Rain and Cloud Perspectives on the Seasonal Cycle over the Western Equatorial Indian Ocean[✉]

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(Manuscript received 29 January 2019, in final form 19 October 2019)

ABSTRACT

Convection over the western equatorial Indian Ocean (WEIO) is strongly linked to precipitation over Africa and Australia but is poorly represented in current climate models, and its observed seasonal cycle is poorly understood. This study investigates the seasonal cycle of convection in the WEIO through rainfall and cloud measurements. Rainfall shows a single annual peak in early austral summer, but cloud proxies identify convective activity maxima in both boreal and austral summer. These diverging measures of convection during boreal summer are indicative of a reduction in the intensity of precipitation associated with a given cloud regime or cloud-top height during this time of year but an increase in the overall occurrence of high-top clouds and convectively active cloud regimes. The change in precipitation intensity associated with regimes is found to explain most of the changes in total precipitation during the period from May to November, whereas changes in the occurrence of convective regimes explains most of the changes throughout the rest of the year. The reduction in precipitation intensities associated with cloud regimes over the WEIO during boreal summer appears to be related to large-scale monsoon circulations, which suppress convection through forcing air descent in the midtroposphere and increase the apparent occurrence of convectively active cloud regimes through the advection of high-level cloud from monsoon-active areas toward the WEIO region.

1. Introduction

Understanding the climate of, and convection in, the western equatorial Indian Ocean (hereinafter referred to as the WEIO) is important for a range of reasons. Convection over the WEIO affects the strength and duration of October–November monsoonal rains in East Africa (Goddard and Graham 1999; Black et al. 2003), the seasonal cycle of precipitation in East Africa (Yang et al. 2015) and, through Rossby-wave teleconnections, rainfall over southern Australia during spring (Cai et al. 2011). Furthermore, the WEIO is the primary formation location of convection anomalies associated with the Madden–Julian oscillation, a large-scale eastward propagating disturbance which affects tropical convection and weather on 30–60-day time

scales (Madden and Julian 1971; Wang and Rui 1990; Madden and Julian 1994). It is also the location of one pole of the Indian Ocean Dipole, a leading mode of interannual sea surface temperature variability in the Indian Ocean (Saji et al. 1999; Webster et al. 1999).

Despite the role convection in the WEIO plays in the East African and Australian climates, current generation climate and weather models do not represent convection in the region well. As an example, models in phase 5 of the Coupled Model Intercomparison Project (CMIP5) overestimate annual precipitation, with the multimodel mean precipitation error in the region being one of the largest globally (Flato et al. 2014). These models also poorly represent the temporal variability of convection in the region—for example, that associated with the Madden–Julian oscillation (Hung et al. 2013; Moise et al. 2015) and precipitation seasonality (Yang et al. 2014; Dunning et al. 2017). This implies that the ability of seasonal and weather prediction models to predict seasonal precipitation in Australia and East Africa is hampered by their poor representation of tropical convection in the WEIO.

[✉] Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-19-0080.s1>.

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To understand discrepancies between modeled and observed precipitation in the region, as well as to better understand the links between WEIO convection and weather in other regions, an improved understanding of the annual cycle of convection and precipitation in the region is required. This topic has drawn little direct attention as compared with interannual and intraseasonal variability of convection in the WEIO, with only a brief seasonal climatology of ERA-40 having been performed, relating precipitation changes to temperature and wind variations associated with the Indian summer monsoon (Slingo et al. 2005).

Because precipitation, clouds, and convection are all closely linked in the tropics, weather states or cloud regimes derived from cloud properties (Jakob and Tselioudis 2003; Rossow et al. 2005) can also be used to study tropical rainfall (Tan et al. 2013; Rossow et al. 2013; Lee et al. 2013). Previous studies using these regimes show that the weather state representing strongly organized convection accounts around half of the total tropical rainfall despite occurring less than 10% of the time, and the majority of observed weather states are generally nonprecipitating (Tan et al. 2013). Weather-state approaches have also been used on climate models (Tan et al. 2018), with many climate models underestimating the rainfall coming from organized cloud regimes, mainly through underestimation of the rain rate associated with organized clouds and suggests improvements in the cloud–precipitation processes of global climate models are needed.

This paper seeks to further investigate the annual cycle of convection over the WEIO beyond previous studies by investigating clouds, rainfall and weather states; and to answer the following questions: first, do cloud-based convective measures and weather states display the same annual cycle over the WEIO as precipitation does, and do these cycles correspond to annual cycles in vertical motion? Second, is the observed annual cycle in precipitation driven more by changes in the frequency of strongly precipitating weather states, or by changes in the intensity of precipitation when precipitating weather states are present throughout the year? Finally, what elements of the climate system drive the changes in the frequency and/or intensity of precipitation and convective weather states over the WEIO? By answering these questions, this study hopes to provide insight into some of the key processes linking clouds and precipitation in the region.

The following section introduces the data and method used in this paper to determine the annual cycle in WEIO precipitation and analyze the factors driving this cycle. Section 3 presents the key results of the study, showing a disconnect between cloud-based and precipitation-based

measures of convection over the WEIO, and is followed by a summary and conclusions in section 4.

2. Data and method

a. Data

Because of its location over the ocean, there are no suitable direct measurements of convective properties such as vertical motion soundings or surface precipitation available to analyze the convective annual cycle in the WEIO region. Instead, precipitation data used here are estimates from version 1.2 of the Global Precipitation Climatology Project (GPCP) One-Degree Daily (1DD) dataset (Huffman et al. 2001, 2016), which blends rain gauge and satellite data into a global dataset of daily precipitation estimates at 1° resolution in longitude and latitude. This is regridded to a 2.5° grid for comparison with other convective proxies.

The cloud-based convection proxies used in this study are outgoing longwave radiation (OLR) and cloud-top infrared brightness temperature T_b . OLR acts as a cloud proxy as it measures the infrared broadband emission from objects on Earth, with high clouds opaque to infrared radiation constituting regions of low OLR when compared with Earth's surface or near-surface clouds. Measurements of OLR as obtained by satellite-based sensors have been long used as proxies for deep convection in the tropics (e.g., Zhang 1993; Waliser et al. 1993), and the top-of-atmosphere OLR in this paper are obtained from the International Satellite Cloud Climatology Project (ISCCP) flux dataset (Zhang et al. 2004). Brightness temperatures are derived by relating the spectral radiance from satellite measurements at several frequencies of radiation to that from an ideal blackbody, with the brightness temperature being the temperature of the ideal blackbody with the same relationship between radiance and frequency, resulting in T_b being lower when cloud tops are higher in the atmosphere. Brightness temperatures in this study were obtained from the Cloud Archive User Service (CLAUS) dataset (Hodges et al. 2000; Environmental Systems Science Centre 2013).

A cloud regime classification dataset, as described in Tan et al. (2013), is also used in this paper. Unlike OLR or T_b , the cloud regime classification combines two different cloud properties, by utilizing ISCCP D1 dataset cloud-top pressure and optical thickness joint histograms over 280 km × 280 km boxes (Rossow and Schiffer 1999) over an extended tropical band (35°N–35°S) and generates eight regime centroids through a k -means clustering upon daytime averages of the joint histograms, corresponding roughly to convectively active

deep cloud (labeled CD), convectively active cirrus (CC), intermediately active mixed cloud (IM), intermediately active thin cirrus (IC), and four convectively suppressed cloud states (ST, SS1, SS2, and SS3). These centroids are used to assign the daily state of the tropics on a $2.5^\circ \times 2.5^\circ$ grid to a regime through a nearest-neighbor comparison between the grid box histogram and the regime centroids. The regimes generated through this process are practically identical to the eight extended tropics weather states from (Mekonnen and Rossow 2011) and (Oreopoulos and Rossow 2011), and the CD, CC, IM, and IC regimes resemble four of the regimes from earlier analyses of smaller tropical regions (Jakob et al. 2005; Rossow et al. 2005; Jakob and Schumacher 2008). Tan et al. (2013) showed that the cloud regimes are reasonable convective proxies, with the active regimes covering a range of precipitation intensities, OLR values and vertical motions, hence their use in this paper to understand convective states of the WEIO throughout the year. To ensure overlap with GPCP, only 1997–2009 are used from the ISCCP cloud regime data.

Large-scale dynamic and thermodynamic fields in the WEIO are obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011). Quantities have been averaged to daily averages from their original 3-hourly resolution, and where directly compared to cloud regimes, have been regridded to the ISCCP 2.5° grid from the 1.5° native ERA-Interim grid. Fields used include zonal, meridional, and vertical velocities (zonal and meridional in meters per second; vertical in pascals per second) to describe the dynamic state of the WEIO atmosphere, divergence (s^{-1}) to determine areas of outflow in the upper troposphere, and specific humidity (unitless, corresponding to kilograms of water per kilogram of air) to describe some of the thermodynamic properties of the atmosphere.

b. Method

Most of the analysis performed in this paper is based on statistical aggregation of data from the datasets mentioned above. However, the ISCCP cloud regime dataset has some missing values, which hamper direct attempts to compare regime data with other fields. This is because precipitation at a time and location where the corresponding ISCCP regime data are missing cannot be assigned to a regime, and total rainfall attributable to the regimes will not equate to total rainfall. Frequencies of occurrence of regimes are expressed as a fraction of nonmissing measurements, and the fractional contribution to precipitation from

each regime is calculated through dividing precipitation from points in each regime by total precipitation, excluding missing regime points from the total rainfall. This fractional contribution is used to calculate net rainfall attributable to each regime over the complete precipitation dataset, including missing regime values, by multiplying the fractional contribution by the complete total precipitation. This assumes that points where regime classifications are missing have regimes and precipitation characteristics distributed in a similar manner to nonmissing values, which may not be the case.

These frequencies of occurrence and precipitation due to each regime are used to calculate the mean intensity of precipitation associated with each regime, where the mean intensity of precipitation is the precipitation due to the regime divided by its frequency of occurrence.

The domain used to define the WEIO in this paper is from 5°N to 10°S , and from 50° to 70°E . This domain was chosen primarily to encompass a region where model performance is particularly poor, and to encompass the band of peak mean precipitation over the western Indian Ocean while not directly capturing a precipitation signal from the Indian monsoon. However, while the exact phasing and magnitude of the seasonality explored in the rest of study changes based on the domain definition, the general behavior of the seasonality in precipitation and convective proxies is resilient to the definition of the domain, provided enough of the region is located west of 60°E (results for other definitions of the WEIO can be found in the online supplemental material).

3. Results

a. Precipitation and convective proxy seasonality over the WEIO

The seasonal cycle of precipitation, calculated as a monthly field mean of GPCP precipitation data over the WEIO, is presented in Fig. 1a. Mean precipitation over the whole region has a maximum of 5.5 mm day^{-1} during January and a minimum of 3.0 mm day^{-1} in March. From the minimum in March, mean precipitation increases to a slightly higher level of 3.5 mm day^{-1} during May and June, and decreases to a secondary minimum of 3.2 mm day^{-1} during July and August. Precipitation then progressively increases over the remainder of the year until January, apart from a small drop of 0.2 mm day^{-1} from October to November.

However, this precipitation is not evenly distributed spatially over the region, and the spatial distribution

FIG. 1. Climatological GPCP monthly mean precipitation rate (mm day^{-1}) (a) displayed as a seasonal cycle averaged over the WEIO domain and (b)–(m) shown by month over the Indian Ocean region.

of rainfall varies over the year, as seen in Figs. 1b–m. In January and February, precipitation occurs mostly south of the equator, with a gradient of increasing rainfall toward the east of the region, with maximum mean precipitation rates of over 8 mm day^{-1} occurring in the $5^{\circ}\text{--}10^{\circ}\text{S}$, $60^{\circ}\text{--}70^{\circ}\text{E}$ subregion. The east–west gradient weakens in March, but rainfall is still primarily south of the equator. In April, the north–south distribution of rainfall changes and precipitation rates increase from the northwest to the southeast with a weak zonal band of increased rainfall at about 6.5°S , and precipitation in May mostly occurs to the east

of 60°E . From June to August, most rainfall occurs along a tongue extending from the eastern edge of the region to 60°E , and from 0° to 5°S , and from September to November this tongue widens while moving south and west until a northwest–southeast precipitation gradient like that in April is observed in November. December displays strong precipitation south of 2°S , and another band of strengthened precipitation north of 3°N .

In comparison with the annual cycle seen in precipitation, convection proxies based on clouds and cloud-top heights display a different cycle over the WEIO, as

FIG. 2. Climatological monthly mean CLAUS T_b (a) displayed as a seasonal cycle with ISCCP OLR averaged over the WEIO domain and (b)–(m) shown by month over the Indian Ocean region.

can be seen in Fig. 2. The seasonal cycle of area-mean outgoing longwave radiation, where lower values are associated with higher, colder clouds and often interpreted as enhanced convection, is strongly semi-annual, with OLR minima of 237.8 and 236.6 W m^{-2} occurring during January and July respectively. The OLR maxima occur during March and October, with the March OLR maxima of 255.8 W m^{-2} being around 6 W m^{-2} greater than the October OLR maxima of 249.8 W m^{-2} . Similarly, CLAUS cloud brightness

temperature has minima in January and July of 278.3 and 277.6 K respectively, and maxima in April and October of 286.1 and 283.8 K . The annual cycle of these measures suggests that increased high cloud occurs over the WEIO during boreal summer, contrary to what is seen in the precipitation annual cycle in Fig. 1.

Cloud-top brightness temperature and outgoing longwave radiation depend on the height of clouds and its value quickly saturates with increasing cloud

FIG. 3. Climatological monthly frequency of occurrence of CD, CC, IM, and IC ISCCP-derived cloud regimes over the WEIO region, and the combined frequency of occurrence of the CD/CC regimes (label C-types), IM/IC regimes (I-types) and ST/SS1/SS2/SS3 regimes (S-types).

optical thickness, and thus can be related to deep convection, or instead be representative of cirrus. Out of the cloud regimes used by Tan et al. (2013), which incorporates cloud thickness as well as cloud-top-pressure information, the two regimes labeled CC and CD are the most convectively active, the regimes labeled IM and IC are associated with weak convection, and the remaining ST, SS1, SS2 and SS3 regimes are associated with suppressed convection. Assuming that the majority of convection and rainfall is associated with the more convectively active CD and CC regimes, the frequency of occurrence (foc) of these regimes (hereinafter C-type regimes) can be used as a

proxy of deep convection, and the foc of the C-type as well as the IM and IC regimes can be used as a proxy of total convective activity. The annual cycle of C-, I-, and S-type regimes are presented in Fig. 3. The foc of C-type regimes matches the cycle in cloud-top heights, with peaks in January and July of 25% and 29%, respectively, and minima in April and October/November, with foci of 11% and 16%, respectively.

Vertical motion in the WEIO region has a seasonal cycle as displayed in Fig. 4. At 950 hPa, within the boundary layer near the surface, upward motion is at its lowest magnitude of $-8 \times 10^{-4} \text{ Pa s}^{-1}$ in March and April, and has its highest magnitude of $-1 \times 10^{-2} \text{ Pa s}^{-1}$ during December and January. At 350 hPa, a level where strong upward vertical velocity is observed in profiles associated with active deep tropical convection (Lin and Johnson 1996; Xie et al. 2010), upward motion reaches a maximum of $-2 \times 10^{-2} \text{ Pa s}^{-1}$ during January, and downward motion is observed during March, August, and September. Higher in the troposphere, the vertical motion cycle becomes semiannual over the WEIO, with maximum upward motion in January and July of -1.6×10^{-2} and $-1.2 \times 10^{-2} \text{ Pa s}^{-1}$ respectively, a local minimum in March, and downward motion in September.

The annual cycle analysis presented so far suggests that cloud properties do not vary in the same way as precipitation does throughout the year. OLR and

FIG. 4. Climatological monthly mean vertical velocity over the WEIO region (Pa s^{-1}) from ERA-Interim.

FIG. 5. A 2D histogram of the frequency of occurrence of all WEIO data points in the OLR–precipitation dataspace, as well as histograms of (top) OLR and (right) precipitation.

C-type regime occurrence suggest that there is increased convection over the WEIO during boreal summer, however, midtropospheric vertical velocity and precipitation suggest that this part of the year corresponds to reduced convection relative to the annual mean. The disconnect between the annual cycles of

cloud-based convective proxies and precipitation in the WEIO will be explored further in the next section.

b. Relationships between precipitation and convective proxies over the seasonal cycle

The relationship between precipitation and OLR in the WEIO region over the entire period 1997–2009 can be seen in the joint histogram shown in Fig. 5. As expected, lower OLR values (and thus higher clouds) are associated with higher precipitation rates, and higher OLR values are associated with lower precipitation rates. Precipitation rates in the lowest bin (below 2 mm day^{-1}) occur in over 60% of the data points, and OLR values above 258 W m^{-2} occur in over one-half of the data points. Furthermore, for any given OLR value above 240 W m^{-2} , precipitation is more likely to be below 2 mm day^{-1} than above 2 mm day^{-1} .

Over the WEIO, although the general shape of the OLR–precipitation histogram does not change during the year, its position in OLR–precipitation space moves, as can be seen for the months of July and October in Fig. 6 (similar plots for January and March can be found in the online supplemental material). In July, the histogram moves toward lower precipitation and OLR values relative to the annual histogram—that is, for a given OLR the distribution of precipitation moves toward lower precipitation rates, and for a given precipitation rate, the OLR distribution moves toward lower OLR. The combined effect of this movement is that the

FIG. 6. Differences between WEIO frequency of occurrence histograms in OLR–precipitation space for (left) July and (right) October and that for the whole year, similar to Fig. 5.

occurrence of OLR values above and below 278 W m^{-2} is reduced and increased, respectively, whereas the occurrence of precipitation rates above and below 4 mm day^{-1} is reduced and increased, respectively. However, in October the histogram is located toward higher precipitation and OLR values relative to the annual histogram, leading to a reduction in the occurrence of precipitation rates below 8 mm day^{-1} and OLR values below 270 W m^{-2} . These plots suggest that clouds with similar cloud-top heights in the WEIO region lead to different surface rainfall rates depending on the time of year, and that this change occurs over the full range of OLR values apart from the very highest bin. Note that in January and March (the extreme months in monthly mean precipitation) the histogram does not move in OLR–precipitation space relative to the annual histogram but the skew of the histogram changes, tending toward low OLR/high precipitation in January and high OLR/low precipitation in March.

Because the annual cycle in the frequency of occurrence of more convective cloud regimes (in Fig. 3) closely resembles that of OLR, we wish to determine whether similar changes in the relationship between these regimes and precipitation occurs during the year, and to what extent precipitation intensity changes and regime frequency changes explain the observed changes in precipitation throughout the year.

To do this, mean monthly rainfall $\bar{p}$ can be decomposed into precipitation due to each cloud regime, which can then in turn be expressed by the product of the mean intensity of precipitation associated with each cloud regime I and the mean frequency of occurrence of each regime f :

$$\bar{p}(\text{month}) = \sum_{\text{regime}} [\bar{I}_{\text{regime}}(\text{month}) \bar{f}_{\text{regime}}(\text{month})],$$

with the mean intensity being calculated from the precipitation rate over each grid point and time step at which a given regime is present. Figure 7 shows the precipitation intensity for the CD, CC, IM, and IC regimes (the S-type regimes are not shown because of their very low associated precipitation). For every month, the CD regime is associated with the most intense precipitation, followed by the CC, IM, and IC regimes in that order. This is consistent with the relationship between ISCCP regimes and precipitation found in previous studies (Tan et al. 2013; Lee et al. 2013). However, noticeable in this figure is that these regimes all display their lowest intensity during June/July, and have their highest intensity during October/November for all regimes excluding IC, which has its highest intensity in December. The differences between the lowest intensity and highest intensity are quite sizable with the highest

FIG. 7. Annual cycle of mean precipitation intensity associated with the CD, CC, IM, and IC cloud regimes over the WEIO region.

intensities being more than twice the lowest for the CC, IM, and IC regimes, and being a third greater than the lowest intensity for the CD regime.

To determine how the changes in regime intensity seen in Fig. 7 and the changes in regime frequency seen in Fig. 3 lead to the observed precipitation changes, we can decompose precipitation changes from the preceding month similar to the way precipitation is decomposed:

$$\Delta \bar{p} = \sum_{\text{regime}} (\bar{I}_{\text{regime}} \Delta \bar{f}_{\text{regime}} + \bar{f}_{\text{regime}} \Delta \bar{I}_{\text{regime}} + \Delta \bar{I}_{\text{regime}} \Delta \bar{f}_{\text{regime}}).$$

The three terms on the right side of the equation above correspond to changes in precipitation due to regime frequency changes, changes due to regime intensity changes, and a cross term that is due to changes in both. In the case in which there is no change in the rainfall intensity related to convective regimes during the year, all monthly rainfall changes are due to changes in the occurrence of the regimes and would appear in the first term; conversely, if the occurrence of regimes remains constant throughout the year then rainfall changes are solely due to intensity changes and appear in the second term.

Using the mean regime precipitation intensities for each month from Fig. 7, along with the mean frequency of occurrence for the regimes as displayed in Fig. 3, the rainfall change from month to month for the CD, CC, IM, and IC regimes are decomposed into these three terms and presented in Fig. 8, along with the sum of each term over all regimes. From the precipitation maximum in January, the decrease in monthly mean precipitation to the minimum in March is primarily due to the decrease in the frequency of the convective regimes, with a smaller amount of the decrease being attributable to a decrease in the intensity of precipitation during the

FIG. 8. (top) Change in mean rainfall from previous month attributable to changes in the frequency of ISCCP cloud regimes ($I\Delta F$), (top middle) changes in the intensity of precipitation associated with ISCCP cloud regimes ($F\Delta I$), (bottom middle) cross terms from changes in both ($\Delta F\Delta I$), and (bottom) the total change in mean rainfall from the previous month for all rainfall as well as that attributable to the CD, CC, IM, and IC regimes.

CC and IC regimes. The small increase in precipitation in April can be attributed to an increase in the frequency and intensity of the IC regime, and the precipitation increase in May can be attributed to an increase in the occurrence of CC regimes and a combined increase in the frequency and intensity of the CD regime.

As also seen in Fig. 3, in June there is a large increase in the frequency of the C-type regimes as well as the IM regime. This would lead to an increase in precipitation during June of over 1 mm day^{-1} , however the decrease in regime intensities between May and June leads to

precipitation decreases, and coupled with the cross term, entirely counteracts the precipitation increase from the increase in regime occurrence, with increased precipitation from the CD, CC, and IM regimes countered by the reduction in IC regime precipitation. The intensity of the regimes decreases further for the CC and IC regimes in July, leading to the reduction in precipitation.

Conversely, over the period from August to October, the frequency of the convective regimes (specifically the CC regime) decreases, however the intensity of the four

convectively active regimes increases. The net effect is that precipitation from the CD, IM, and IC regimes increase over the period and precipitation from the CC regime remains roughly constant, leading to a net increase in precipitation. The small dip in precipitation in November can be associated with a small decrease in the intensity of CD and CC regimes, and the increase in precipitation from November to January is predominantly due to the increase in the CD and CC regime frequency.

We have now established that over the WEIO from May to July, there is a reduction in the level of precipitation associated with both given OLR values and with ISCCP cloud regimes while the occurrence of lower OLR values and C-type regimes increases, leading to the apparent disagreement between convective activity from these cloud proxies and that from precipitation. From August to October, OLR values increase and the frequency of CC and IC regimes decreases, while the intensity of precipitation increases for all the convective regimes such that total precipitation increase. The remaining question is what leads to the changes in precipitation and OLR/regime frequency in the period from December to April without large changes in the precipitation intensity, and what drives the differing changes in regime frequency and intensity from May to July and from August to October.

c. Factors behind regime precipitation intensity and frequency changes

The increase in precipitation from November to January, and the decrease from then until April appears to be due to the establishment and decay of the intertropical convergence zone (ITCZ) over the southern region of the WEIO domain during this time, which is observed in cloudiness (Meenu et al. 2007) and 1000–850-hPa convergence (Berry and Reeder 2014). As the ITCZ strengthens deep convection in the area becomes more likely until January. January also corresponds to the maximum occurrence of both CD and IM regimes, suggesting that optically thick clouds with cloud tops above 650 hPa are most common over this part of the year, as would be expected from strong ITCZ surface convergence.

The large increase in low OLR values/high cloud-top regimes during April–July without an increase in precipitation, but instead a decrease in precipitation intensity when these OLR values or regimes are present suggest an increase in the level of high, nonprecipitating cloud during this time of year. In combination with mean descent of air between 300 and 850 hPa over the WEIO during boreal summer shown in Fig. 4, this suggests that the upper-level cloud is not related to convective activity

within the WEIO region and deep moist convection is instead suppressed within the WEIO. Two questions arise from this point: What is the source of the upper-level cloud? What circulation is responsible for the descending air in the midtroposphere and suppression of deep convection over the WEIO during boreal summer?

While WEIO convection is weak during boreal summer, this is not the case in nearby tropical areas where the Indian and Southeast Asian summer monsoon is active, with strong convection occurring over the Indian subcontinent, the Bay of Bengal, and Southeast Asia. Furthermore, boreal summer also corresponds to the rainy phase of the Ethiopian monsoon (also known as the long rains) with intense convection occurring over the northwest of the country (Funk et al. 2015), and to enhanced rainfall over the eastern Indian Ocean (Yuan and Miller 2002). It is possible that cirrus emanating from the nearby convectively active regions could travel toward the WEIO region, leading to the upper-level cloud in the region. The main candidate is from the Indian/Southeast Asian monsoon, due to the continuous area of elevated convection and decreased OLR extending from this region to the edge of the WEIO box (as can be seen in Fig. 2, there is no such continuous band extending from Africa eastward, ruling out the Ethiopian monsoon as the source of this upper-level cloud).

Transects of the frequency of occurrence anomaly from the annual mean for the four convectively active regimes are shown in Fig. 9 for January and July, with the transect extending northeast from the southwest corner of the WEIO region up through the northeast corner and onward across the southern tip of India. In January, all convective regimes occur more frequently than the annual mean in the southwest part of the WEIO region, and occur less frequently than the annual mean toward the northeast except for the IM regime, which has a slightly elevated occurrence frequency over the entire WEIO section of the transect. This is entirely consistent with an ITCZ positioned along the southern part of the WEIO region. In addition, the elevated IM occurrence over the entire WEIO is part of a noted pattern with an organized convection peak flanked by increased disorganized convection, with enhanced thin cirrus located on the edge of this disorganized convection (the IC anomaly located near 8°N, 74°E). Conversely, in July the largest increase in convective occurrence (as measured by the frequency of the CD regime) along the transect is located over the Bay of Bengal, decreasing toward to the southwest. This is flanked by a transect maximum in elevated CC occurrence along the east coast of India, a transect

FIG. 9. (a) Location of transect (red line) across the WEIO region (black-outlined box) and Southern India, and plots of departures in mean frequency of occurrence of the CC, CD, IM, and IC ISCCP cloud regimes along the transect from the annual mean during (b) January and (c) July. The WEIO region is marked as a purple line along the x axis, and land is marked as a thick black line along the x axis.

maximum in elevated IM regime occurrence over India and its west coast, and a transect maximum in elevated IC regime occurrence over the northeast part of the WEIO region. However, despite the CD and CC regimes having their greatest elevated frequency along the transect over the Bay of Bengal, elevated frequencies along the transect extend into the WEIO region—about two-thirds into the WEIO for the CD regime and about four-fifths into the WEIO transect for the CC regime. The elevated occurrence of CD and CC regimes in the WEIO section of the transect is consistent with the annual cycle in the mean frequency of occurrence of the regimes over the WEIO shown in Fig. 3, and the increased occurrence in these regimes from May to July shown in Fig. 8, but the continuous elevated occurrence of these regimes extending from

the Bay of Bengal into the WEIO region suggests that the increased high level cloud in the WEIO during boreal summer is related to the Indian monsoon.

The transport of high level cloud and moisture from the active areas of the Indian monsoon toward the WEIO region during boreal summer is further apparent in the circulation and location of relative specific humidity anomalies (defined by the anomaly divided by the annual mean) along the transect during July, shown in Fig. 10. The layer between 700 and 400 hPa is anomalously moist or dry over the monsoon-active region or WEIO region of the transect, respectively, with wind along the transect being from the WEIO region to the monsoon region. However, higher in the troposphere (above 350 hPa), wind along the transect is from the monsoon region to the WEIO region, and positive specific humidity anomalies extend deep into the WEIO region, with the furthest extension occurring at around 200 hPa. This pressure range is characteristic of cirrus clouds (Sassen 2002), further suggesting the high-level cloud seen over the WEIO region in boreal summer is transported there from the active monsoon regions located across India and Southeast Asia.

The transport of high-level cirrus clouds from monsoon active areas to the WEIO during boreal summer has previously been shown through forwards Lagrangian tracking through the upper troposphere from areas of deep convection (Luo and Rossow 2004). Figure 5 of Luo and Rossow (2004) shows that over the WEIO cirrus clouds tend to result from formation events farther east in the Indian Ocean, either south of India or in the eastern Bay of Bengal near Thailand and Myanmar, and are transported roughly west to southwesterly over the region, decaying as they approach the African coast. Notably, they find that cirrus clouds from convective events in the eastern Indian Ocean and western Pacific Ocean build to a higher level and persist for a longer period than for other regions. This suggests that while the wind vectors in Fig. 10 show that high-level outflows from the Indian summer monsoon are directed toward the WEIO region of the transect, there is also a strong out-of-transect flow and that outflows from monsoon convection farther east than the Indian subcontinent are likely to be another source of both cirrus and the descending air in the midtroposphere.

As to the circulation which leads to the anomalously dry midtroposphere over the WEIO during boreal summer seen in Fig. 10 and subsequent suppression of deep convection in the region, Fig. 4 suggests that there is horizontal convergence of air over WEIO region at around 400–300 hPa, leading to a net descent of air below this level and through it a drying of the midtroposphere.

FIG. 10. Winds (vectors; m s^{-1}) and relative anomalous specific humidity (color contours, where a value of 1 corresponds to 2 times annual mean specific humidity, 0 corresponds to annual mean specific humidity, and -1 corresponds to 0 specific humidity) along the transect in Fig. 9a during July, with vertical winds scaled by a factor of 1000. The WEIO and land areas are marked as in Fig. 9.

To elucidate the nature of this convergence, the divergence at the 350-hPa level during July is displayed in Fig. 11 and shows that the WEIO region as well as the Mascarene high to the south are local maxima in convergence at 350 hPa. Areas of strong divergence at 350 hPa are located over Ethiopia, over the Pacific near Papua New Guinea and the Philippines, and over Myanmar, with an area of divergence extending from Ethiopia across southern Arabia and the southern foothills of the Himalayas to the Pacific minimum, and extending from the Pacific minimum across Indonesia and along 15°S to about 65°E . The location of these divergence bands suggests that outflow at the 350-hPa level from convection in the active monsoon regions over Ethiopia and south Asia, as well as from the eastern Indian Ocean, leads to upper-level

convergence over the WEIO and thus suppressed deep convection over the region. Note that this is not to imply that the air from the monsoon outflows at this level travel toward the WEIO as the nondivergent winds over the WEIO at the 350-hPa level are generally easterly. Instead, the monsoon outflows generate a strong zonal gradient in the magnitude of the easterly winds over the tropical Indian Ocean, with magnitudes higher in the east and lower in the west.

4. Summary and discussion

This paper explored three main questions: first, what are annual cycles of precipitation, clouds, cloud regimes and vertical motion over the WEIO and what are the

FIG. 11. Mean climatological divergence (shading) and horizontal winds (vectors) over the tropical Indian Ocean at the 350-hPa level during July.

similarities and differences between them; second, how do changes in precipitation throughout the year come from changes in the frequency of more convective states and changes in the precipitation during these states; and third, what seems to drive the changes observed in answer to the second question?

Analysis of the seasonal cycles of rainfall and convective proxies over the WEIO show a single annual peak in precipitation and midtropospheric vertical motion in early austral summer, however outgoing longwave radiation, cloud-top brightness temperatures and occurrence of convectively active cloud regimes show a marked semiannual cycle with convective activity peaking in both austral and boreal summer. While the annual cycle analyses for precipitation and cloud measures suggest different levels of convection in boreal summer, the general trend of higher cloud (as measured by low OLR) correlating with higher precipitation still holds during this period. However, the position of the relationship moves such that a given OLR measurement corresponds to lower rainfall than the rest of the year, and consequently the decrease in mean OLR during boreal summer results in no large mean precipitation changes. Similar behavior is seen with cloud regimes—during boreal summer, the mean intensity of each cloud regime in the WEIO decreases relative to the rest of the year. Decomposition of the changes in mean monthly precipitation to that due to changes in regime frequency and regime intensity shows that the decrease in WEIO precipitation during boreal spring is related to a reduction in highly convective cloud regimes; however, an increase in the occurrence of these regimes during boreal summer is entirely negated by the reduction in their intensity. Furthermore, small decreases in the occurrence of highly convective regimes during austral spring are counteracted by the mean intensity of these regimes returning to that seen through most of the year, resulting in the total mean precipitation increasing.

We suggest that the decrease in mean regime intensities in boreal summer, and likewise in precipitation associated with a given OLR value, is due to two factors: first, increased high-level cloud that is advected into the WEIO region from nonlocal convection and, second, air descending in the midtroposphere making deep convection less favorable. The first factor explains why cloud-based measurements suggest increased convection in the region—because the presence of high cirrus decreases the measured OLR or shifts the distribution of clouds in the cloud-top-pressure/optical-thickness histograms at the heart of cloud-regime classification toward that characteristic of the more convectively active regimes. The second factor explains why the non-cloud measures do not show increased convection—the

circulations associated with monsoons during boreal summer produce a net descent over the area, and as such the likelihood of significant deep convection is reduced. In combination, these factors suggest that, although they are frequent convective proxies, the use of OLR and other cloud-based measurements as proxies assumes that the clouds are representative of processes occurring in the local neighborhood of the clouds. However, when clouds are long-lived enough and there is sizable shear, be it from large scale circulations or otherwise, such assumptions are incorrect and they fail to be good convective proxies.

The annual cycle in WEIO precipitation and outgoing longwave radiation examined in this paper is compared with the annual cycles of 29 members of the CMIP5 project in Fig. 12. As suggested in the previous studies noted in the introduction, over the WEIO region both the annual mean precipitation and the magnitude of precipitation during the boreal winter is above the GPCP magnitude in all 35 models. In terms of the seasonality, all models replicate the reduction in precipitation from January through to March. However, the behavior of the models varies widely over the period from April through October, with some models having a secondary rain maxima of around 6 mm day^{-1} during May–June before a dryer period until September, and other models instead having a dryer period that lasts for June–July before a period of increased precipitation. The seasonality in WEIO top-of-atmosphere OLR across the CMIP5 members shows similar agreement in trend from December–January through to March–April, with OLR over the WEIO increasing for all models. Similar as to precipitation, the models disagree widely from April onward, with some having long periods of elevated OLR and some long periods of decreased OLR, with few to none having the strong semiannual cycle seen in the ISCCP data.

This result suggests that the CMIP5 models are not sufficiently representing some of the real-world features that drive the observed seasonality in both precipitation and OLR. The effect that the boreal summer monsoon circulations appear to have on mean precipitation in the WEIO suggests that improving representations of the monsoon circulation may help models to accurately represent the seasonality of convection over the WEIO during the monsoon period as well the remote effects of this convection. Currently, the CMIP5 ensemble performs poorly at placing summer monsoon convergence and precipitation in the right location, with too little precipitation in the Bay of Bengal and east of the Philippines and too much precipitation off the west coast of India during July and August (Freychet et al. 2015). While the rainfall bias to the west of India may

FIG. 12. Climatological (a) monthly mean precipitation rate (mm day^{-1}) and (b) top-of-atmosphere OLR (W m^{-2}) averaged over the WEIO domain during the historical experiment for 29 members of the “r1i1p1” (where r is realization, i is initialization, and p is physics) CMIP5 ensemble (colored lines as in legend) in comparison with the multimodel mean (dashed black line) and GPCP 1DD [in (a)] and ISCCP climatologies [in (b)] (solid black line).

directly lead to the increased WEIO rainfall seen in the models during boreal summer, the combined effect of all the errors in model monsoon performance may be a factor that leads to the poor representation of seasonality over the WEIO region seen in the CMIP5 ensemble, and suggests a direction for future work in order to better simulate the climate. However, similar or better improvements could be gained from better understanding why climate models strongly overestimate the boreal winter precipitation maxima over the WEIO region, as this is the largest contributor in the CMIP5 models bias over the whole year and is currently under investigated.

Acknowledgments. This work was supported and funded by the Australian Research Council (ARC)

Centre of Excellence for Climate System Science, grant CE110001028. ERA-Interim data used were obtained via the ECMWF MARS web-server. Thanks to Jackson Tan for the cloud-regime dataset used, to Harry Hendon and Abhik Santra for their helpful insight and input, and to the anonymous reviews for their comments and queries that helped in improving the final manuscript.

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