses in the troposphere for the five cases. The stratosphere–troposphere exchanges involve different mechanisms: formation of a filament by wind shear between the upper level circulation of the MD and an anticyclone (GERARD), by interaction between the upper-level circulation of the MD and a jet-front system forming a Rossby wave breaking (DIWA, HENNIE), or the formation of a cutoff lows event (DT7, FELICIA). It seems that the dynamic structure of MDs, with particularly strong winds in the periphery, favors the weakening of the tropopause and the induction of stratosphere–troposphere exchange. However, ozone measurements have not been performed inside the filaments and ozone peaks have not been observed in the five cases studies.

[37] It is important to continue to study the dynamics of MD, in the scope of understanding the dynamics and meteorology, and evaluating their impact on the tropospheric ozone budget. All the results presented in this study depend on the quality of ECMWF analyses. More conclusive results could be from the future with specific campaigns including airborne measurements such as the future experiment SWICE (Southwest Indian Ocean tropical Cyclone Experiment) which will be organized in 2011 or 2012.

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