



# Influence of Indian Ocean SST regionality on the East African short rains

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## Abstract

Three sub-regions of the Indian Ocean in which SSTs significantly influence the equatorial East African short rains on interannual timescales are identified, and the physical processes of this influence are studied using regional climate model simulations from the Weather Research and Forecasting model (WRF). Five 20-year ensemble integrations are generated to represent a control climate and to simulate the individual and combined effects of SSTAs in the influential regions. SSTAs in the western Indian Ocean exert a stronger influence on the equatorial East African short rains than central and eastern Indian Ocean SSTAs both in terms of the coverage of significantly-changed precipitation and the magnitude of the precipitation response. Positive western Indian Ocean SSTAs significantly increase the short rains over 95% of the equatorial East Africa domain (30°–40°E, 5°S–5°N), while only 30% of the region responds to central and eastern Indian Ocean SSTAs. Evidence of an influential Indian Ocean dipole mode does not emerge from the analysis. The mechanisms of this influence are diagnosed using atmospheric moisture budget and moist static energy analyses, with reference to Kelvin and Rossby wave generation as in the Gill model, but in the presence of complicated topography and nonzero background flows. Wind convergence anomalies in a moist environment primarily support precipitation anomalies in all cases, while changes in atmospheric instability are largely controlled by low-level moisture. Central and eastern Indian Ocean SSTAs change circulations and precipitation locally, but the remote influence on East Africa is weaker than that of the western Indian Ocean SSTAs.

**Keywords** East african precipitation · Short rains · Kenya precipitation · Indian ocean dipole · Interannual variability · Tropical africa rainfall · SST forcing · Greater horn of africa · WRF

## 1 Introduction

The tropical East African population is dependent on rain-fed agriculture and, therefore, vulnerable to climate variation and change. The rainfall is marginal in most areas across East Africa and exhibits complicated seasonality and prominent regionality. Extreme rainfall events, both flooding and droughts, can significantly impact the agriculture, water and food security, and public health (Anyah et al. 2012; Lyon and DeWitt 2012). There are critical needs to improve our fundamental understanding of the region's precipitation, including its seasonality.

Precipitation over tropical East Africa (EA) exhibits pronounced regional variation and complicated seasonality characterized by a boreal spring “long rains” season and a boreal fall “short rains” season. Past studies (e.g., Behera et al. 2005; Ummenhofer et al. 2009) link interannual variability of the EA short rains with SST variability in the Indian Ocean. However, there are unsolved questions about the relationship between Indian Ocean SSTs and EA short rains. For example, are some regions of the Indian Ocean more influential, or do the basin-wide Indian Ocean SSTs impact EA short rains equally?

The purpose of this study is to investigate the regionality of the Indian Ocean SST forcing and the mechanisms by which this forcing influences the equatorial EA short rains in the current climatology (1998–2017). We identify hot spots of influential SSTAs, and study the hydrodynamics of the resulting perturbations in isolation and in combination using regional model (RCM) simulations. Background on the EA short rains is provided in Sect. 2. The RCM simulations,

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reanalyses, observational datasets, and methodologies used are described in Sect. 3, with a model evaluation presented in Sect. 4. The main results are presented in Sect. 5, and conclusions are summarized in Sect. 6.

## 2 Background

Rainfall in equatorial EA is often characterized by two rainy seasons. One rainy season occurs during the boreal spring, referred to as “long rains”, and a second rainy season in the boreal fall is referred to as “short rains”. Although the long rains provide a larger amount of rainfall to EA than the short rains (Hastenrath et al. 1993), the latter exhibit more interannual variability (Black et al. 2003; Hastenrath et al. 2011; Lyon 2014). This study, therefore, focuses on the EA short rains and associated mechanisms.

EA precipitation exhibits complex seasonality and pronounced regionality (Nicholson 1996, 1998, 2000; Herrmann and Mohr 2011; Liebmann et al. 2012; Lyon 2014). For example, Herrmann and Mohr (2011) classify the seasonality of precipitation over Africa based on multiple precipitation datasets and find equatorial Africa often exhibits two wet seasons. They suggest that the complex topography over EA and location near the equator result in the continent’s richest mix of different seasonality classes. In Lyon (2014), the annual cycle of EA precipitation shows great spatial heterogeneity owing in large part to its complex terrain.

Previous studies (Hastenrath et al. 1993; Black et al. 2003) relate the interannual variability of EA short rains to Pacific SSTs. However, a basinwide large-scale coupled mode, referred to as the Indian Ocean Dipole (IOD) or Indian Ocean Zonal Mode (IOZM), is found to be more influential for EA short rains than the El Niño–Southern Oscillation (ENSO) (Saji et al. 1999; Webster et al. 1999; Behera et al. 2005). For example, Bahaga et al. (2015) investigated the roles of the IOD and ENSO in the variability of EA short rains using observations and GCMs, and found that the main driver of EA short rains is the IOD and ENSO provides only a minor contribution.

EA analysis domains vary in the studies relating EA short rains to the IOD. For example, Black et al. (2003) analyze composites of extreme years using station data for 1900–97 over (37.5°–41.25°E, 2.5°S–2.5°N) and (37.5°–41.25°E, 10°–12.5°S), and find extreme EA short rains are associated with large-scale SSTA patterns in the Indian Ocean during IOZM events. They indicate that only IOZM events that reverse the zonal SST gradient for several months trigger high rainfall. Behera et al. (2005) find that extreme years (1961, 1994, and 1997) of EA short rains over (5°S–5°N, 35°–46°E) can be explained by intrinsic zonal variability related to the

IOD. Wenhaji Ndomeni et al. (2018) find a strong positive correlation between October–December precipitation over (5°S–20°N, 28–52°E) and the IOD index in observational datasets, suggesting the positive-phase IOD plays a dominant role in driving EA short rains.

The correlations between the Indian Ocean SSTs and EA short rains shows interdecadal variability. Clark et al. (2003) study the rainfall during October–December along the coast in Kenya and Tanzania, and find correlations between Indian Ocean SSTs and EA precipitation in 1950–1982 and 1983–1993 nearly reverse in sign.

Several modeling studies (Ummenhofer et al. 2009; Bahaga et al. 2015) show the western Indian Ocean plays a dominant role in impacting EA short rains. Ummenhofer et al. (2009) study the relationship of October–November rainfall over (10°N–1°S, 31°–45°E) and IOD using ensemble simulations with an atmospheric general circulation model (GCM). They assess the contributions of individual (and combined) poles of the IOD to above-average precipitation over EA. They show that increased EA short rains during positive IOD are driven mainly by warming over the western Indian Ocean (38°–70°E, 12°S–12°N), leading to a reduction in sea level pressure over the western half of the Indian Ocean. Converging wind anomalies over EA lead to moisture convergence and increased convective activity.

Past regional modeling studies have shown that RCMs are able to reproduce the EA observed rainfall seasonality and regional circulation patterns (e.g., Sun et al. 1999; Segele et al. 2009; Cook and Vizy 2012; Cook and Vizy 2013; Endris et al. 2013; Ogwang et al. 2016; Han et al. 2019). The coarse resolution of current GCMs limits their capability in capturing the important regional forcing features, such as complex topography and coastlines (Cook and Vizy 2013; Endris et al. 2013; Ogwang et al. 2016). Shongwe et al. (2011) approximate the observed sensitivity of EA precipitation during October–December to Indian Ocean SSTAs, but the GCMs do not generally produce an accurate representation of the EA climate. A regional simulation with higher resolution simulations is needed because of the important role of topography in determining EA rainfall distributions (Hession and Moore 2011; Lyon 2014) and the observed complex regionality of the rainy seasons (Herrmann and Mohr 2011).

Regardless of the numerous studies discussed above, there are unsolved questions about the relationship between Indian Ocean SSTs and EA short rains. Are some regions of the Indian Ocean more influential, or do the basin-wide Indian Ocean SSTs impact EA short rains equally? In any case, what are the mechanisms of the Indian Ocean SSTs impacting EA short rains? In this study, we will address these questions to improve our understanding of the influence of Indian Ocean SSTs on EA short rains variability.

### 3 Methodology

#### 3.1 Regional model simulations

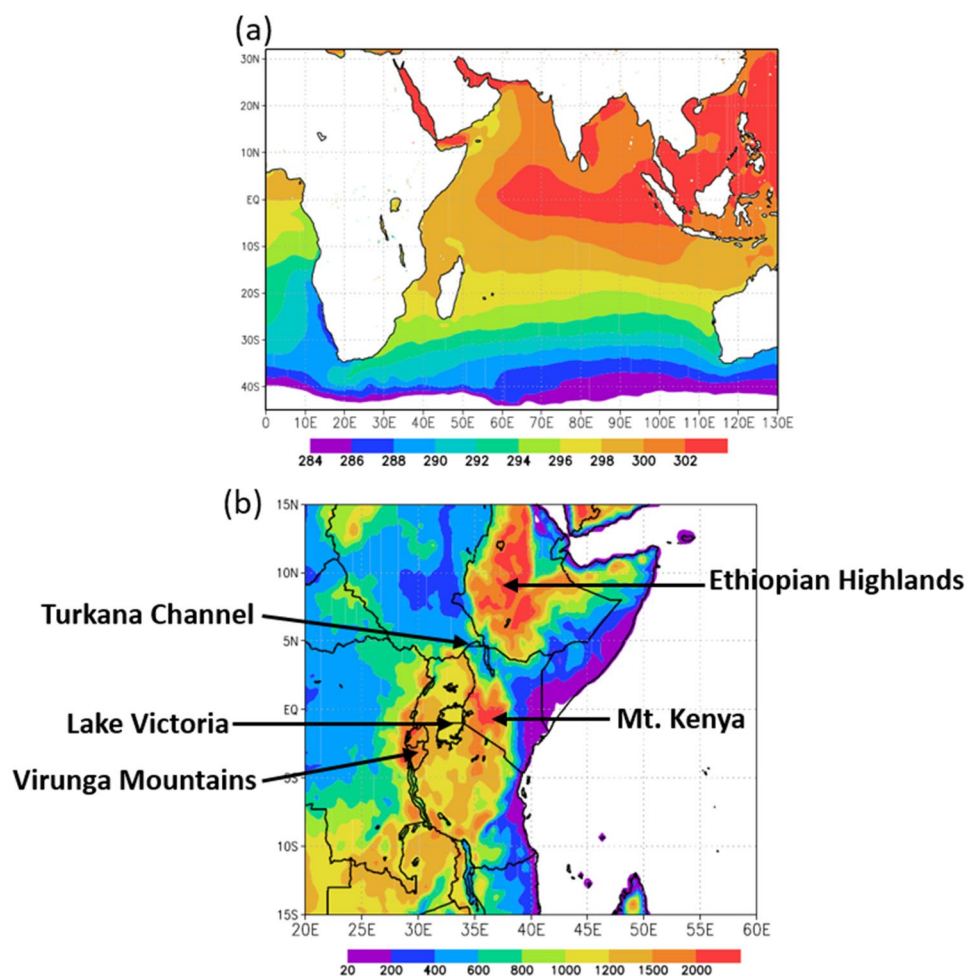
The National Center for Atmospheric Research—National Oceanic and Atmospheric Administration Weather Research and Forecasting (WRF; Skamarock et al. 2008), version 3.8.1, is used to conduct RCM simulations over a large domain ( $0^{\circ}$ – $130^{\circ}$ E,  $44^{\circ}$ S– $32^{\circ}$ N; Fig. 1a) covering most of the Africa continent, and tropical and sub-tropical Indian Ocean. The simulation domain is chosen to minimize the effects of lateral boundary constraints over EA and Indian Ocean, and to allow the development of sub-tropical anticyclone over the Indian Ocean. Other synoptic and large-scale circulations (e.g., the low-level Turkana jet and the overturning zonal circulation over the equatorial Indian Ocean) are also included in the simulation domain.

The model is run with 30-km horizontal resolution and 40 vertical levels, with the top of the atmosphere is set at 10 hPa. The time step is 60 s and model output is saved every 3 h. A combination of parameterizations is used as

it has been shown to reproduce the African climate and the seasonality of EA precipitation realistically (Vizy and Cook 2009; Cook and Vizy 2012; Vizy et al. 2013, 2015; Cr  tat et al. 2014; Han et al. 2019). Physical parameterizations selected include the Lin et al. microphysics scheme (Lin et al. 1983; Rutledge and Hobbs 1984; Chen and Sun 2002), the rapid radiative transfer model longwave radiation scheme (Mlawer et al. 1997), the Dudhia shortwave radiation scheme (Dudhia 1989), the Monin–Obukhov Similarity surface layer scheme (Skamarock et al. 2008), the Yonsei University boundary layer scheme (Hong et al. 2006), the Noah land surface model (Chen and Dudhia 2001), and Kain–Fritsch (new Eta) cumulus scheme (Kain and Fritsch 1993).

Figure 1b shows prominent topographical features over tropical East Africa as resolved in the model, which is the topography from the US Geological Survey digital elevation model interpolated to 30-km resolution. The Ethiopian highlands and the East African highlands are separated by the Turkana channel. Model elevation maxima are Mt. Kenya ( $37^{\circ}$ E,  $0^{\circ}$ N) at 2783 m ( $\sim$  730 hPa), and the Virunga

**Fig. 1** **a** Simulation domain and SSTs (K) in the control simulations. **b** Elevation (m) as resolved in the 30-km simulation. Thin black lines indicate country outlines



Mountains in the Rwanda-Burundi region at 1878 m (~815 hPa). Lake Victoria is located between these two peaks.

Five sets of simulations, or ensembles, are run with the only difference between the ensembles being in the prescribed SSTs. One is a control climate, and four ensembles have idealized Indian Ocean SSTAs added to the control climate SSTs. The SSTAs are Gaussian-shaped and derived from observed correlations between EA precipitation and Indian Ocean SSTs (see Sect. 5.1). Each ensemble consists of 20 simulations that are initialized on August 1 of a different year (1998–2017), allowing the model to spin up, and run for 153 days to December 31. The simulation with 20 ensemble members is designed to suppress variability on synoptic to interannual timescales, to study the EA short rains from a climatological perspective. The model output from each 20-year run is averaged to form a climatology. Here we define a climatology as an average over the 20-year period (1998–2017) that is the focus of this study. The year 1998 is chosen as the start year to coordinate with NASA TRMM Multi-satellite Precipitation analysis (TRMM; Huffman et al. 2007) availability. Initial, lateral, and surface boundary conditions are specified at 6-hourly intervals from the European Centre for Medium-Range Weather Forecasts Interim reanalysis (ERA-Interim; Dee et al. 2011; 1.5° latitude × 1.5° longitude) for each year (1998–2017).

In the control ensemble, climatological SSTs from ERA-Interim averaged from 1998 to 2017 (Fig. 1a) are used except for Lake Victoria (centered on 33°E, 2°S), where the daily Operational Sea Surface Temperature and Sea Ice Analysis (OSTIA; Donlon et al. 2012) values averaged from 2012 to 2017 are used. OSTIA values are available on a global 1/20° (~6 km) grid from 2007 for daily SSTs and lake surface temperatures from November 2011. OSTIA is used over Lake Victoria because it captures the observed SST gradient across Lake Victoria, while the coarser resolution (1.5°) ERA-Interim has only one grid point over Lake Victoria. Past studies (Thiery et al. 2015; Argent et al. 2015; Woodhams et al. 2018) indicate that Lake Victoria SST gradients impact regional precipitation. In addition, ERA-Interim SSTs are lower than OSTIA SSTs averaged over Lake Victoria by ~2 K, and they change only at monthly intervals. For these reasons, OSTIA is commonly used for regional studies of Lake Victoria and adjacent East Africa (Thiery et al. 2015; Argent et al. 2015; Woodhams et al. 2018). One disadvantage of OSTIA for our study is that the number of years of OSTIA SSTs (6 years from 2012–2017) is less than for ERA-Interim (20 years from 1998–2017).

### 3.2 Observational/reanalysis datasets

To evaluate the accuracy of the simulated hydrodynamics, we compare with the 6-hourly ERA-Interim reanalysis and

the 6-hourly Japan Meteorological Agency's Japanese 55-Year Reanalysis (JRA-55; Kobayashi et al. 2015; 1.25° latitude × 1.25° longitude). These two reanalyses realistically capture the circulation over Africa and Indian Ocean, and are commonly used in regional studies of EA (Vizy and Cook 2012, 2019; Cook and Vizy 2013).

To evaluate precipitation in the simulations, we use TRMM at 0.25° resolution, the NOAA Precipitation Estimation from Remotely Sensed Information using an Artificial Neural Network Climate Data Record (PERSIANN; Ashouri et al. 2015; 0.25° resolution), The Climate Hazards Group Infrared Precipitation with Satellite Data (CHIRPS; Funk et al. 2015; 0.05° resolution), and The NOAA Climate Prediction Center morphing technique precipitation V2 bias-corrected dataset (CMORPH; Joyce et al. 2004; 8-km resolution).

To examine the observed correlations of Indian Ocean SSTs and EA short rains, the daily National Oceanic and Atmospheric Administration Optimum Interpolation observed SSTs (NOAA OI SST; Reynolds et al. 2007) version 2.0 at 0.25°-resolution and the monthly Hadley Centre Sea Ice and SST dataset (HadISST; Rayner et al. 2003) at 1°-resolution are used.

### 3.3 Analysis methods

The atmospheric column moisture budget is used to connect precipitation with the large-scale circulation and moisture fields. The climatological precipitation rate can be decomposed according to

$$P = E + C + A + R, \quad (1)$$

where  $P$  is precipitation,  $E$  is evapotranspiration,  $C$  represents contributions from horizontal wind convergence in a moist environment,  $A$  is the vertically integrated horizontal advection of moisture, and  $R$  is a residual term that includes orographic precipitation, the effects of transient eddies, and numerical error (Lenters and Cook 1995; Vizy and Cook 2001; Cook and Vizy 2013).

The convergence term,  $C$ , which is shown to provide the strongest contribution to the total precipitation,  $P$ , in Sect. 5.3, is calculated as

$$C = -\frac{1}{g\rho_w} \int_{p_s}^{p_{top}} [q\nabla_h \cdot \vec{V}] dp, \quad (2)$$

where  $g$  is gravitational acceleration,  $\rho_w$  is the density of water,  $p_s$  and  $p_{top}$  are the pressure at the surface and the top of the atmosphere, respectively,  $q$  is specific humidity, and  $\vec{V}$  is the horizontal wind.  $C$  is decomposed into zonal,  $C_z$ , and meridional components,  $C_m$ , as follows:



$$C_z = -\frac{1}{g\rho_\omega} \int_{p_s}^{p_{top}} \left( q \frac{\partial u}{\partial x} \right) dp \quad (3)$$

and

$$C_m = -\frac{1}{g\rho_\omega} \int_{p_s}^{p_{top}} \left( q \frac{\partial v}{\partial y} \right) dp. \quad (4)$$

Moist static energy (MSE) is used to measure the local instability of the atmosphere. MSE is defined as the sum of the sensible heat, latent heat, and geopotential energy contents of a parcel according to

$$MSE = c_p T + Lq + gz. \quad (5)$$

In Eq. (5),  $c_p$  is the specific heat of air at constant pressure,  $T$  is the air temperature,  $L$  represents the latent heat of water vaporization, and  $z$  is the geopotential height. MSE profiles measure atmospheric stability and discriminate the individual roles of temperature and moisture in generating instability. MSE increases with altitude indicate a stable atmosphere, and a neutral profile generally reveals the presence of convection.

## 4 Model evaluation

Figure 2a–c show climatological (1998–2017) precipitation averaged over October 1–December 15 in PERSIANN, TRMM, and the control ensemble, respectively. The primary analysis region ( $30^\circ$ – $40^\circ$ E,  $5^\circ$ S– $5^\circ$ N) is shown by the black box, located over equatorial East Africa to the east of the Congo basin. As discussed below, sensitivity to domain boundaries is low. The averaging period for the short rains in the EA analysis domain is chosen as October 1–December 15, when precipitation rates in both TRMM and PERSIANN exceed 2 mm/day (Fig. 2d). The definition of the short-rains season in EA varies in the literature (Black et al. 2003; Clark et al. 2003; Yang et al. 2015; Hirons and Turner 2018). Other periods (October–November and October–December) are tested, and we find that the results are not sensitive to the choice of the short-rains averaging period in our study.

In PERSIANN, TRMM, CMORPH, and CHIRPS (Fig. 2a–d), rainfall maxima are located over Equatorial Guinea ( $15^\circ$ E,  $0^\circ$ ), Rwanda ( $28^\circ$ E,  $2^\circ$ S). The coastal regions over East Africa (i.e., eastern Kenya and southern Somalia) and southwestern Ethiopia have higher precipitation rates than northern Kenya ( $35^\circ$ E,  $2^\circ$ N) by 1–2 mm/day. The control ensemble (Fig. 2c) captures the precipitation pattern in observations, but has a wet bias over some regions, especially over the Congo basin, consistent with past RCM studies (Vizy and Cook, 2012; Cr  tat et al. 2014). Precipitation in the control climatology exceeds the observed rainfall by 2–5 mm/day over the equatorial central Indian Ocean and

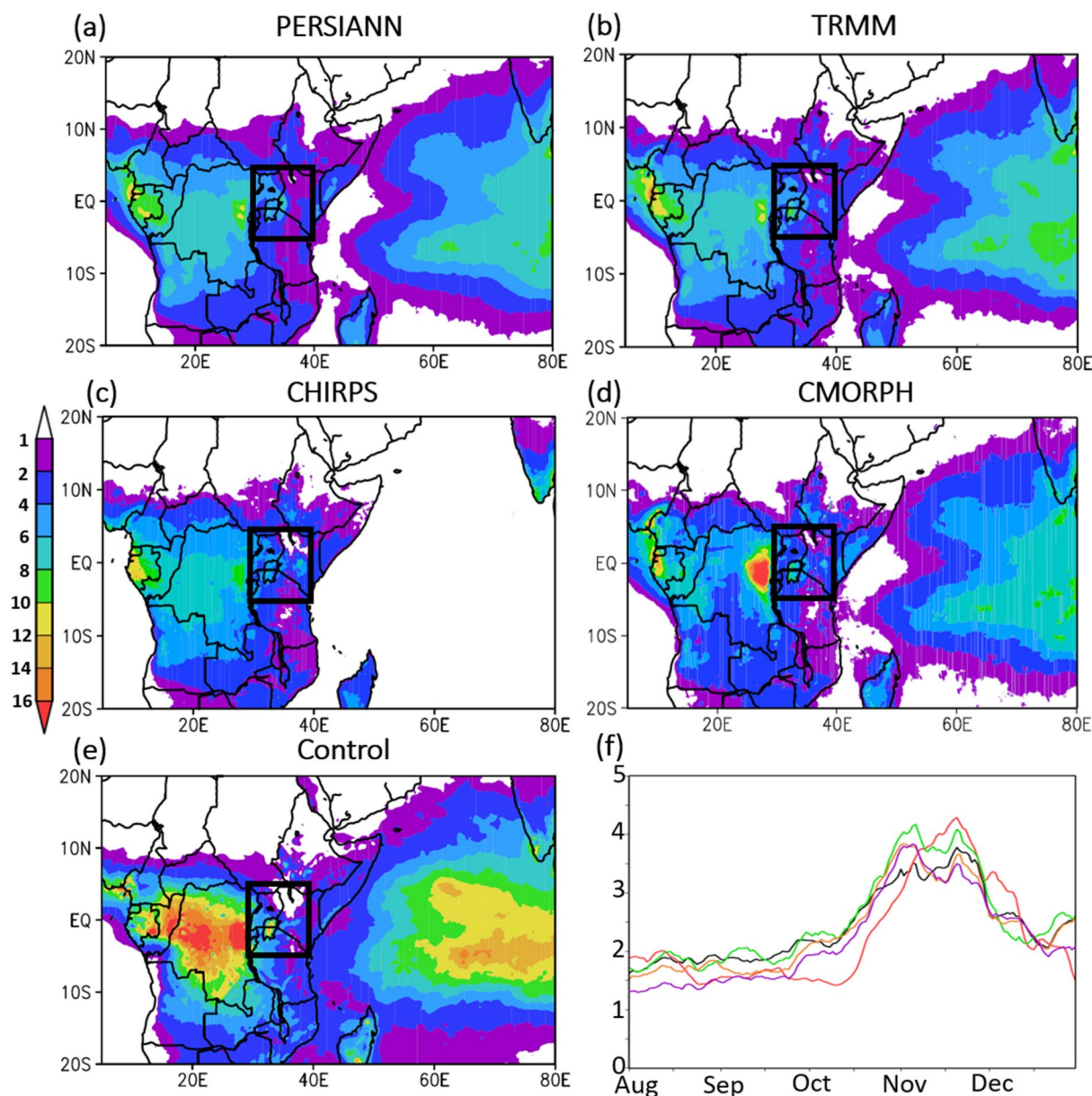
by 5–9 mm/day over the Congo Basin. Over the EA analysis domain (black box), the control ensemble simulates precipitation patterns and magnitudes reasonably showing high rainfall rate (2–6 mm/day) over the adjacent region of Lake Victoria and low rainfall rate (< 2 mm/day) over ( $35^\circ$ – $40^\circ$ E,  $1^\circ$ – $5^\circ$ N), except for a wet bias over central Lake Victoria.

Figure 2d shows 11-day running means of precipitation averaged over the EA domain in PERSIANN (black), TRMM (green), CMORPH (purple), CHIRPS (orange), and the control ensemble (red). Applying the 11-day running mean effectively helps to filter out the synoptic variations. Differences between precipitation from the 20-year control simulation and the observations are insignificant at the 90% confidence level using the two-tailed Student's-t test, and the simulated seasonality is realistic. Using a threshold of 2 mm/day, the short-rains season in the control ensemble starts on October 20 and persists to December.

Figure 3a–c show climatological 850-hPa wind and specific humidity from ERAI, JRA-55, and the control ensemble, respectively. The control ensemble captures the winds in the two reanalyses, including westerlies in the equatorial Indian Ocean, northeasterlies from the Arabian Sea to Somalia, and easterlies/southeasterlies over the southern Indian Ocean between  $10^\circ$ S and  $25^\circ$ S. The model-simulated specific humidity also agrees well with ERAI and JRA-55 in terms of patterns and magnitudes. Specific humidity maxima ( $> 12 \text{ g kg}^{-1}$ ) are located over the Congo basin and the Maritime Continent. Other vertical levels are also examined (not shown) to conclude that the 30-km RCM provides a reasonable simulation of the 1998–2017 climatology.

Figure 3d–f show cross sections of the climatological zonal wind along the equator from ERAI, JRA-55, and the control ensemble, respectively, from 1000 to 500 hPa. Westerlies over the equatorial Indian Ocean and easterlies over the EA domain are divergent over coastal EA (eastern Kenya) in both the control ensemble and reanalyses. Low-level westerlies over the Congo basin are convergent with easterlies from the EA coast ( $\sim 42^\circ$ E). The westerlies over the Congo Basin in the two reanalyses (Fig. 3d, e) are located between the surface and 900 hPa, with magnitudes of 0–2 m/s. The model-simulated westerlies are stronger and deeper than in the reanalyses, extending from the surface to 750 hPa with a maxima of 5 m/s. Thus, moisture convergence (not shown) associated with the westerly flow over the Congo Basin and the easterly flow over the EA domain in the control climatology is stronger than in the reanalyses, which is consistent with the wet bias over the Congo Basin in the RCM (Fig. 2c).

Moisture budget components in the control ensemble and ERAI are compared to evaluate how well the RCM captures the mechanisms of precipitation. Figure 4a–f show the P-E, A, R, C,  $C_z$ , and  $C_m$  terms in the moisture budget (Eqs. 1–4), respectively, in the control ensemble. P-E is positive over the



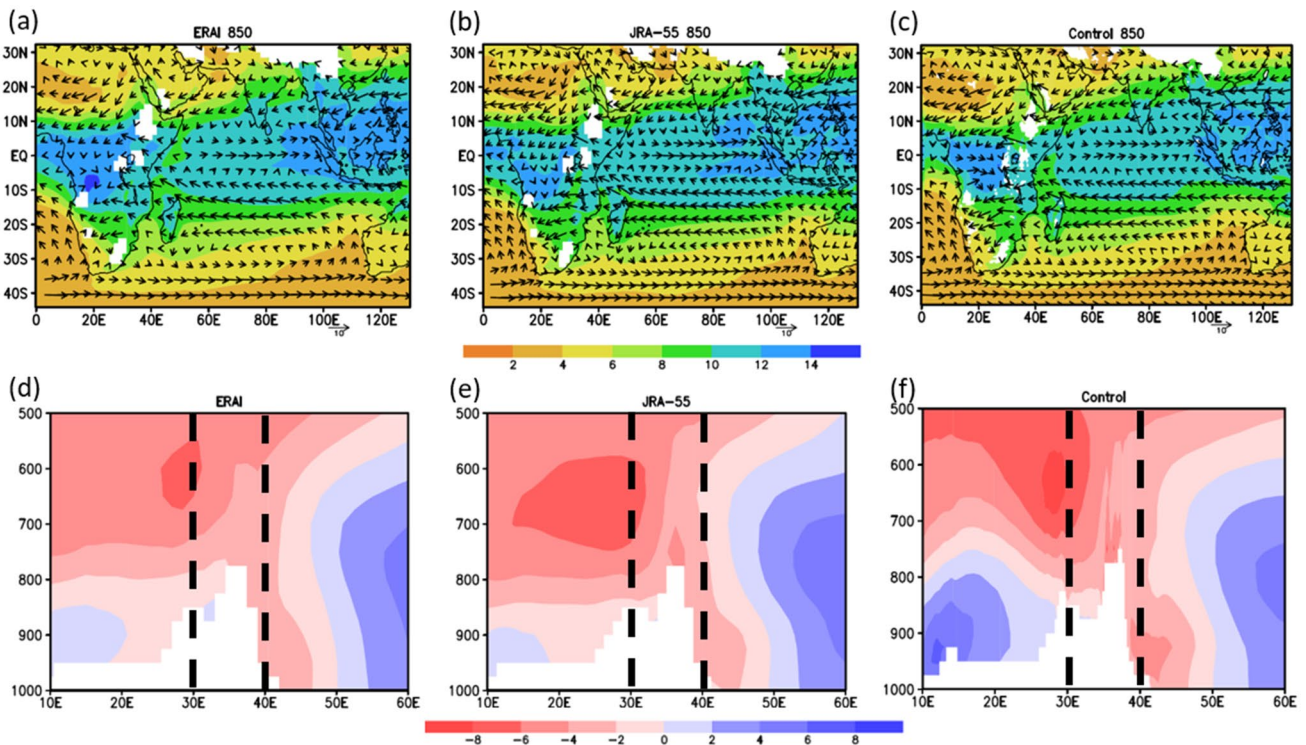
**Fig. 2** Climatological (1998–2017) precipitation averaged over October 1–December 15 from **a** PERSIANN, **b** TRMM, **c** CHIRPS, **d** CMORPH, and **e** the control ensemble. Black box denotes the EA analysis domain. Thin black lines indicate country outlines. **f** Daily

precipitation smoothed using an 11-day running mean averaged over the EA domain in PERSIANN (black), TRMM (green), CMORPH (purple), CHIRPS (orange), and the control ensemble (red). Units: mm/day

Congo Basin and the Indian Ocean east of 50°E, with a maximum of 10 mm/day (Fig. 4a). Moisture advection (Fig. 4b) is negative with a magnitude below 2 mm/day, except for a region of positive values centered on (35°E, 4°N) in the Turkana Channel where southeasterly flow transports relatively wet air from the Indian Ocean (Nicholson 2016; Vizy and Cook 2019). The residual term (Fig. 4c) largely reflects topographic patterns, with strong positive values (> 8 mm/day) over the EA domain east of 35°E. Moisture convergence (Fig. 4d) with magnitudes of 2–10 mm/day is located

over the Congo Basin and the EA domain to the west of Mt. Kenya (35°E), and it supports positive P-E in the same regions. Moisture divergence (Fig. 4d) is present over the EA domain to east of 35°E. A clear east–west contrast of moisture convergence term exists over the EA domain (Fig. 4d).

Figure 4e and f show the zonal and meridional components of the moisture convergence term (Eqs. 3 and 4). Moisture convergence over the Congo Basin (Fig. 4d) is mainly supported by the zonal component (Fig. 4e) in association with westerlies over the Congo basin and easterlies from the



**Fig. 3.** 850-hPa wind (vectors;  $\text{m s}^{-1}$ ) and specific humidity (shaded;  $\text{g kg}^{-1}$ ) averaged over October 1–December 15 in **a** ERAI, **b** JRA-55, and **c** the control ensemble. Cross sections of the zonal wind ( $\text{m s}^{-1}$ )

along the equator averaged over October 1–December 15 in **d** ERAI, **e** JRA-55, and **f** the control ensemble. Black dashed lines locate the EA analysis domain. Topography is masked out

EA coast (Fig. 3f). Moisture divergence over the EA domain east of  $35^\circ\text{E}$  (Fig. 4d) is primarily supported by the zonal component (Fig. 4e) and it is associated with easterlies over the EA domain and westerlies over the ocean (Fig. 3f).

It is clear from the moisture budget analysis that precipitation in the western and eastern portions of the EA domain is supported by different mechanisms, especially where the residual (Fig. 4c) and the moisture convergence terms (Fig. 4d) are concerned. The short rains over the western EA domain are mainly supported by moisture convergence while they are mainly maintained by the residual term over the eastern EA. The contribution of residual term to the precipitation over eastern EA domain indicates the possible importance of topographic effects of mountains (such as Mt. Kenya) in determining precipitation. Since it is important to average over homogenous regions, we divide the EA analysis domain into western and eastern parts separated at  $35^\circ\text{E}$  in the following analysis.

Figure 4g–m are the same as Fig. 4a–f but for the ERAI climatology. Similar to the control ensemble (Fig. 4a–f), ERAI indicates that the moisture convergence and residual terms have opposite signs over the western and eastern EA domains. A comparison with the simulated moisture budget (Fig. 4a–f) indicates that the control ensemble produces an accurate decomposition of the processes that maintain

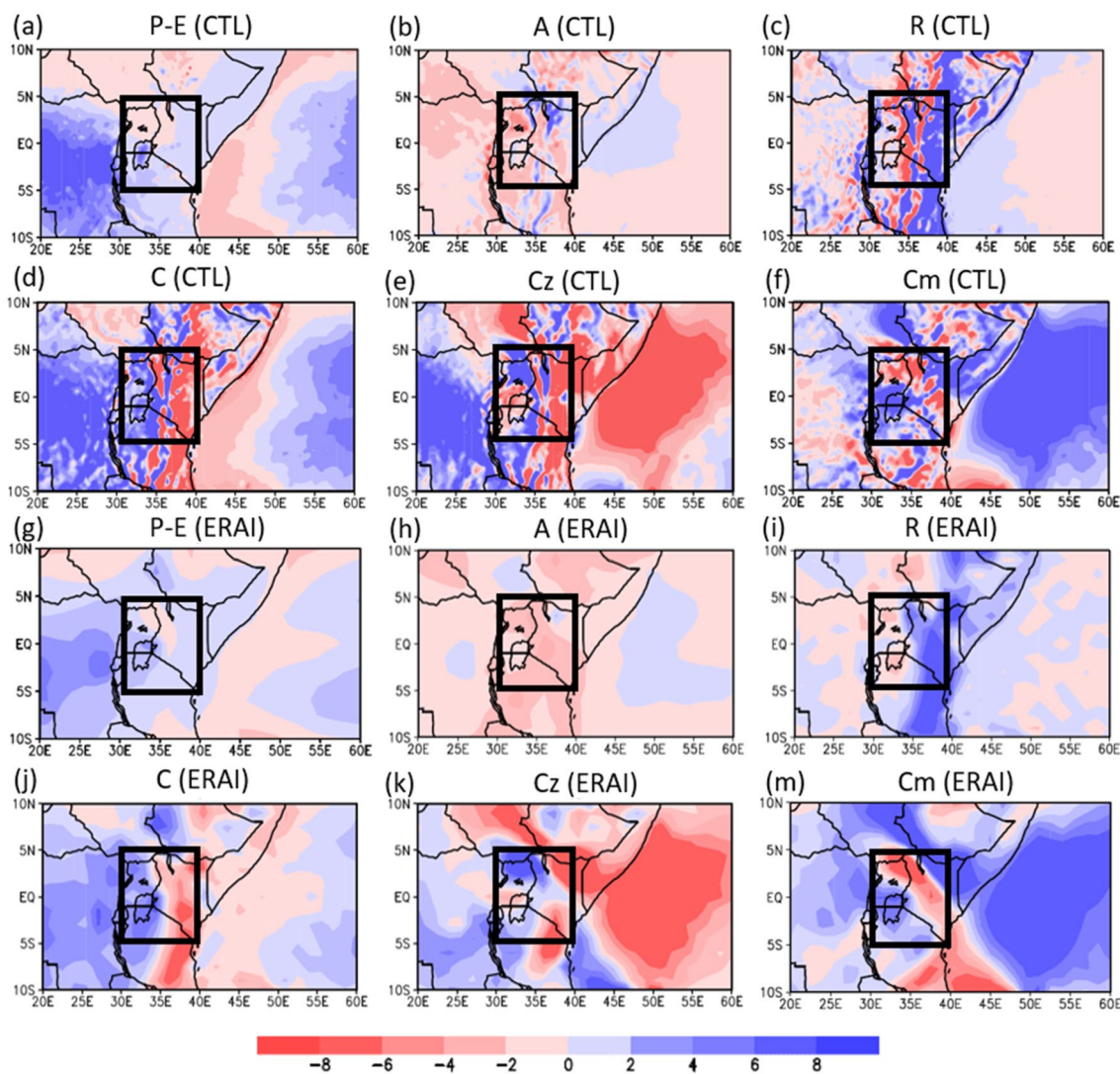
rainfall in the analysis domain, adding to our confidence in the regional model simulations. The simulations (Fig. 4b–e) show more structure in the moisture budget terms along the topography ( $35^\circ\text{E}$ ) than ERAI (Fig. 4h–k) as the topography is resolved at a higher resolution in the simulations.

## 5 Results

### 5.1 Observed correlations of EA short rains and Indian Ocean SSTs

Figure 5a and b display correlations of TRMM rainfall averaged over the full EA domain ( $30^\circ\text{--}40^\circ\text{E}$ ,  $5^\circ\text{S--}5^\circ\text{N}$ ) and Indian Ocean SSTs averaged over October 1 to December 15 in the 1998–2017 climatology using ERAI and NOAA OI SSTs, respectively. Correlations using PERSIANN (not shown) produce similar patterns. The EA short rains are positively correlated with western and central Indian Ocean SSTs along the equator, and negatively correlated with eastern Indian Ocean SSTs at approximately  $5^\circ\text{S--}10^\circ\text{S}$ . The correlations are as high as 0.8 in the western and central Indian Ocean, and  $-0.7$  in the eastern Indian Ocean. Correlations using EA short rains averaged





**Fig. 4** **a** Precipitation minus evapotranspiration, **b** moisture advection term, **c** residual term, **d** moisture convergence term, **e** zonal moisture convergences term, and **f** meridional moisture convergence term in the moisture budget [Eq. (1)] averaged over October 1 to December

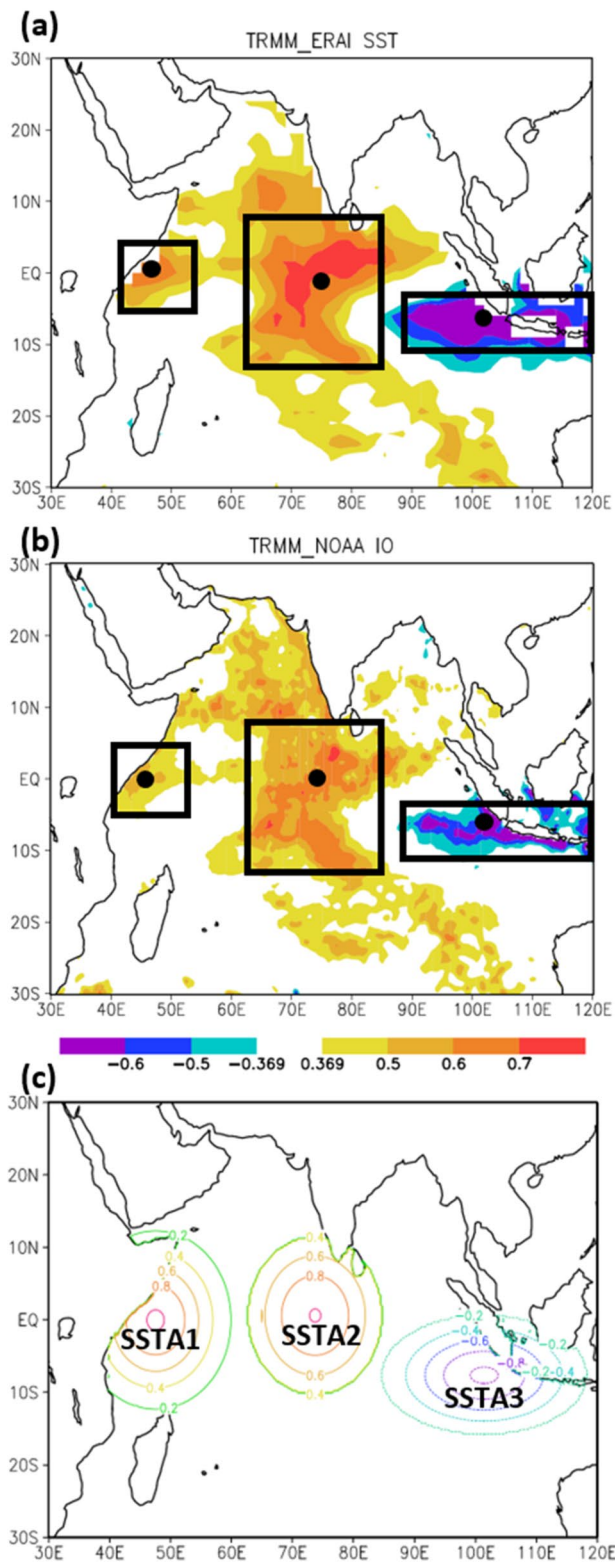
15 in the control climatology. Unit: mm/day. The black box denotes the EA analysis domain. The black contours show country outlines. **(g)–(m)** are the same as **(a)–(f)**, respectively, but using ERAI

over the western and eastern EA domains separately (not shown) produce similar patterns over the western, central, and eastern Indian Ocean.

To understand the extent to which the three SSTAs occur independently, correlations between SSTs averaged over the black boxes in Fig. 5 are calculated. Results using NOAA OI SST, ERAI SST, and HadISST are similar. The western and eastern Indian Ocean SSTAs are not significantly correlated at the 90% confidence level, with a correlation coefficient of  $-0.24$ . SSTAs in the western and eastern Indian Ocean have opposite signs in 11 (55%) out of 20 years from 1998 to 2017. In contrast, SSTAs in the western and central Indian Ocean are highly correlated (>99% confidence level) with a correlation of 0.92.

Past studies (e.g., Saji et al. 1999; Saji and Yamagata 2003; Li et al. 2003; Luo et al. 2010) report a dipole mode of SST variability in the tropical Indian Ocean (IOD), quantified by the SST difference between the tropical western Indian Ocean ( $50^{\circ}\text{E}$ – $70^{\circ}\text{E}$ ,  $10^{\circ}\text{S}$ – $10^{\circ}\text{N}$ ) and the tropical south-eastern Indian Ocean ( $90^{\circ}$ – $110^{\circ}\text{E}$ ,  $10^{\circ}\text{S}$ – $0^{\circ}$ ), referred to as the IOD index. The averaging regions used in calculating the IOD index are different from the regions identified here as being especially influential for the EA short rains in Fig. 5. Further, our results suggest that the western and eastern Indian Ocean SSTAs are not significantly anticorrelated using the averaging boxes in Fig. 5 for 1998–2017. Other studies (Nicholls and Drosowsky 2001; Dommenget and Latif 2002; Zhao and Nigam 2015; Wang et al. 2019) find a





**Fig. 5** Correlations of TRMM rainfall averaged over the EA domain with **a** ERAI SSTs and **b** NOAA IO SSTs for the October 1 to December 15 climatology. Only correlations with confidence levels higher than 90% in a two-tailed Student’s t-test are shaded. Black dots are the locations of maxima of the SSTAs in **(c)**. Black boxes are averaging regions for western, central, and eastern Indian Ocean SSTAs. **c** SSTAs (K) in the perturbation simulations

**Table 1** Ensemble descriptions

Ensemble name	SSTs used in ensembles
Control (CTL)	Climatological SSTs
Western Indian Ocean (WIO)	Climatological SSTs + SSTA1
Central Indian Ocean (CIO)	Climatological SSTs + SSTA2
Eastern Indian Ocean (EIO)	Climatological SSTs + SSTA3
Warm-cold Indian Ocean (ALL)	Climatological SSTs + SSTA1 + SSTA2 + SSTA3

similar lack of anticorrelation using the averaging regions defined for the IOD index. The fundamental mechanisms of the IOD are beyond the purpose of this study and not discussed further.

Figure 5c shows the locations and magnitudes of the idealized SSTAs used in the perturbation ensemble simulations. The choice of these SSTAs is guided by the correlations shown in Fig. 5a, b. In addition to the western, central, and eastern Indian Ocean, significant correlations also exist over (65°–70°E, 8°–12°N) in the Arabian Sea in Fig. 5a, b. However, we do not apply SSTAs over the Arabian Sea in the perturbation simulations because the correlations are less robust in the sense that they are sensitive to the exact choices of EA averaging region.

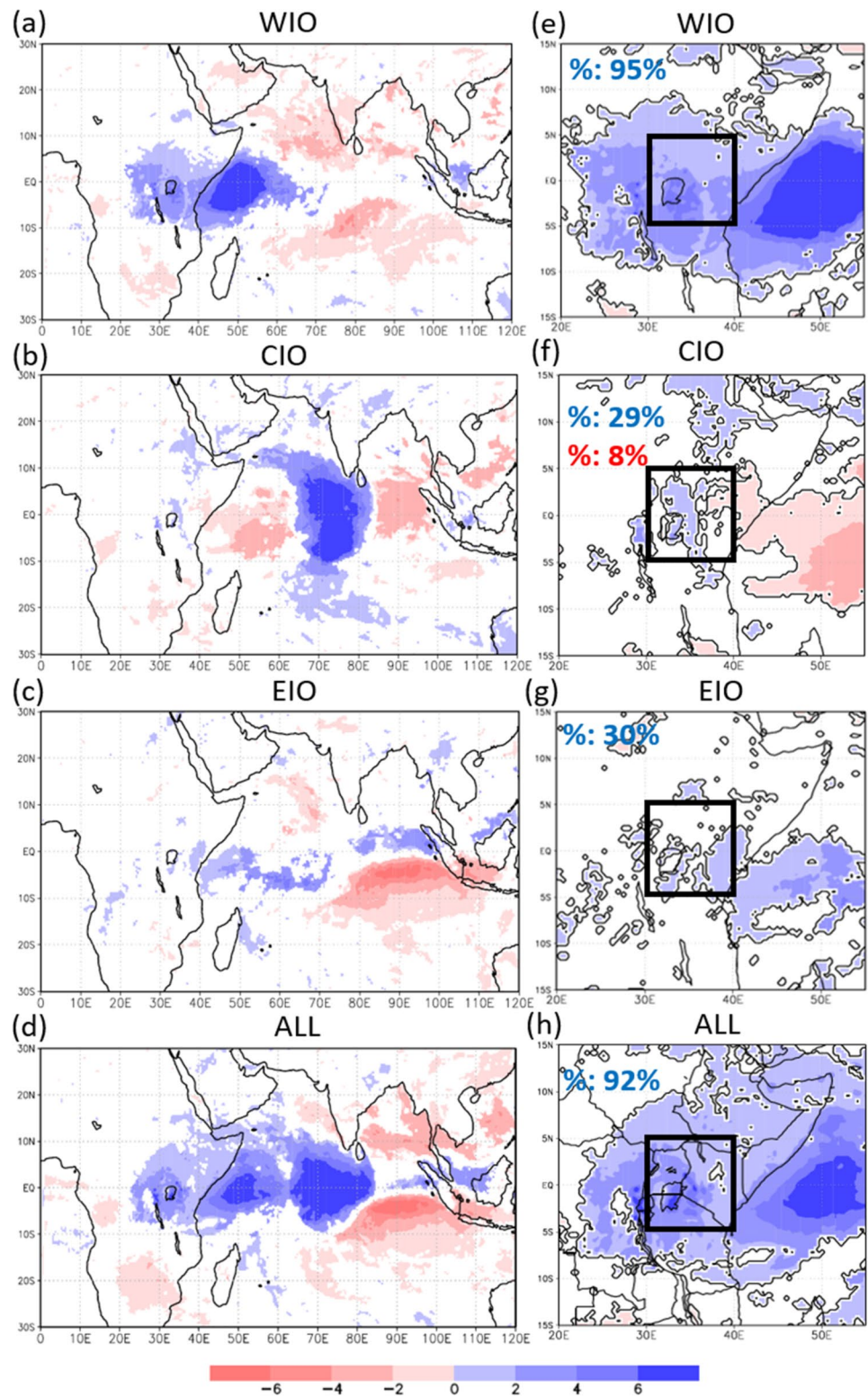
The three SSTAs shown in Fig. 5c are Gaussian in shape and have maximum amplitude of +1 K in the western and central Indian Ocean, and –1 K in the eastern Indian Ocean, covering the regions with significant correlations shown in Fig. 5a, b. The SSTA maximum amplitudes are larger than the standard deviations of the SSTAs in ERAI (0.3–0.4 K) to produce diagnosable responses, but similar in magnitude to those of the most extreme years.

Table 1 provides a reference for naming the five ensemble simulations. The three SSTAs in Fig. 5c are imposed individually in the WIO, CIO, and EIO ensembles. The ALL ensemble includes all three SSTAs shown in Fig. 5c, comparing with the linear superposition of the WIO, CIO, and EIO ensembles to evaluate the potential for interference among the individually-forced responses.

### 5.2 Precipitation anomaly in perturbation ensembles

Figure 6a–d show anomalous precipitation, defined as differences from the CTL climatology, for the WIO, CIO, EIO, and ALL climatologies (see Table 1). Only areas with confidence levels higher than 90% are shaded. The same precipitation anomalies are shown in Fig. 6e–h, but enlarged to focus on the EA domain and adjacent regions. The percentages of the grid boxes in the EA domain that are associated with significantly-changed precipitation is indicated in the upper left of each panel.

**Fig. 6** Anomalous precipitation (mm/day) for the **a** WIO, **b** CIO, **c** EIO, and **d** ALL simulations for October 1–December 15 climatology. Only areas at or above the 90% confidence level are shaded. **e–h** are the same as **(a)**, **(b)**, **(c)** and **(d)**, respectively, but near and in the analysis domain. The black contours show country outlines. Black boxes denote the EA analysis domain. The percentages of the grid boxes in the EA analysis domain that are associated with significantly-changed precipitation is indicated in the upper left of each panel



Precipitation anomalies associated with a warm western Indian Ocean (Fig. 6a) cover a large region of tropical EA (22°–41°E, 10°S–8°N), with decreased precipitation in the Arabian Sea (60°–80°E, 6°–15°N) and the south Indian

Ocean (80°–100°E, 7°–16°S). 95% of the grid points in the EA analysis domain have significantly-increased precipitation (0–4 mm/day) due to the warm western Indian Ocean (Fig. 6e). Eastern Kenya, southern Somalia, and southern

Ethiopia also show significantly-increased precipitation (1–2 mm/day).

In the CIO simulation, precipitation increases over the imposed SSTA as in the WIO simulation, and there is decreased precipitation over the western and eastern Indian Ocean near the equator (Fig. 6b). The dry region in the western Indian Ocean extends onto the African coast (southern Somalia), and precipitation responses are not homogeneous over the EA domain (Fig. 6f). The north-eastern quadrant (29% of the EA domain) becomes significantly drier due to the warm central Indian Ocean SSTA forcing, while regions west of 38°E (8% of the EA domain) become significantly wetter by 0–2 mm/day.

When cool SSTAs are imposed in the eastern Indian Ocean in the EIO ensemble (Fig. 6c), precipitation decreases over the SSTAs, and increases over the western equatorial Indian Ocean (40°–70°E, 0°–8°S) and to the north of the cold anomaly (80°–110°E, 0°–5°N). 30% of the EA domain shows significantly increased precipitation (Fig. 6g).

Comparing the precipitation responses in the three simulations, the percentages of the grid boxes in the EA domain that are associated with significantly-changed precipitation in the WIO ensemble are about three times those in the CIO and EIO ensembles.

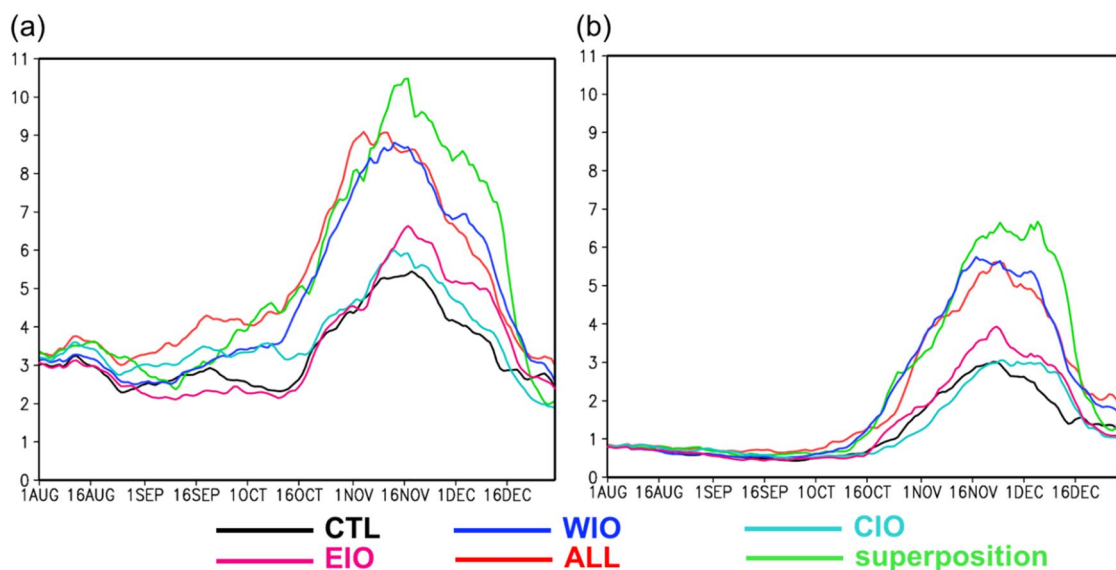
ALL simulations are significantly wetter over the western and central IO, and the EA (22°–55°E, 10°S–10°N), including Kenya, Somalia, Ethiopia, and Uganda (Fig. 6d), and drier over the southeastern IO and the Bay of Bengal than the control simulations. 92% of the grid points in the EA

analysis domain have significantly-increased precipitation (Fig. 6h), similar to the WIO simulations (Fig. 6e).

Since the precipitation anomalies shown in Fig. 6 are averaged over two and a half months, they could represent changes in magnitudes, or/and changes in the timing of the onset or demise of the rainy season. Figure 7a shows the 11-day running mean of precipitation averaged over the western EA domain (30°–35°E, 5°S–5°N) in the CTL (black), WIO (dark blue), CIO (light blue), EIO (pink), and ALL (red) climatologies, along with the linear superposition of the WIO, CIO, and EIO ensembles (green). The WIO ensemble (dark blue line) shows significantly (> 90% confidence level) increased precipitation compared with the CTL ensemble (black line) with a maximum of 9 mm/day on November 11.

The CIO (light blue line) and EIO (pink line) simulations also show increased rainfall compared with the CTL ensemble but in smaller magnitudes than for the WIO ensemble. Precipitation anomalies are up to 1.1 mm/day in the CIO ensemble and 0.8 mm/day in the EIO ensemble. However, the CIO and EIO simulations do not produce significant (90% confidence level) changes of precipitation from the control climatology during the short rain season. The precipitation changes are significant only on specific dates rather than over the entire period like for WIO.

The response of EA precipitation to the imposition of all three SSTAs together (the ALL ensemble; red line) does not equal the sum of the responses to the SSTAs individually (green line). When multiple SSTAs are applied, nonlinearities are introduced. The maximum precipitation anomaly in



**Fig. 7** a 11-day running mean of precipitation averaged over the western domain (30°–35°E, 5°S–5°N) in CTL (black), WIO (dark blue), CIO (light blue), EIO (pink), ALL (red), and the linear super-

position of the WIO, CIO, and EIO ensembles (green). b Same as a but averaged over the eastern domain (35°–40°E, 5°S–5°N). Unit: mm/day



ALL is smaller than the superposition by 1.5 mm/day. The daily evolution of precipitation in the ALL simulation is more similar to the WIO simulation in magnitude than to the superposition of the three SSTAs. Although the ALL ensemble changes the large-scale zonal gradient of Indian Ocean SSTs, the response of EA precipitation is not significantly different from the WIO ensemble.

Figure 7b shows the 11-day running means of precipitation averaged over the eastern EA domain ( $35^{\circ}$ – $40^{\circ}$ E,  $5^{\circ}$ S– $5^{\circ}$ N) in multiple simulations similar to Fig. 7a. Eastern-domain precipitation anomalies in the perturbation simulations are similar to those in the western EA domain (Fig. 7a) but with smaller magnitudes. The statistical significance of the precipitation responses in the different ensembles are also similar to those in the west (Fig. 7a), with significantly-increased short rain precipitation in the WIO ensemble and insignificantly increased precipitation in the CIO and EIO.

Both Figs. 6 and 7 indicate that western Indian Ocean SSTAs exert a stronger influence on EA short rains than central and eastern Indian Ocean SSTAs. The central and eastern Indian Ocean SSTAs produce significantly-changed precipitation over some regions of the tropical Indian Ocean, but their influence on African precipitation in boreal fall is less than the western Indian Ocean SSTAs.

### 5.3 Mechanisms of the Indian Ocean SSTs impacting the EA short rains

To connect precipitation responses in the WIO simulations to circulations, Fig. 8 depicts anomalies from the CTL ensemble in the components of the atmospheric moisture budget [Eq. (1)] for the WIO simulations. Positive  $P - E$  anomalies (Fig. 8a) exist over the Congo Basin, the EA analysis domain, and the western Indian Ocean. Evaporation anomalies (not shown) are not influential to precipitation anomalies with magnitudes less than 1 mm/day. The magnitudes of  $A$  anomalies (Fig. 8b) are generally under 1 mm/day, except for 2–3 mm/day near the Turkana Channel. The anomalies of the residual term (Fig. 8c) show structures along the topography. The increased  $P - E$  (Fig. 8a) extending from the Congo basin to the western Indian Ocean is not explained by  $A$  (Fig. 8b) and  $R$  anomalies (Fig. 8c).

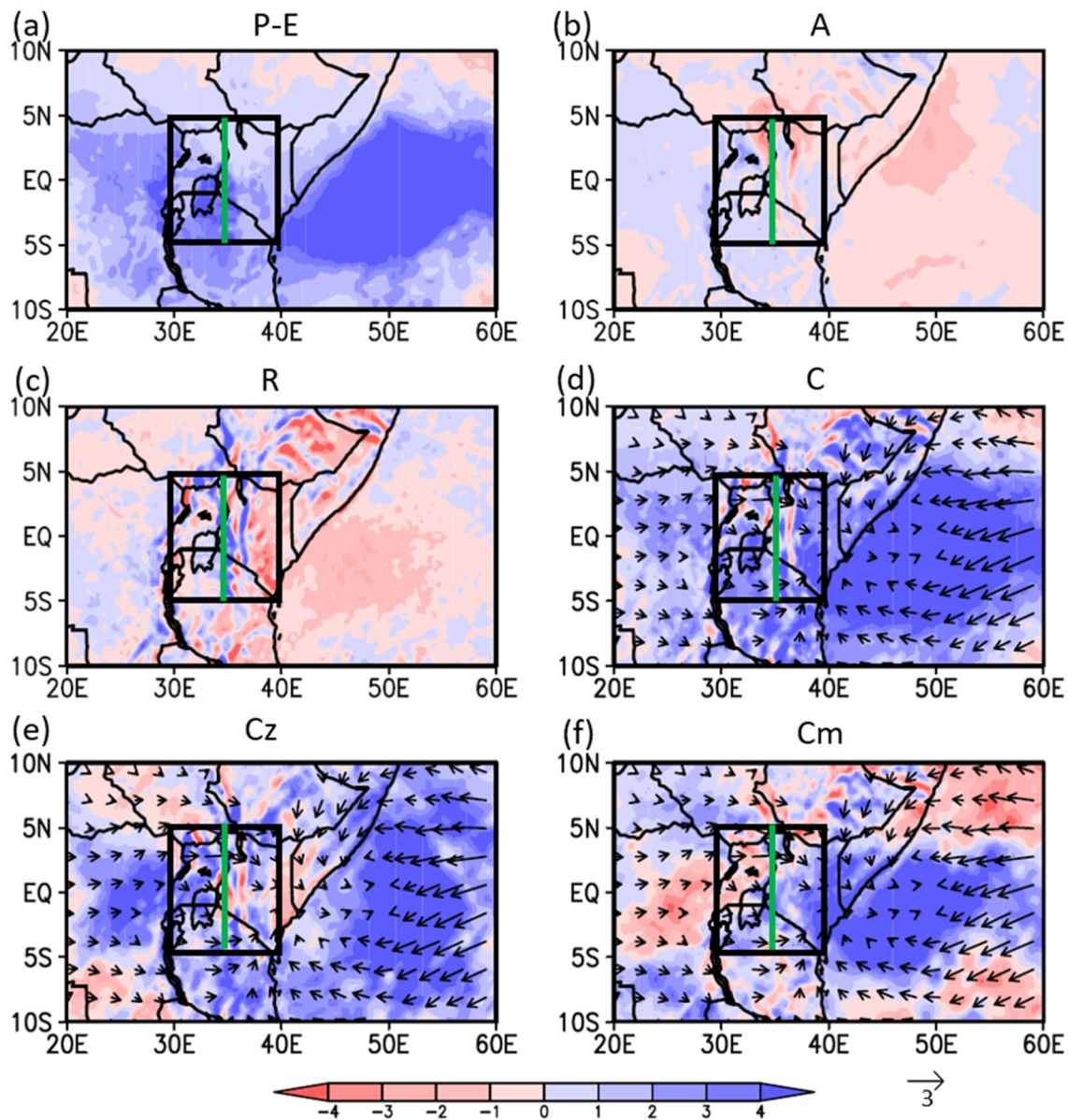
Figure 8d shows the difference in moisture convergence term [Eq. (2)] between the WIO and CTL simulations, representing the convergence of wind perturbation in a moist environment, along with anomalous winds at 850 hPa. Winds at 850 hPa are characteristic to represent low levels (975–700 hPa; not shown). The anomaly of  $C$  term most closely resembles  $P - E$  anomaly (Fig. 8a). Anomalous wind convergence is found over equatorial Africa on the east of  $20^{\circ}$ E and the western Indian Ocean. The warm western

Indian Ocean SSTAs strengthen the wind convergence in a moist environment over the western EA domain in the CTL (Fig. 4d) and weaken the wind divergence over the eastern EA domain (Fig. 4d). The anomalous 850-hPa winds (Fig. 8d) are mainly zonal from the Congo basin to the western EA domain while they are more meridional along the African coast (Somalia and eastern Tanzania).

Figure 8e and f show zonal [Eq. (3)] and meridional [Eq. (4)] components of  $C$  anomaly, respectively, along with the same anomalous winds as Fig. 8d. Patterns of both components indicate complicated heterogeneity. Both the eastern and western parts of the EA domain show positive  $P - E$  anomalies due to the warm western Indian Ocean SSTAs (Fig. 8a), but they are controlled by different components of  $C$  anomaly. Specifically, 70% of anomalous  $P - E$  (Fig. 8a) over the western EA domain is supported by  $C_z$  (Fig. 8e). Over the eastern EA domain, 125% of the anomalous  $P - E$  (Fig. 8a) is supported by  $C_m$  (Fig. 8f) but it is suppressed by  $R$  term (Fig. 8c). The meridional winds from southern and northern hemispheres are convergent on the equator near the coast over southern Somalia and eastern Kenya (Fig. 8f), associated with positive  $C_m$  anomalies over the eastern EA domain. The westerlies from the Congo basin proceed to the northern and southern sides of Mt. Kenya, associated with negative  $C_m$  anomalies over ( $30^{\circ}$ – $34^{\circ}$ E,  $1^{\circ}$ S– $5^{\circ}$ N) (Fig. 8f).

To understand the changes of low-level winds in large-scale circulation, Fig. 9 shows anomalous geopotential heights and winds at 850 hPa in the WIO ensemble. Geopotential height anomalies and wind anomalies show patterns similar to the response to steady thermal forcing on the equator from the shallow water equations (Gill 1980). The western Indian Ocean heating generates negative geopotential height anomalies in the tropical Indian Ocean between  $10^{\circ}$ N and  $10^{\circ}$ S and the western Indian Ocean, over which the geopotential height anomalies are not centered exactly on the equator but in the southern hemisphere. A possible explanation for this asymmetry is that land north of the equator (e.g., the Horn of Africa) extends to the east and the western Indian Ocean SSTAs (SSTA1 in Fig. 5c) spatially cover a greater area south of the equator, hence, the SST forcing is weighted more towards the southern hemisphere. Positive geopotential height anomalies (0–4 gpm) are found over India, southeastern Asia, and the southern Indian Ocean centered at ( $75^{\circ}$ E,  $14^{\circ}$ S).

Anomalous zonal flow is confined near the equator ( $10^{\circ}$ S– $10^{\circ}$ N), consistent with Gill (1980). Considering the generations of the Rossby and Kelvin waves in Gill (1980), the Kelvin wave to the east of the western Indian Ocean heating maximum (yellow dot) produces anomalous zonal flow into the heating close to the equator (Fig. 9) across the equatorial Indian Ocean. The equatorial zonal flow to the



**Fig. 8** Anomalous **a** precipitation minus evapotranspiration, **b** moisture advection term, **c** residual term, **d** moisture convergence term, **e** zonal moisture convergences term, and **f** meridional moisture convergence term in the moisture budget [Eq. (1)] for the WIO ensemble climatology. Unit: mm/day. Vectors in (d)–(f) are anomalous winds

( $\text{m s}^{-1}$ ) at 850 hPa for the WIO ensemble climatology. Topography is masked out. Black boxes denote the full EA analysis domain. Green lines are the boundaries between western and eastern EA analysis domain

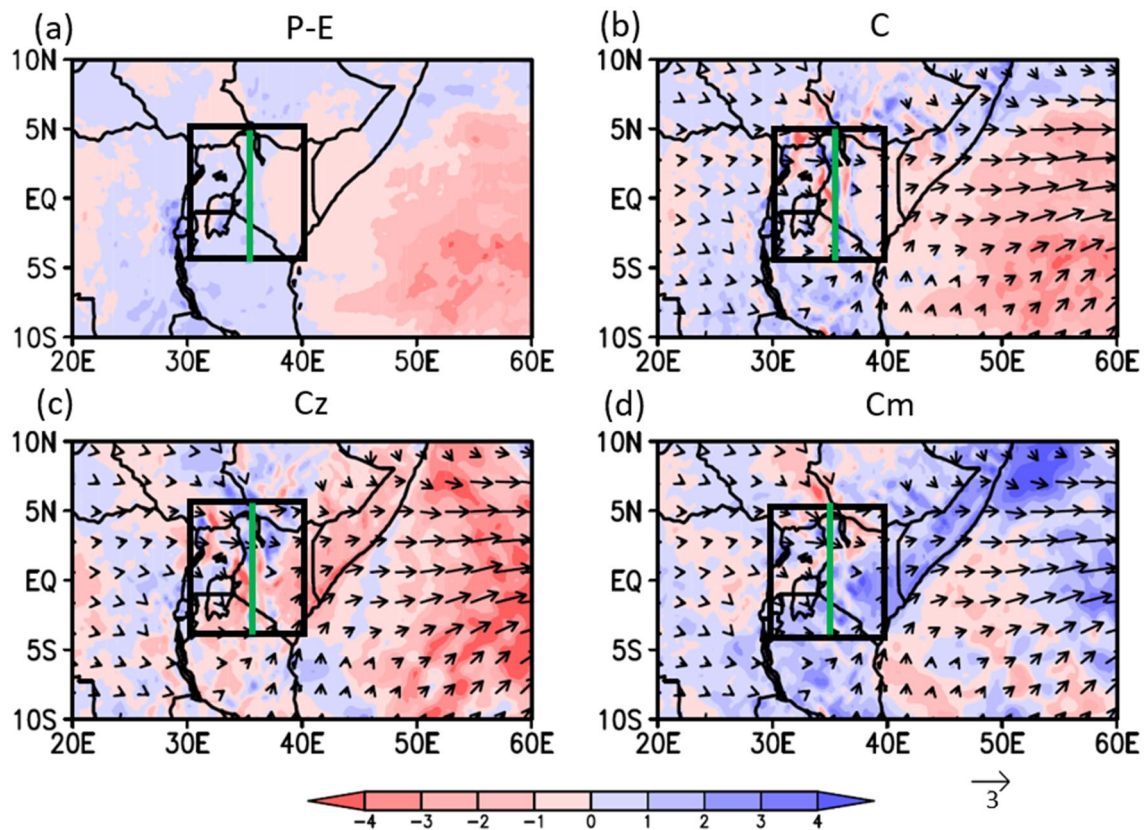
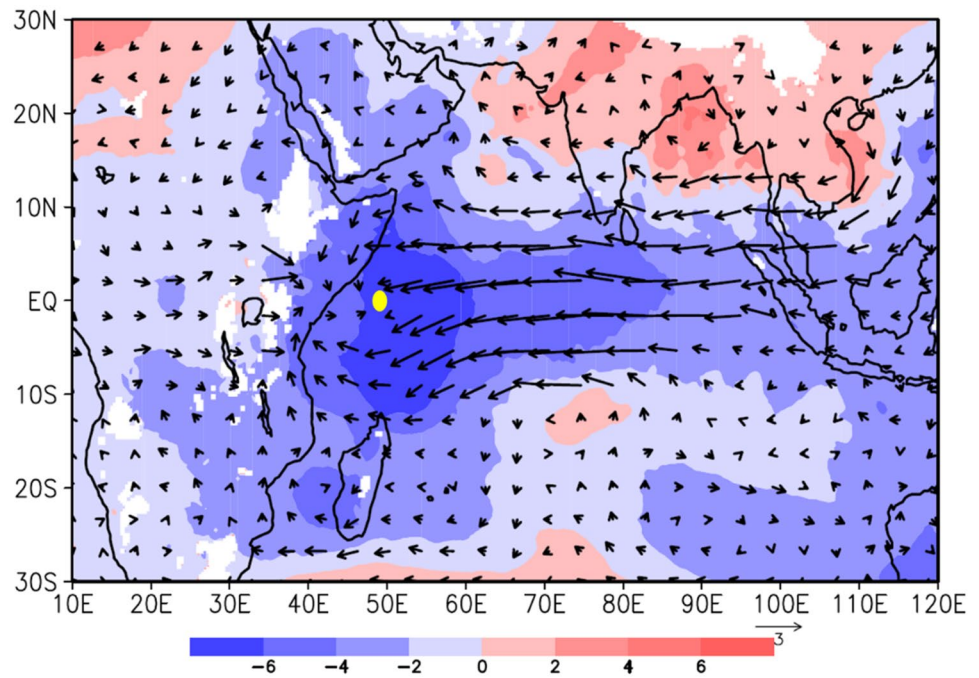
west of the heating are weaker than to the east, different from the classic Gill type responses (1980), possibly due to the influence of the highlands over East Africa and Ethiopia (up to 700 hPa). The anomalous westerlies/southwesterlies over the Turkana Channel (Fig. 9) are associated with the negative *A* anomalies over the same region (Fig. 8b). Rossby wave response occurs to the west of western Indian Ocean heating. The presence of the Rossby mode is indicated by the

cyclonic anomaly over (35°–55°E, 0°–15°S) and meridional wind perturbations over (40°–50°E, 0°–15°N), associated with the equatorward wind perturbations and the positive *Cm* anomalies over the coast (Fig. 8f).

The classic Gill model (1980) assumes a quiescent basic state while the basic-state flow is non-zero in our case. Considering the basic-state flows in the control simulations (Fig. 3c, f), the western Indian Ocean heating weakens the



**Fig. 9** Anomalous geopotential heights (shaded; gpm) and winds (vectors;  $\text{m s}^{-1}$ ) at 850 hPa in the WIO ensemble for October 1–December 15 climatology. The yellow dot shows the location of the maximum of SSTAs in the WIO simulations. Topography is masked out



**Fig. 10** Anomalous **a** precipitation minus evapotranspiration, **b** moisture convergence term, **c** zonal moisture convergences term, and **d** meridional moisture convergence term in the moisture budget [Eq. (1)] for the CIO ensemble climatology. Unit: mm/day. Vectors

in **d–f** are anomalous winds ( $\text{m s}^{-1}$ ) at 850 hPa for the CIO ensemble climatology. Topography is masked out. Black boxes denote the full EA analysis domain. Green lines are the boundaries between western and eastern EA analysis domain



equatorial westerlies over the Indian Ocean and the southeasterlies over the Turkana Channel by producing zonal flows in the Kelvin wave response, and strengthens the equatorward flows over the African coast (Somalia and eastern Tanzania) associated with the meridional perturbations in the Rossby wave.

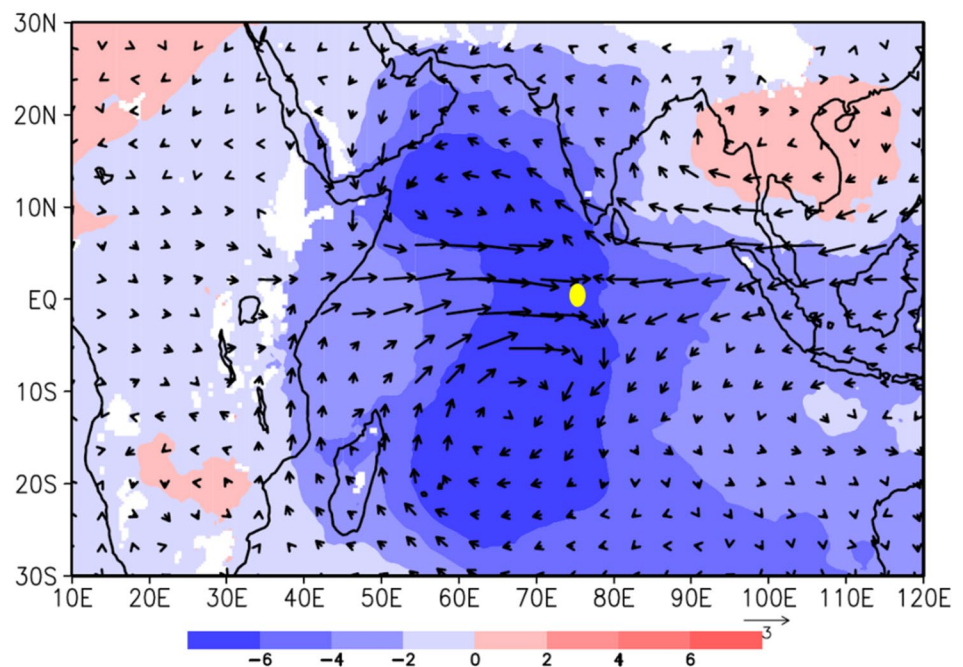
To connect precipitation responses in the CIO simulations to circulations, Fig. 10 shows the CIO minus CTL differences in  $P - E$ ,  $C$ ,  $Cz$ , and  $Cm$  terms in the moisture budget. Vectors in Fig. 10d–f are anomalous winds at 850 hPa for the CIO ensemble average. Unlike the WIO simulations, the anomalous  $P - E$  in CIO simulations is inhomogeneous with positive  $P - E$  anomaly over the western EA domain, and negative  $P - E$  anomaly over the eastern EA domain and the western Indian Ocean. The magnitudes of  $P - E$  over the continent ( $< 1$  mm/day) in the CIO simulations are smaller than the WIO simulations (1–6 mm/day). The anomalous  $A$  and  $R$  terms (not shown) are not influential to the patterns of  $P - E$  anomalies, similar to Fig. 8b, c.

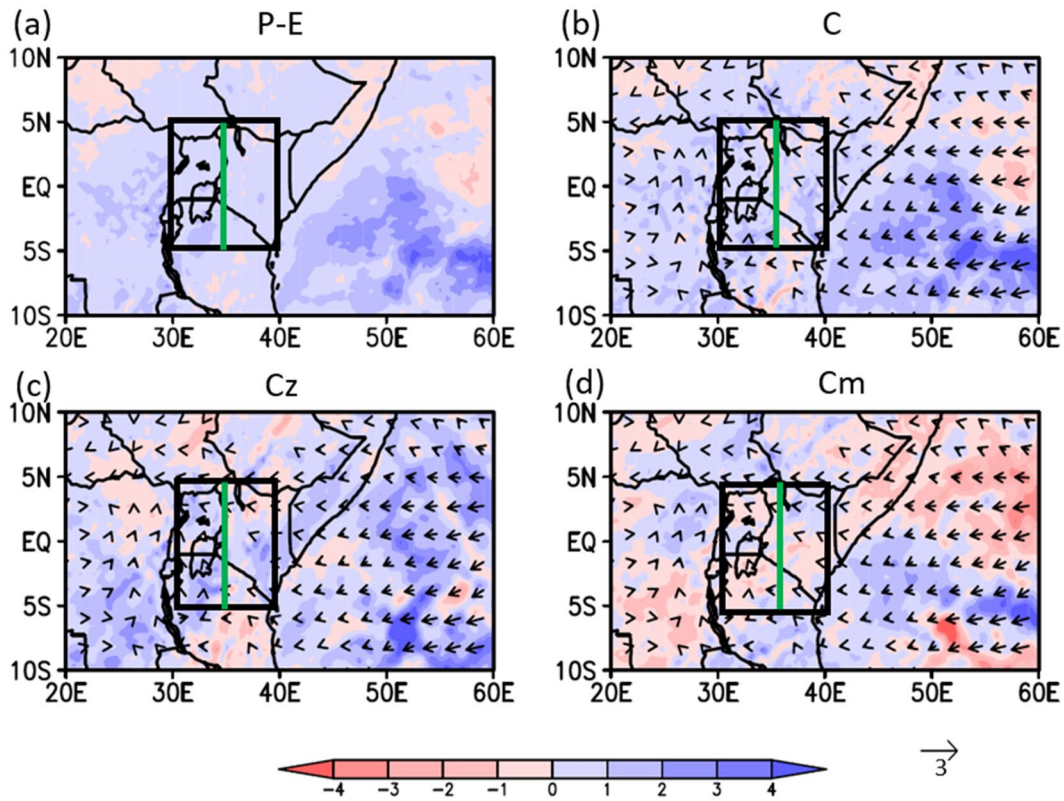
The pattern of the anomalous  $C$  term (Fig. 10d) resembles  $P - E$  anomalies (Fig. 10a), similar to WIO simulations, except for structures along the topography near  $35^\circ\text{E}$ . The negative  $P - E$  anomaly over the eastern EA domain and the ocean (Fig. 10a) is dominated by zonal wind divergence (Fig. 10c), associated with offshore wind anomalies. Positive  $Cm$  anomalies (Fig. 10d) are found over Somalia, northern Kenya, and Tanzania, associated with meridional wind convergence, but the magnitudes of  $Cm$  anomalies are smaller than  $Cz$  (Fig. 10c) by approximately 0.3 mm/day.

Figure 11 shows differences in geopotential heights and winds at 850 hPa between the CIO and CTL simulations. Similar to the classic Gill model, anomalous westerlies near the equator ( $5^\circ\text{S} - 5^\circ\text{N}$ ) occur on the west of central Indian Ocean SSTAs and anomalous easterlies exist on the east of central Indian Ocean SSTAs in the presence of Kelvin wave. These anomalous equatorial zonal flows are associated with anomalous zonal wind convergence in a moist environment over the central Indian Ocean (not shown) and, therefore, the significantly-increased precipitation (Fig. 6b). The anomalous westerlies from the African continent to the central Indian Ocean heating are associated with a strengthening of the offshore low-level flow over equatorial East Africa in the CTL (Fig. 4c), associated with the negative  $Cz$  anomalies near the coast (Fig. 10c). There are two cyclonic anomalies centered at ( $65^\circ\text{E}, 15^\circ\text{S}$ ) and ( $65^\circ\text{E}, 8^\circ\text{N}$ ) on the southwest and northwest of the central Indian Ocean SSTA maximum (yellow dot) in the presence of Rossby wave. Poleward wind anomalies over  $75^\circ - 85^\circ\text{E}$  in both hemispheres are meridional perturbations of Rossby wave. The return flows of the Rossby wave response are the equatorward flows over the African coast, associated with the positive  $Cm$  anomalies over Somalia, northern Kenya, and Tanzania (Fig. 10d).

Figure 12 shows the EIO minus CTL differences in  $P - E$ ,  $C$ ,  $Cz$ , and  $Cm$  terms in the moisture budget. Anomalous  $P - E$  (Fig. 12a) increases by 0–1 mm/day over the EA domain, which is much less than in the WIO simulations. Again, the anomalous  $C$  term (Fig. 12b) resembles the anomalous  $P - E$  term (Fig. 12a) with changes in  $Cz$  (Fig. 12c) mostly responsible for the positive anomalies

**Fig. 11** Anomalous geopotential heights (shaded; gpm) and winds (vectors;  $\text{m s}^{-1}$ ) at 850 hPa averaged over October 1–December 15 in the CIO ensemble. The yellow dot shows the location of the maximum of SSTAs in the CIO simulations. Topography is masked out

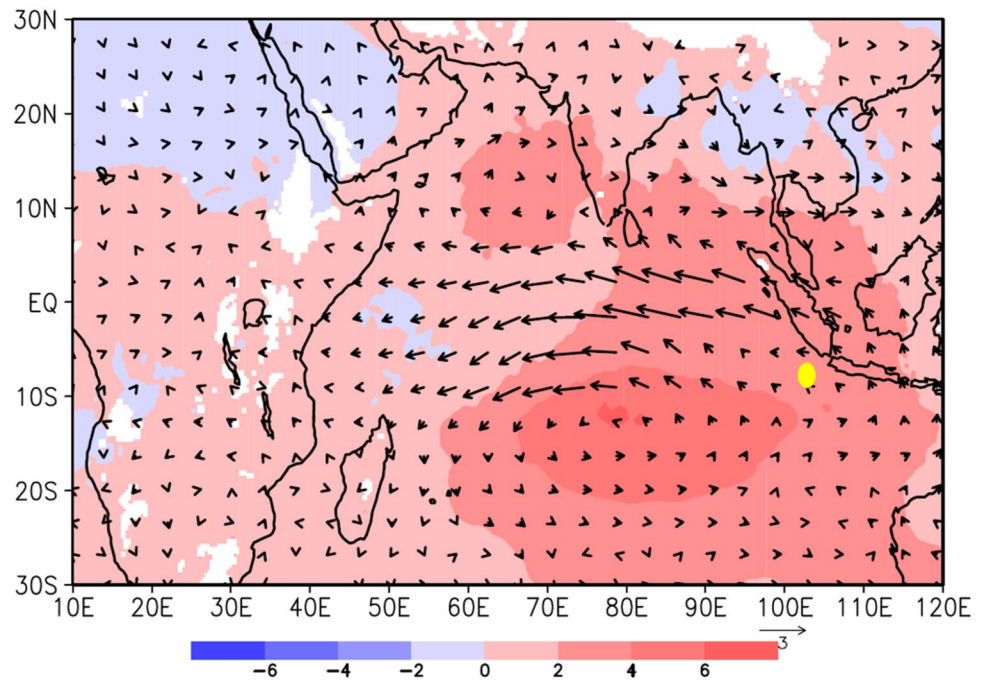




**Fig. 12** Anomalous **a** precipitation term minus evapotranspiration term, **b** moisture convergence term, **c** zonal moisture convergences term, and **d** meridional moisture convergence term in the moisture budget [Eq. (1)] for the EIO ensemble climatology. Unit: mm/day.

Vectors in **(b)–(d)** are anomalous winds ( $\text{m s}^{-1}$ ) at 850 hPa for the EIO ensemble climatology. Topography is masked out. Black boxes denote the full EA analysis domain. Green lines are the boundaries between western and eastern EA analysis domain

**Fig. 13** Anomalous geopotential heights (shaded; gpm) and winds (vectors;  $\text{m s}^{-1}$ ) at 850 hPa in the EIO ensemble. The yellow dot shows the location of the minimum of SSTAs in the EIO simulations. Topography is masked out



while being slightly offset by  $C_m$  (Fig. 12d) as there is anomalous low-level easterly flow from the Indian Ocean into the equatorial East Africa. That being said, the anomalous flow is weaker in magnitude for EIO compared to the WIO and CIO cases.

Figure 13 shows 850-hPa geopotential height and wind anomalies for the EIO ensemble. The anomalous low-level height response is asymmetric about the equator as the center of the SSTAs is located more off the equator ( $10^{\circ}\text{S}$ ) for EIO compared to the WIO and CIO cases. There is a Gill-like anomalous anticyclonic anomaly pattern to the west of eastern Indian Ocean SSTAs with the stronger height anomalies in the southern hemisphere. Along the equator, the anomalous flow is easterly, indicating a weakening of the low-level equatorial westerly flow across the basin (Fig. 3c). While the anomalous equatorial zonal flow is still easterly to the west of  $55^{\circ}\text{E}$ , it is much weaker than over the eastern half of the Indian Ocean basin as the SST forcing is much further away compared to the other cases.

To compare changes in the instability of the atmosphere due to different SSTAs, Fig. 14 displays anomalous MSE (black),  $c_p T$  (green), and  $Lq$  (red) [Eq. (5)] averaged over the western (solid) and eastern EA domains (dashed) in the WIO, CIO, and EIO simulations. Note the  $g_z$  component anomalies are minimal, and thus are not shown in Fig. 14. MSE in the CTL simulation average (not shown) decreases from surface to 600 hPa over both western and eastern EA domains indicating the low-level atmosphere is unstable.

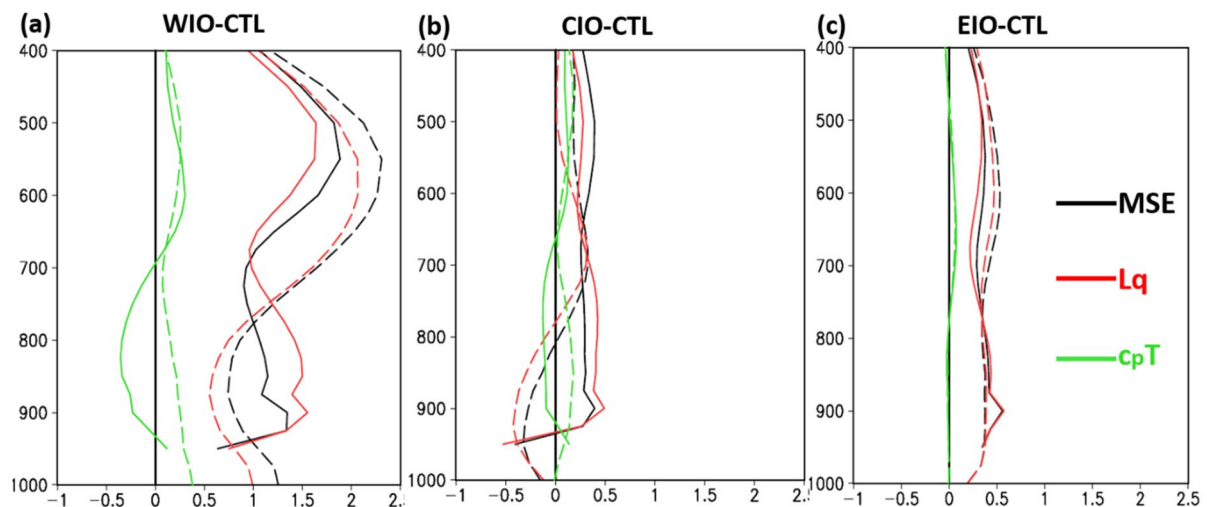
For WIO (Fig. 14a), anomalous low-level MSE (black line) has a negative slope over both domains but at different levels. For example, increases in atmospheric instability over

the western domain (black solid line) occur between 925 and 700 hPa, but occur between 1000 and 850 hPa over the eastern EA domain (black dashed line). Clearly, the elevation differences between these two regions has a considerable impact (Fig. 1b). Changes in the vertical MSE profiles over both domains (black line) are primarily associated with changes in  $Lq$  (red line). Low-level atmospheric moisture content is enhanced due to the warm western Indian Ocean SSTAs and thus leads to a destabilization of the low-level atmosphere.

For CIO (Fig. 14b), the vertical profile of anomalous MSE is neutral over the western EA domain (black solid line), but the atmosphere between 900 and 700 hPa is more stable over the eastern EA domain (black dashed line), consistent with the wetter west region and drier east region rainfall pattern in Fig. 6f. Anomalous MSE (black line) is dominated by changes in  $Lq$  (red line) over both domains as the low-level moisture (surface to 800 hPa) increases over the western EA domain, and decreases over the eastern EA domain.

For EIO (Fig. 14c), changes in  $c_p T$  are approximately zero indicating minimal temperature changes over the two EA domains. The MSE profile is, therefore, supported by changes in  $Lq$ . Anomalous  $Lq$  (red line) is positive from the surface to 400 hPa indicating enhanced atmospheric moisture. The eastern Indian Ocean SSTAs increase atmospheric moisture, but only generate a neutral vertical profile of MSE (black line), less influential than the western SSTAs (Fig. 14).

Changes in vertical speeds over the EA domain and the Indian Ocean for the different simulations are also



**Fig. 14** Anomalous MSE (black),  $c_p T$  (green), and  $Lq$  (red) in **a** WIO, **b** CIO, and **c** EIO simulations averaged over western EA domain ( $30^{\circ}$ – $35^{\circ}\text{E}$ ,  $5^{\circ}\text{S}$ – $5^{\circ}\text{N}$ ) (solid) and eastern EA domain ( $35^{\circ}$ –

$40^{\circ}\text{E}$ ,  $5^{\circ}\text{S}$ – $5^{\circ}\text{N}$ ) (dashed) averaged over October 1–December 15. Topography is masked out. Unit:  $10^3 \text{ m}^2 \text{ s}^{-1}$



investigated. Results from this analysis (not shown) show that the warm western Indian Ocean SSTAs are associated with the strongest anomalous upward vertical motions over the EA domain, indicating a weakening of the subsidence near the EA coast. Central and eastern Indian Ocean SSTAs induce anomalous vertical motions over the ocean, but their influence on the EA subsidence is minor.

## 6 Summary and conclusions

The boreal fall short rains over equatorial East Africa (30°–40°E, 5°S–5°N) exhibit considerable interannual variability and regionality over the complex topography (Black et al. 2003; Hastenrath et al. 2011; Lyon 2014). Past studies (e.g., Behera et al. 2005; Ummerhofer et al. 2009) associate this interannual variability with SST variations in the Indian Ocean. Our purpose is to identify which regions of the Indian Ocean most strongly influence the boreal fall equatorial East African short rains on interannual timescales, and to investigate the physical mechanisms responsible.

Three “hot spot” regions are identified, located on or near the equator in the western, central, and eastern Indian Ocean. A regional climate model (WRF) with 30-km horizontal resolution is used to conduct 5 sets of simulations in which 20-member ensembles represent a control climate and climates with idealized SSTAs. The control simulations use climatological (1998–2017) SSTs. In the idealized simulations, various Gaussian-shaped SSTAs derived from observed correlations between East African precipitation and Indian Ocean SSTs are imposed (Fig. 5c). One additional ensemble includes all three SSTAs to evaluate the potential for interference among the individually-forced responses.

The results are summarized as follows:

- The control simulations capture the observed patterns and seasonality of East African precipitation (Fig. 2) and circulation (Fig. 3) with reasonable accuracy, albeit with a wet bias over the Congo basin. The simulated processes that support regional rainfall as revealed by the atmospheric moisture budget also agree well with reanalyses. This analysis shows that the mechanisms that support precipitation in the western and eastern parts of the equatorial East Africa analysis domain are fundamentally different due to the presence of complex topography (Fig. 4). The short rains over the western half of the domain are mainly supported by wind convergence in a moist environment, while orographic uplift plays a dominant role farther east.
- Short rains over the analysis domain are positively correlated with western and central Indian Ocean SSTs, and negatively correlated with eastern Indian Ocean SSTs in observations from 1998–2017 (Fig. 5). The western and eastern Indian Ocean SSTAs are not significantly anti-correlated at the 90% confidence level, consistent with some past studies that also reveal an absence of the Indian Ocean Dipole as a coherent mode of variability (Nicholls and Drosowsky 2001; Dommengot and Latif 2002; Zhao and Nigam 2015; Wang et al. 2019).
- SSTAs in the western Indian Ocean are shown to exert a stronger influence on the equatorial East African short rains than central and eastern SSTAs in terms of both the coverage of significantly-changed precipitation (Fig. 6) and the magnitude of the precipitation response (Fig. 7), consistent with past studies (Ummerhofer et al. 2009; Bahaga et al. 2015). Specifically, western Indian Ocean SSTAs produce significantly enhanced precipitation over 95% of the analysis domain, while only 30% of the region responds to central and eastern Indian Ocean SSTAs (Fig. 6). The maximum precipitation anomaly associated with western Indian Ocean SSTAs is about three times larger than the anomalies due to central and eastern SSTAs in the simulation (Fig. 7). The response of the precipitation field to the three SSTAs together does not equal the sum of the responses to the SSTAs individually indicating the presence of nonlinearity (Fig. 7).
- The atmospheric moisture budget analysis shows that the anomalous wind convergence closely resembles the precipitation anomaly in all cases (Figs. 8, 10, 12). The patterns of geopotential heights and winds anomalies caused by positive western and central Indian Ocean SSTAs are similar to the Gill response (1980) to steady heating on the equator (Figs. 9, 11), while the simulations with negative eastern Indian Ocean SSTAs are similar to the off-equator cooling case (Fig. 13). The main differences between the classic Gill response (1980) and our cases are due to the presence of complex topography (especially in the simulations with western Indian Ocean SSTAs) and background flows.
- The increased precipitation over the western domain due to western Indian Ocean SSTAs is mainly supported by zonal wind convergence anomalies, associated with anomalous westerlies from the Congo basin in the Kelvin wave response. The meridional wind convergence anomaly contributes to increased precipitation over the eastern domain most strongly, in association with the meridional perturbations of the Rossby wave (Fig. 8).
- The vertical profiles of anomalous MSE are dominated by the moisture component in all three cases (Fig. 14). The increases in low-level moisture that accompany western Indian Ocean warming destabilize the low-level atmosphere over equatorial East Africa. SSTAs in the central and eastern Indian Ocean also modify the region's atmospheric moisture, but they are less influential than western Indian Ocean SSTAs in changing atmospheric instability.

This study shows that western Indian Ocean SSTAs are much more influential for perturbing the equatorial East African short rains on interannual time scales than SSTAs elsewhere in the Indian Ocean. In particular, we find no evidence of connections between an Indian Ocean dipole mode and equatorial East African rainfall. The mechanisms of the forcing from the western Indian Ocean are related to the shallow-water Gill model (1980), but in the presence of complicated topography and nonzero background flows. East African rainfall, including its supporting mechanisms, is highly regional, encouraging caution when choosing averaging regions in future studies of precipitation variability on all time scales. Our study designed the idealized SSTAs based on the correlations of the East African short rains and Indian Ocean SSTs in the current climatology (1998–2017). The next step is to consider the decadal and multi-decadal variability of this correlation and investigate the relationship of Indian Ocean SSTs and East African short rains over a longer time period to better understand the robustness of the results presented here.

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