Unique relationship between tropical rainfall and SST to the north of the Mozambique Channel in boreal winter

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Abstract

In this study, we investigate a possible mechanism for the dichotomy in climatology of marine rainfall and sea surface temperature (SST) over a part of the southwestern Indian Ocean (SWIO) during boreal winter (January and February) with state-of-the-art satellite and reanalysis datasets. Rainfall to the north of the Mozambique Channel, bounded 10°S-5°S and 40°E-50°E, is found to be quite feeble despite being in the warm sea surface temperature (SST) regime of up to 29-29.5 °C. The rainfall intensity is rather found to be highly associated with the atmospheric surface divergence. The vigorous rainfall is associated with the more convergence over the Inter-tropical Convergence Zone (ITCZ), while the weak rainfall is linked with the divergence to the north of the Mozambique Channel. The surface divergent flow to the north of the Mozambique Channel is associated with a deep southward penetration of the northerly Indian Winter Monsoon (IWM). Corresponding to the surface divergent field, a relatively high sea level pressure (SLP) compared to the SLP in the ITCZ, weak subsidence, and low-level stratiform clouds are formed to the north of the Mozambique Channel, despite the warm, tropical SST. These atmospheric conditions are most likely conductive to the inhibition of cumulus convection over the region and the unique relationship between rainfall and SST seems peculiar. Our analysis also shows that the rare occurrence of tropical cyclones over the area is attributed to a high-pressure surge and the associated positive surface vorticity (anti-cyclonic). This study suggests that the area to the north of the Mozambique Channel is dynamically interesting for climatological studies.

1. Introduction

The Islands and territories of the Southwest Indian Ocean (SWIO) are often facing damages from weather-induced disasters, such as tropical cyclones and floods (e.g., du Plessis, 2012; Reason and Keibel 2004; Malherbe et al., 2012; Woodruff et al., 2013). The tropical cyclones forming over the SWIO are approximately 14% of the global total tropical cyclones (e.g., Mavume et al., 2013). The severe weather effects are most likely exacerbated by the effects of climate change. The effect of climate change on the regional characteristic of the SWIO rainfall is unclear but is likely important given the large precipitation totals in the boreal winter season (January and February) (e.g., Jury 2016).

The SWIO is dominated by the cross-equatorial northeasterly/northwesterly Indian Winter Monsoon (IWM) flow originating from the Indian Subcontinent. This basin-scale monsoon flow forces an ocean monsoonal circulation system in the Indian Ocean (e.g., Schott and McCreary, 2001; Schott et al., 2009; Talley et al., 2011). The IWM forms the Intertropical Convergence Zone (ITCZ) or monsoon trough over the SWIO region by colliding against the southeasterly/easterly trade winds associated with the Mascarene High over the southern Indian Ocean. Deep cumulus convection occurs frequently (e.g., Roca et al., 2002) and tropical cyclones and monsoon depressions are triggered in the ITCZ over the SWIO due to the underlying warm sea surface and low-level atmospheric convergent flow (e.g., Jury, 1993; Waliser et al., 1993; Klinman and Reason, 2008; Fauchereau et al., 2009; Baray et al., 2010).

The ITCZ over the SWIO is connected to a rainfall belt associated with the Tropical Temperature Troughs (TTT, e.g., Macron et al., 2014) over the southern African Continent through the Mozambique Channel and Madagascar as shown

Fig.1a. Jury (2016) has investigated the austral summer climate (December-to-March) over Madagascar comprehensively and concluded the following : rainfall activity in the December-to-February period over Madagascar is positively well-correlated with the IWM and the cyclonic circulation over the Mozambique Channel. The diurnal cycle and high-elevated topography (up to 1700 m) causes more rainfall interacting with these background winds over Madagascar. Macron et al. (2016) showed a connection among Madagascar rainfall intra-seasonal variability, the MJO and TTT in austral summer season. Reason (2007) suggested that a cyclonic anomaly can be a favourable condition for the development of the tropical cyclone Dera (initiated over the Mozambique Channel) that caused the severe flooding disaster over Mozambique during 9-11 March in 2001.

Referring to Fig. 1a, there is a latitudinal discontinuity of the rainfall belt associated with the ITCZ over the SWIO, which is as follows: between 30°E and 50°E, the centre of the rainband tilts in northwest-southeast direction with a small angle. While the rainband becomes weakened slightly over the southern part of the Mozambique Channel (30°S-20°S and 35°E-40°E), the vigorous rainfall sits over the northern Mozambique Channel towards Madagascar (20°S-15°S). Along the eastern coast of Madagascar, cumulus convection is still highly vigorous, which is also associated with a diurnal variation of land breeze circulation (Jury 2016) and interaction with the easterly trade winds. To the east of Madagascar, the rainband core jumps suddenly up to 7.5°S eastward over the SWIO. On the other hand, there is an area where rainfall activity is weak (1-4 mm/day) at the northern entrance of the Mozambique Channel (10°S-5°S and 40°E-50°E) and over the subtropical SWIO to the east of Madagascar (20°S-15°S and 52°E-80°E). In particular, the area to the north of the Mozambique Channel is located at the same tropical latitude as the ITCZ over the SWIO in Fig. 1a, while the eastern Madagascar is almost in the subtropical zone.

There are, in general, few studies on the boreal winter (January and February) rainfall climatology and associated dynamical processes over the Madagascar and Mozambique regions (e.g., Matyas 2015). Furthermore, the western and northwestern areas of Madagascar are less investigated than the eastern Madagascar. Hence, further investigation on the boreal winter rainfall climatology and its dynamical perspective is important because of the following (1) despite being in the tropical ocean region, the area to the north of the Mozambique Channel is relatively dry and (2) there is a latitudinal discontinuity of the ITCZ during boreal winter, and such discontinuity in the ITCZ has not been reported elsewhere in the tropics. This study investigates the boreal winter rainfall associated with the IWM, focusing on these two aspects.

The rest of this paper is constructed as follows. Section 2 provides the details of datasets utilized in this work. We will describe climatological states to the north of the Mozambique Channel and build a relationship between the rainfall and other key atmospheric variables over the region in Section 3. Finally, Section 4 will summarize the results of analysis with a discussion.

2. Data and Methodology

In this study, we use various datasets of state-of-the-art satellite observational and reanalysis products. The data length was chosen based on the availability. For the satellite observations, the 3-hourly Tropical Rainfall Measuring Mission (TRMM-3B42, Huffman et al., 2007) of rainfall for 1998-2012, the daily Optimum Interpolated Sea Surface Temperature (OISST, Reynolds et al., 2002) of SST for

1982-2012, the daily QuikSCAT (Mears et al., 1999) of surface wind over the ocean for 2000-2008, the International Satellite Cloud Climatology Project (ISCCP, Schiffer and Rossow, 1983) of low-level cloud fraction for 1983-2000 are utilized. We investigate the climatological mean of rainfall, surface wind, SST, and clouds and their relationships over the SWIO in boreal winter. Additionally, the International Best Track Archive for Climate Stewardship (IBTrACS, Knapp et al., 2010; Levinson et al., 2010) for 1900-2010 will be used for a brief investigation on cyclogenesis over the SWIO.

For the reanalysis, we use the monthly Modern Era Retrospective-analysis for Research and Applications (MERRA, Rienecker et al., 2011) for the investigation of monsoon-related atmospheric fields. The MERRA is strong in the better representing hydrological cycle with data assimilation than the previous products (e.g., Wong et al., 2011; Posselt et al., 2012). In particular, MERRA has improved rainfall and water vapour climatology. The observation and reanalysis products are summarized in Table 1. We investigate on possible mechanism through atmospheric diagnostics utilizing MERRA datasets. We focus on the January-February throughout the paper based on the monthly mean and its climatology. In addition, a lag correlation and regression between daily climatological SST and rainfall will be performed in order to investigate the response of rainfall to the underlying SST.

3. Results: Climatological state of IWM around Madagascar

In this section, we investigate the climatological state to the north of the Mozambique Channel. With comparison to other part of the SWIO, a relationship among rainfall, SST and the other key atmospheric variables will be established in the region.

We begin our analysis with from satellite datasets. Figures 1b and 1c provide satellite-monitored boreal winter (January-February) climatology of surface winds from QuikSCAT and sea surface temperature (SST) from OISST, respectively. With a macroscopic view, the northeasterly winds associated with the IWM prevails from the Indian Subcontinent to the Arabian Sea and the IWM changes its direction to northwesterly after the equator reaching 10°S, as shown in Fig. 1b. The southeasterly or easterly trade wind blows in the south of the domain and reaches around 10°S. The vigorous rainfall is approximately located between these northwesterly and southeasterly winds. On the other hand, the IWM intrudes deeply into the Mozambique Channel down to approximately 20°S. Nassor and Jury (1998) have shown that this monsoon deep penetration activates cumulus convection over Madagascar. Remarkably, the meridional component of the surface wind still remains -6 to -4 m/s to the north of the Mozambique Channel while that shrinks to more than -2 m/s over the SWIO from 50°E to 80°E. This southward deep penetration of the IWM seems to generate the cyclonic circulation with the southeasterly wind associated with the Mascarene High over the Mozambique Channel.

According to Fig. 1c, the SST is quite warm (above 28 °C) everywhere the SWIO and it can be expected that deep cumulus convection tends to be generated frequently here. In fact, one of the warm peaks is located broadly between 60°E and 80°E at 2°S where it is adjacent to the strong rainfall zone over the SWIO (Fig. 1a). Another warm SST is found along the Madagascan western coast and the rainfall is also vigorous there, as shown in Fig. 1a. Interestingly, the SST to the north of the Mozambique Channel, where the rainfall is infrequent or weak (Fig. 1a), is also a warm SST peak (up to 29 °C) in the SWIO. Another feature worth mentioning is that the SST is relatively cool in the western basin of the Arabian Sea and a cold-tongue-like structure is formed along the eastern coast of the Arabian Peninsula to Somalia. This cool SST co-exists with the IWM and a similar co-existence can be seen in other sub-basin of the South China Sea (e.g., Koseki et al., 2013; Thompson et al., 2016). The latent heat flux is relatively stronger along this cool SST in the Arabian Sea (not shown). This high evaporation contributes to the cooling of the SST in the Arabian Sea (e.g., Prasanna Kumar and Prasad, 1996; Schott et al., 2009).

Although the rainfall is slightly weaker, particularly, over the Mozambique Channel, the southward intrusion of the IWM into the Mozambique Channel (wind speed in MERRA is also relatively weaker than QuikSCAT) and the dry area over the warmest underlying sea temperature to the north of the Mozambique Channel is well represented in the MERRA reanalysis (Figs.1d and e). In addition, there is a qualitative agreement between MERRA and QuikSCAT in terms of the surface cyclonic circulation over the Mozambique Channel, shown in Figs.1b and e. The location and latitudinal discontinuity of the ITCZ are also reproduced realistically. On the other hand, the eastern/western coastal rainfall is relatively weak over Madagascar. The coastal rainfall is mainly due to diurnal variation indicating that MERRA may not represent the local sea/land breeze well. Although Fig. 1f shows ocean skin temperature, warm ocean temperature is geographically consistent with the observed SST (see Fig. 1c). A relationship between the tropical marine rainfall and underlying sea water temperature is stated more clearly in Fig. 2. This scatter plot is shown in three different boxes in the following: (i) the ITCZ over the SWIO (Box-A (black), 50°E-80°E and 20°S-5°S, black), (ii) to the north of the Mozambique Channel (Box-B (red), 30°E-50°E and 10°S-5°S, red) and (iii) the Mozambique Channel (Box-

C (blue), 30°E-50°E and 20°S-10°S, blue) only over ocean grids. Figure 2a from satellite observations shows that the rainfall intensity increases monotonically as the SST warms up until approximately 28 °C, as seen in Box-A, and the rainfall appears to be independent of the SST between 28 and 28.5 °C, even though the intensity is still largely high. The modest marine rainfall to the east of Madagascar (c.f. Fig. 1a) is due to a relatively cool SST (c.f. Fig. 1c). This rainfall-SST relationship appears to be consistent with the results and conclusions of previous studies have concluded (e.g., Graham and Barnett, 1987; Waliser et al., 1993; Sabin et al., 2013).

Over Box-C, where the SST is slightly warmer than that in Box-A, the rainfall is still strong and the relationship between rainfall and the SST seems to be the same as that over the ITCZ. Conversely, the relationship in Box-B differs extremely from that in the other two boxes. Although some grids are overlapping with those in the Mozambique Channel (this is because two boxes are connected meridionally, the overlapping scatters may be in a marginal zone between two boxes), there is a main cluster of scatters located in an area of weak-rainfall (approximately 2 mm/day) and warm-SST (29 °C). In particular, consolidating with the scatters of the Mozambique Channel, a width of rainfall variation at 29 °C ranges from approximately 1 mm/day to 16 mm/day, which is wider than the range of rainfall over the ITCZ between 26.5 and 28.5 °C of the SST. Waliser et al. (1993) discussed that the intensity of deep convection drops down after 29.5 °C over the tropical oceans based on satellite observations. Indeed, the SST on some grid cells over the MC exceeds to this SST threshold and the rainfall is somewhat moderate (10 mm/day), although the number of grid cells may not be enough to prove a statistical significance. Sabin et al. (2013) also showed that 29-29.5 °C is a threshold of intense deep cumulus convection and the decreasing of rainfall as warming SST exceeds to the threshold is remarkable especially over the warm pool in the tropical Pacific and Indian Oceans. With respect to discussions by Sabin et al. (2013), our results on the rainfall-SST relationship over Box-B seems to be singular because the rainfall intensity is quite weak despite not exceeding to the SST traditional criteria of 29-29.5 °C. The MERRA also draws this extraordinary relationship between the rainfall and SST to the north of the Mozambique Channel as shown in Fig. 2b, while the rainfall of the MERRA is relatively moderate over the Mozambique Channel compared to that of the observation (Box-C). Another satellite rainfall dataset, TMI (TRMM Microwave Imager, e.g., Gentemann et al., 2010), also illustrated similar singularity between rainfall and SST to the north of the Mozambique Channel (not shown).

The simultaneous relationship suggests that the rainfall activity is explainable by the classical relationship with the underlying SST over the ITCZ (Box-A), but the relationship north of the Mozambique Channel (Box-B) differs from this. On the other hand, it has been concluded that deep cumulus convection continues to be intensified after meeting the criteria of 29-29.5 °C of the SST over tropical oceans (e.g., Wu and Kirtman, 2005; Nair and Rajeev, 2013; Roxy 2014). In particular, Roxy (2014) found that there is a time lag of several days when rainfall responds to the SST in the North Indian Ocean by lag-regression analysis. Here, we perform a lag correlation and regression analysis over Box-A and Box-B and investigate the time lag of rainfall response to the SST in the southwest Indian Ocean. For this analysis, the daily climatology of TRMM and OISST (1998-2012) is used from January 1st to February 28th.

Figure 3 presents plots of lag correlation and regression coefficients between SST and rainfall rate. In Box-A, the highest correlation coefficient of approximately 0.6 is found around minus 10 days. Correspondingly, the precipitation is regressed strongly to SST by a 10 day lag. This indicates that precipitation over the ITCZ is enhanced by the warm SST after 10-day. This result is consistent with results by Roxy (2014) for over the North Indian Ocean during the Indian Summer Monsoon. Roxy (2014) concluded that the SST-regressed precipitation increases monotonically after the traditional threshold of 29-29.5 °C. Our result also suggests that such a monotonic increase in precipitation with SST can be detected over the Southwest Indian Ocean during the boreal winter. However, based on satellite data of OISST, climatological daily SST in Box-A rarely exceeds this criterion of SST during January to February (not shown), while SST warmer than the criteria is observed frequently in the North Indian Ocean (Roxy 2014).

On the other hand, the lag correlation is quite small for the whole of lagged time period in Box-B, while relatively high correlation is seen around minus 5 days (but still smaller than 0.2 which is not statistically significant). The lag regression coefficient reaches 2.0 mm/day/°C, which is comparable with the results of Roxy (2014). However, this high value of regression is induced from the small daily variability of SST (not shown). Since the correlation coefficient is insignificant in this context, so is the regression coefficient. This small lag-correlation suggests that rainfall is not sensitive to the underlying SST to the north of the Mozambique Channel.

Figures 4a shows a surface atmospheric divergence obtained from satellite observation. A strong convergence is located over the ITCZ where the intense rainfall is generated (see Fig. 1a). Additionally, there is a relatively strong convergence over the Mozambique Channel. These convergent zones are well consistent with the intense rainfall (Figs.1a and d). On the other hand, the divergent surface flow is dominant to the north of the Mozambique Channel, elongating from the Arabian Sea along the east African coast. In according to another scatter plot between rainfall and surface divergence (Figs. 4c), the rainfall over the SWIO is highly related to the surface divergence as follows: the vigorous rainfall is over the surface convergence (Box-A) and weak rainfall concentrates over the divergence (Box-B). Over the Mozambique Channel, the relationship between rainfall and divergence seems to be weaker than the other two regions, although a relationship of strong rainfall and convergence is still seen. Over the ITCZ (Box-A), the precipitation seems strongly dependent on both the underlying SST and surface divergence. This result may suggest the three-way relationship among precipitation, SST and divergence suggested by Lau et al. (1997) and Roxy et al. (2013). On the other hand, the precipitation is not dependent on the warm SST, but only on the surface divergence to the north of the Mozambique Channel (Box-B) indicating that the three-way relationship is not applied to this region. In the three-way relationship, the warm SST plays a role in affecting the atmospheric circulation. However, our analysis suggests that the underlying SST does not influence the above atmosphere to the north of the Mozambique Channel. This suggestion is supported by the lagged analysis shown in Fig.3.

The MERRA also captures the relationship between the rainfall and the surface divergence shown in Figs. 4b and d although a range of surface divergence is relatively narrow. In particular, the southward intrusion of the divergence into the Mozambique Channel is well represented (Fig. 4b). Therefore, we mainly focus on the MERRA to survey what induces this unique relationship to the north of the Mozambique Channel, henceforth.

Here, more details of other atmospheric variables over the SWIO are investigated as shown in Fig. 5. The distribution of lower sea level pressure (SLP) appears to be consistent with that of the ITCZ and the Mozambique Channel, shown in Figs.1a and d. Higher SLPs are found in both the northern and southern sides of the domain, indicating the northeasterly monsoon-associated high over the Arabian Sea and the Mascarene High over the subtropical southern Indian Ocean, respectively. It is worth of pointing out that the relatively high SLP spreads along the east African coast and the Arabian Peninsula to the north of the Mozambique Channel and the SLP ridge forms between 40°E and 50°E (note that the SLP interval is exaggerated between 1010 and 1012 hPa in Fig. 5a). The distribution of vertical motion at 500 hPa is consistent roughly with that of the SLP in Fig. 5a. The intense upward motion exists around the ITCZ and the Mozambique Channel with a good agreement with the intense rainfall. Interestingly, a cross-equatorial weak subsidence is detected along the eastern African coast where the relatively high SLP penetrates southward. The weak subsidence still survives in the north of the Mozambique Channel, although the underlying SST is warmest in the SWIO (Figs. 1b and 1e).

Corresponding to the higher SLP and downward motion, a part of the SWIO is covered by low-level clouds due to large-scale condensation process shown in Fig. 5b. One dominant, low-level cloud formation is over the subtropical southern Indian Ocean. This low-level cloud may be associated with the Mascarene High (e.g., Wood 2012). In general, subtropical stratocumulus cloud cover is noted over the subsidence region (e.g., Klein and Hartman, 1993). Another low-level cloud formation is remarked over the southwestern Arabian Sea to the north of the Mozambique Channel along the east coast of Africa. This low-level cloud also co-occurs with the relatively higher SLP along the east African coast elongating from the Arabian Sea (Fig. 5a). On the other hand, the low-level cloud is infrequent over the ITCZ and the Mozambique Channel where deep cumulus convection is supposed to be strong. Supportively, Fig.

5c shows that the low-level cloud is relatively dominant from the Arabian Sea towards the Mozambique Channel in a satellite observation. Because Figs.5b and 5c are different quantities, it does not make sense to argue about the two plots quantitatively. However, their qualitative distributions are roughly identical. Bony et al. (2000) showed a frequent low-level cloud formation over the Arabian Sea and east African coast during January to February with other satellite observations.

A vertical-longitude section also provides another unique characteristics of the north of the Mozambique Channel with respect to those in the ITCZ, shown in Fig. 6a. From the surface up to 900 hPa, the atmospheric boundary layer over the tropical SWIO is highly wet (climatological relative humidity exceeds 85%) everywhere (40°E-80°E), as shown in Fig. 6a. On the other hand, from 850 hPa up to 250 hPa, the atmosphere to the north of the Mozambique Channel (40°E-50°E) is relatively dry and that which is over the ITCZ (50°E-80°E) is wet. The relatively wetter middle-troposphere (up to 600-500 hPa) in the ITCZ indicates that cumulus convection occurs there and condensation occurs quite effectively. The drier middle/upper-troposphere to the north of the Mozambique Channel suggests less cumulus convection and, additionally, that the subsidence transports a drier air-mass from the upper to the lower troposphere because the cooler air, in general, contains less water vapour, based on Clausius-Clapeyron's relation.

This singularity to the north of the Mozambique Channel can be summarized in Fig. 6b. The rainfall and SLP shows a straightforward relationship over the SWIO (less rainfall/higher SLP and more rainfall/lower SLP). Correspondingly, the surface divergence can also explain the rainfall longitudinal variation over the SWIO. On the other hand, the sea skin temperature (a proxy of SST) is warmest between 40°E and 50°E and decreases eastward (although the range of values is small). Even though the warmest temperature does not exceed to the SST-criteria for deep cumulus convection (Waliser et al.,1993; Sabin et al., 2013), the atmospheric boundary layer to the north of the Mozambique Channel bears relatively unfavourable conditions for deep cumulus convection because of the weak subsidence (Fig. 7b) and corresponding divergent flow (Fig. 4) there.

Additionally, we analyse the cyclogenesis of tropical cyclones over the SWIO that can be related to the IWM. The surface relative vorticity has a clear contrast between the 40°-50°E and 50°-80°E longitudinal zones (Figs. 7a and 7b). Associated with the high-pressure surge, the anti-cyclonic vorticity forms along the Somali coast to the north of the Mozambique Channel. Inversely, a cyclonic vorticity is generated over the SWIO and the Mozambique Channel. In general, genesis of tropical cyclones is a function of low-level relative vorticity in addition to Coriolis forcing, underlying SST, vertical wind shear and atmospheric low-level humidity (e.g., Camargo et al., 2007; Matyas, 2015). There is a geographical agreement between convergence and negative vorticity over the ITCZ and the Mozambique Channel and vice versa north of the Mozambique Channel. The cyclogenesis over the SWIO seems to reflect this surface vorticity pattern shown in Fig. 7c as follows: an occurrence of tropical cyclones is largely high over the ITCZ and Mozambique Channel whereas the cyclogenesis is relatively low to the north of the Mozambique Channel. In particular, there is no occurrence in 5°S-10°S and 40°E-45°E even though this area is located over the warm SST. In addition to the positive vorticity, the relative dry middle troposphere (see Fig. 6b) can also contribute to the inhibition of the tropical cyclogenesis to the north of the Mozambique Channel.

4. Discussion and Concluding Remarks

This study has investigated a latitudinal discontinuity of the Indian winter monsoonal ITCZ over the southwest Indian Ocean (SWIO) in January and February using state-of-the-art satellite and reanalysis datasets. Deep cumulus convection, and thus intense rainfall over the SWIO and the Mozambique Channel is due to the interaction of the northeasterly and northwesterly IWM with the southerly trade winds. On the other hand, deep cumulus convection is suppressed strongly over the northern entrance of the Mozambique Channel where the latitude is the same as the ITCZ over the SWIO. Nevertheless, the SST in this region is warmest (29-29.5 °C) in the SWIO. This peculiar relationship of warm SST and extremely weak cumulus convection differs from what previous studies have concluded (e.g., Waliser et al.,1993; Sabin et al., 2013).

Further, it is evident from the lagged correlation analysis (Fig. 3) that rainfall is not sensitive to the underlying warm SST to the north of the Mozambique Channel. Rather the feeble rainfall north of the Mozambique Channel can be explained by the surface divergence (Fig. 4). There seems to be a three-way relationship among warm SST, strong rainfall and surface convergence (e.g., Lau et al.,1997; Roxy et al., 2013) in the ITCZ. Conversely, the north of the Mozambique Channel is only characterized by weak rainfall and surface divergence, which is a two-way relationship. That is, the underlying warm SST does not control cumulus convection in the north of the Mozambique Channel.

The inhibition of deep cumulus convection to the north of the Mozambique Channel can be attributed to the monsoonal high-pressure surge and this is associated with the weak subsidence over the region. Correspondingly, the low-level stratiform cloud forms more frequently from the western Arabian Sea to the north of the Mozambique Channel even over the tropical warm ocean. Co-existence of high-pressure SLP and low-level stratus clouds are, in general, ubiquitously observed in the basin-scale subtropical anti-cyclone systems (e.g., Klein and Hartmann, 1993). Our study reveals that a similar co-occurrence is also detected over the tropical warm ocean. The frequent occurrence of stratus/stratocumulus is probably due to a relatively cool SST in the western Arabian Sea to the Somali coast (Figs. 1b and 1e). Further, the strong latent heat flux is found to be roughly consistent with this cool SST along the Arabian Sea and Somali coast (not shown). This latent heat flux may also enhance the low-level cloud formation in this region. The low-level clouds are often associated with cooling and high SLP features (e.g., Koseki et al., 2012). The high-pressure surge over the north of the Mozambique Channel may also be influenced thermodynamically by the low-level clouds. Coinciding with the surface divergent field, the surface relative vorticity is negative over the ITCZ and Mozambique Channel. The surface relative vorticity is positive to the north of the Mozambique Channel (Fig. 7). These vorticity distributions seem to be related to the cyclonegenesis over the SWIO.

Whereas our present study can conclude that the unusual or unique SST-rainfall relationship to the north of the Mozambique Channel is due to the cross-equatorial monsoonal high-pressure surge into this area, there arises some research questions of interest. One of possibilities is to understand what dynamical thermodynamical processes determine such the IWM horizontal distribution in terms of climatology. For example, The other monsoonal systems are affected by the regional cool SST allowing the monsoon flows to penetrate more deeply (e.g., Okumura and Xie, 2004; Koseki et al., 2013). It can be expected that the cool SST in the Arabian Sea also influence the IWM. Other is to perform a model simulation to understand why the monsoonal high-pressure can survive even over the tropical warm ocean under conditions of low-level stratus cloud formation. These research topics will be taken into account in our future work.

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Figure Captions

Figure 1. Climatology of JF-mean (a) TRMM 3B42 rainfall rate (mm/day) for 1998-2014 period, (b) QuikSCAT surface wind (vector, m/s) for 2000-2008 period and its meridional component (shading, only shown wind less than 3 m/s), (c) OISST sea surface temperature (°C) for 1982-2012 period; and MERRA climatology during 1979-2010 period for (d) rainfall rate (mm/day), (e) 10m wind (m/s, vector) and its meridional component (shading, only shown wind less than 3 m/s), and (f) sea skin temperature (°C). The 3 boxes are regions for scatter plot in Figs.2 and 4.

Figure 2. Scatter plots of JF-mean climatological rainfall versus sea temperature for (a) observation and (b) MERRA over Indian Ocean monsoon trough (box-A, 20°S-5°S and 50°E-80°E), northern entrance of the Mozambique Channel (box-B, 10°S-5°S and 30°E-50°E), and Mozambique Channel (box-C, 20°S-10°S and 30°E-50°E). The plots are only over the ocean grid. The black dashed-line denotes 29.5 °C that is the threshold by Waliser et al. (1993). The box for each region is shown in Fig.1a.

Figure 3. Lag correlation (solid) and regression (dashed) coefficients between daily-mean precipitation and SST over the ITCZ (box-A, 20°S-5°S and 50°E-80°E, shown by black) and northern entrance of Mozambique Channel (box-B, 10°S-5°S and 40°E-50°E, shown by gray). Label on left (right) is for lag correlation (lag regression).

Figure 4. JF-mean climatology of surface divergence for (a) QuikSCAT and (b) MERRA. (c) and (d) same as Fig.4, but for rainfall versus surface divergence for QuikSCAT and MERRA, respectively. For (c), QuikSCAT data is interpolated into MERRA's grid box.

Figure 5. JF-mean climatology of (a) SLP (color) and vertical motion at 500hPa (contour, dashed is negative and solid is positive) and (b) mixing ratio of cloud water due to large-scale condensation at 925 hPa from MERRA in 1979-2010. Note that the color scale is exaggerated between 1010 and 1012 hPa and the contour interval in (a) is 0.01 and 0.005 Pa/s for negative and positive values, respectively. (c) JF-mean climatology of low-level cloud fraction between 1000 and 680 hPa obtained from ISCCP in 1983-1999.

Figure 6. Pressure-longitude section of (a) JF-mean climatological relative humidity averaged between 10°S and 5°S. (b) Latitude-averaged (10°S-5°S) plots of sea level pressure (solid), rainfall (dashed), skin temperature (dot), and surface divergence (solid with triangle marker). All plots are from MERRA.

Figure 7. JF-mean climatology of surface relative vorticity for (a) QuikSCAT (2000-2008) and (b) MERRA (1979-2010). (c) JF cyclongenesis over the SWIO estimated form IBTrACS in 1900-2010. Only the initial location of each tropical cyclone is binned into 2°×2° grid.

Table 1. A detailed list of data sets used in this study.

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