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Authors

Sena, ACT
Magnusdottir, G

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Influence of the Indian Ocean Dipole on the Large-Scale Circulation in South America[✉]

ANA C. T. SENA^a AND GUDRUN MAGNUSDOTTIR^a

^a *Department of Earth System Science, University of California, Irvine, Irvine, California*

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ABSTRACT: The influence of each phase of the Indian Ocean dipole (IOD) on the large-scale circulation in South America is investigated using rainfall observations, fully coupled, large-ensemble, historical simulations (LENS), and forced experiments using the coupled model's atmospheric component. IOD events often occur when El Niño–Southern Oscillation (ENSO), the largest source of interannual variability of precipitation in South America, is active. To distinguish from effects of ENSO, only cases during neutral ENSO conditions are analyzed in LENS and observations. During the positive IOD polarity, a perturbation in the local Walker circulation leads to increased convection over equatorial South America, resulting in wet anomalies in the Amazon basin. This signal is the opposite of what is expected during El Niño events. Tropical convection anomalies in the Indian Ocean also force an extratropical Rossby wave train that reaches subtropical South America. During positive IOD, the moisture flux from the Amazon to central and southeastern Brazil weakens, resulting in a drying of the area associated with the South Atlantic convergence zone. Meanwhile, the South Atlantic subtropical high strengthens, contributing to a drying in southeastern Brazil. During negative IOD, the induced wave train from the Indian Ocean leads to increased moisture transport to the La Plata basin, leading to wet anomalies in the region.

KEYWORDS: Atmosphere; Indian Ocean; South America; Dynamics; Teleconnections; Interannual variability

1. Introduction

The Indian Ocean dipole (IOD) consists of a dipole in sea surface temperature (SST) anomalies in the Indian Ocean. During its positive phase, warm SST anomalies are observed in the western part of the Indian Ocean, while cool SST anomalies occur in the eastern sector. During the negative phase, these anomalies are reversed (Saji et al. 1999). The IOD usually starts developing around June, peaks in September and October, and rapidly decays during the austral summer (Lim and Hendon 2017).

IOD events often occur when El Niño–Southern Oscillation (ENSO) is active. Positive IOD is often concurrent with the positive phase of ENSO, while the negative phases of IOD and ENSO often occur in the same years (Sun et al. 2015; Wang and Wang 2014). There is an ongoing debate whether or not IOD and ENSO are really two separate physical phenomena. Some studies suggest that the IOD develops not as an independent process, but as a consequence of the wind anomalies in the eastern sector of the Indian Ocean generated by ENSO (Stuecker et al. 2017; Zhao and Nigam 2015). However, Wang et al. (2016) argue that IOD can be generated even in the absence of ENSO as a result of anomalies in Indonesian convection amplified by surface wind response to the precipitation anomalies and subsurface conditions. There is also evidence that IOD may precede ENSO's development and contribute to it (Wang et al. 2019; Izumo et al. 2010; Annamalai et al. 2005). In this paper, we examine the remote effects of the SST pattern in the Indian Ocean corresponding to the IOD on the South American circulation, regardless of how the IOD pattern may

have been generated. Henceforth, we will refer to IOD events during ENSO neutral conditions as pure IOD events.

ENSO is the dominant source of interannual variability of rainfall over South America and it is challenging to isolate the influence of other modes of climate variability in the short observational record. El Niño is commonly associated with drought over the Amazon and wetting over southern Brazil. The opposite signals are associated with La Niña (Cai et al. 2020; Grimm 2011).

In addition to climate influences in adjacent areas such as the Indian monsoon region (Ashok et al. 2004) and Australia (Ashok et al. 2003; Cai et al. 2011), convection associated with SST anomalies over the Indian Ocean can generate Rossby waves that disturb the atmospheric circulation over remote locations such as Antarctica (Nuncio and Satheesan 2014; Li et al. 2015; Feng et al. 2019) and South America (Drumond and Ambrizzi 2008; Chan et al. 2008; Saji et al. 2005).

Earlier studies have shown that the circulation in South America is influenced by SST anomalies in the Indian Ocean. During the austral summer, Drumond and Ambrizzi (2008) found in modeling experiments that a monopole of warm SST anomalies in the southern Indian Ocean is associated with drier conditions in the areas commonly related to the South Atlantic convergence zone (SACZ) and wetter conditions in southern Brazil. Chan et al. (2008) found that the positive phase of the IOD is linked to negative rainfall anomalies over the areas commonly associated with the SACZ in a coupled ocean–atmosphere global climate model. They isolate the IOD from ENSO influences by selecting only years with an active IOD during neutral ENSO. As ENSO and IOD SST anomalies often occur simultaneously, the number of cases analyzed in that study is reduced and no significant anomalies were found during negative IOD events. These studies did not find significant precipitation anomalies in tropical South America.

Although the above studies are focused on effects during the austral summer, the IOD usually reaches its peak during the

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Corresponding author: Ana C. T. Sena, athomese@uci.edu

austral spring, which coincides with the onset of the South American monsoon (Sena and Magnusdottir 2020; Gan et al. 2005). The IOD has also been linked to surface temperature anomalies in South America during the austral spring, with warm anomalies in the subtropical South America during positive IOD events and cold anomalies in the area during negative events (Saji et al. 2005).

The objective of this study is to get a better understanding of the mechanisms by which the sea surface temperature patterns associated with the IOD and the associated perturbed convection can influence the large-scale circulation in the South American monsoon during austral spring and summer. First, we analyze how rainfall over South America responds to pure IOD events in observations. As the observational record is short and an extended record is only available over land areas, we proceed to analyze the response in historical simulations of the Community Earth System Model's (CESM) Large Ensemble Simulations (LENS; Kay et al. 2015). LENS consists of 40 realizations of the same coupled climate model, that only differ in initial conditions consisting of random round off error perturbations in the initial temperature field. Sena and Magnusdottir (2020) evaluated the representation of the South American monsoon system (SAMS) in LENS compared to observations. They examined the seasonal precipitation anomaly and its spatial distribution, SAMS onset and retreat dates, and wet-season length in LENS and observations. They found that the ensemble mean of LENS captures the spatial distribution of the rainfall anomaly both during the peak of the wet season (DJF) and the peak of the dry season (JJA). This gives us confidence to examine LENS in terms of the South American response to the remote forcing of the IOD. The large number of realizations increases the sample size and allows us to compare the responses in LENS with the ones found in observations. Finally, we use the atmospheric component of LENS [the Community Atmospheric Model, version 5 (CAM5); Hurrell et al. 2013] to run forced experiments where we prescribe the dipole of sea surface temperature associated with both positive and negative IOD. These forced experiments allow us to isolate the response to the IOD SST pattern.

The paper is organized as follows. Section 2 describes the data and methods used. The results are described in section 3 and summarized and discussed in the conclusions (section 4).

2. Data, models, and methods

a. Observations

We examined the IOD and South American rainfall in observations using the following datasets. The Hadley Centre Sea Ice and Sea Surface Temperature (HadISST; Rayner et al. 2003) dataset is used to determine the conditions of the IOD and ENSO. The data are available monthly, from 1870 to the near present at a 1° spatial resolution. Monthly estimates of precipitation from the Global Precipitation Climatology Centre (GPCC; Schneider et al. 2014) are used to examine rainfall patterns in observations. GPCC is available globally over land from 1901 to the near present, at a 0.5° spatial resolution. In this paper, we use both observational datasets from 1901 to 2017.

The IOD state in observations was determined using the dipole mode index (DMI; Saji et al. 1999), defined by the difference between SST anomalies over the eastern Indian Ocean (10°S – 10°N , 50° – 70°E) and western Indian Ocean (10°S – 0° , 90° – 110°E). To compare the IOD state in observations and models, we consider IOD positive events when the DMI averaged over September–November is greater than its climatological mean plus one standard deviation. Similarly, IOD negative events are defined by years with DMI lower than the mean minus one standard deviation. We isolate years with ENSO neutral conditions by considering SST anomalies over the Niño-3.4 area (5°S – 5°N , 120° – 170°W) that are within one standard deviation from the mean. The SST pattern associated with IOD events depends on the dataset used in the analysis (Yang et al. 2020). HadISST was chosen due to the availability of a long continuous record and the similarities with other SST datasets in terms of magnitude and evolution of the DMI index (Fig. S1 in the online supplemental material) and SST patterns associated with the IOD (Fig. S2). We will discuss results in section 3.

It is important to point out that due to the association between IOD and ENSO there is only a small number of cases of IOD active during ENSO neutral conditions. This makes it difficult to attribute any response found in observations to IOD. To interpret the observational record, we also looked for a response in global climate models both fully coupled and in atmosphere only experiments as we describe in the following sections.

b. Large Ensemble simulations

LENS (Kay et al. 2015) is used to examine variability and to determine how the anomalies found in observations are reproduced in a fully coupled ocean–atmosphere global climate model (OAGCM). LENS consists of 40 realizations of the CESM1.1 model. Each ensemble member is run with historical radiative forcing from 1920 to 2005 and then follows the representative concentration pathway (RCP) 8.5 from 2006 to 2100. The ensemble members differ only in initial conditions of the surface temperature by a random roundoff error. LENS is available daily and monthly at a 1° spatial resolution. In this project we only analyze the historical period (1920–2005). In a previous study (Sena and Magnusdottir 2020) we found that LENS captures the evolution of the South American monsoon system including the seasonal distribution of rainfall anomalies, in spite of a mean rainfall bias in the model. The main circulation characteristics of the monsoon are captured in LENS. ENSO's seasonality is also well reproduced in CESM1 (Bellenger et al. 2014).

Pure IOD cases were isolated in each ensemble member. The ENSO state was determined based on SST over the Niño-3.4 area. Cases with the Niño-3.4 index within 0.6 standard deviation of the mean in each ensemble member were defined as ENSO neutral events. The IOD state was calculated based on the DMI index for the ENSO neutral conditions, averaged between September and November. Cases with DMI index greater than the mean plus one standard deviation were considered positive IOD events, while cases with the DMI index lower than the mean minus one standard deviation were

FIG. 1. SST forcing for each month for the (a) IODn and (b) IODp experiment.

considered negative IOD events. Note that due to the large number of cases in LENS, we chose a more restricted definition of neutral ENSO in LENS than in observations to further reduce the impact of ENSO in the final results. This threshold was chosen arbitrarily, as a way to maintain at least 250 cases for each phase of the IOD.

c. Forced CAM5 experiments

When we analyze observations and LENS, the responses we find may have been influenced by other perturbations concurrent with the IOD. Although by confining the analysis to ENSO neutral events reduces the effect of ENSO, some significant SST anomalies remain in the Pacific in LENS (Fig. S3), similar to what is found in the transitions between ENSO phases (Lin and Qian 2019). To isolate the IOD response, we performed forced atmospheric global climate model experiments using CAM5 to investigate the remote effects of the SST patterns associated with the IOD on the atmospheric circulation in the absence of other forcing. As CAM5 is the atmospheric component of LENS, we have a direct comparison with a fully coupled model, however, neglecting any response in surface turbulent flux.

A control experiment was performed, initialized with the annual cycle of the climatological SST from HadISST for the 1979–2008 period and run for 100 model years. We obtained the forcing field corresponding to the positive and negative IOD from the observed SST. Following Saji et al. (1999), we performed an empirical orthogonal function (EOF) analysis using the monthly (September, October, November) HadISST SST anomalies in the Indian Ocean (30°S–30°N, 30°–120°E) for the 1901–2017 period. We define IOD positive events as cases where the second EOF's reconstructed time series is greater than the mean plus one standard deviation. Similarly, the IOD negative events are cases where the reconstructed time series is less than the mean minus one standard deviation. In this way, we composited the SST anomalies over the Indian Ocean for

each month between April and December. The IOD positive (negative) experiment was forced by 2 times the composite of the SST anomalies for each month, as shown in Fig. 1. Hereafter, we will refer to these experiments as IODp (for positive IOD) and IODn (for negative IOD). One hundred realizations were performed for each experiment, with a 1° horizontal spatial resolution. Each experiment was run from 1 April to 31 December. Each ensemble member of the two perturbation experiments was initialized by branching from the control run on 1 April of each model year. That year of control was paired with the corresponding ensemble member to calculate the response.

The IOD develops in austral winter (Fig. 1), and rapidly decays after December. We chose to start our modeling experiment in April to provide 2 months of spinup time before the first signs of the SST dipole pattern starts to appear in June. In this study, we focus on the responses found from September to December.

d. Magnitude of the Walker circulation

The magnitude of the Walker circulation in each case was calculated based on the zonal component of the effective wind for water vapor transport, hereafter termed effective wind. The effective wind was determined following Sohn and Park (2010), and it is given by

$$\mathbf{V}_E = \sum_{i=p_0}^p W(i) \mathbf{V}_\chi(i), \quad (1)$$

where $\mathbf{V}_\chi$ is the divergent part of the horizontal wind field and W is the layerwise contribution of the water vapor amount to the total precipitable water in the atmospheric column, $p_0 = 1000$ hPa, and $p = 50$ hPa.

e. Rossby wave source

We quantified the Rossby wave source (RWS; Sardeshmukh and Hoskins 1988) in the extratropics following James (1994), defined as

$$S = -\nabla \cdot (\zeta \mathbf{V}_\chi) = -(\zeta \nabla \cdot \mathbf{V}_\chi + \mathbf{V}_\chi \cdot \nabla \zeta), \quad (2)$$

where $\mathbf{V}_\chi$ is the divergent component of the horizontal wind and ζ is the absolute vorticity. In this study, the RWS is calculated at 200 hPa.

3. Results

a. IOD representation in LENS

The seasonality of IOD is well reproduced in CESM (Fig. S1b), with the greatest variability occurring between September and November. The frequency of occurrence of IOD events is also well reproduced in LENS. The averaged period between two IODn years is 6.3 years in HadISST and 7.6 years in the ensemble mean of LENS for 1920–2005. For IODp, the mean period between two events is 7.9 years in observations and 8.4 years in the ensemble mean of LENS. In both observations and LENS, IODp events become more frequent at the end of the twentieth century. In the ensemble mean of LENS, there are seven IODp events and two IODn events during the 1979–2005 period. Cai et al. (2009) linked the increased frequency of IODp events to the strengthening of the easterly winds in the equatorial Indian Ocean due to a weakening of the Pacific Walker circulation and a stronger land–ocean temperature contrast in the Asian monsoon region.

The difference between the SST anomalies in the eastern and western Indian Ocean, measured by the DMI amplitude, is overestimated in LENS (Fig. S1a). CMIP5 climate models have a similar bias in the IOD amplitude, mainly due to an unrealistic representation of the thermocline slope, resulting in a thermocline–SST feedback (Cai and Cowan 2013; Ng et al. 2014; Yao et al. 2016). The current state-of-the-art CMIP6 models have similar overly strong biases (McKenna et al. 2020).

IOD is the first mode of variability for the September–November SST in LENS, corresponding to 31% of the explained variance. The spatial pattern of the IOD is reasonably well represented in LENS (Fig. S2) with a dipole between the SST in the eastern and western Indian Ocean.

Considering only neutral ENSO years, we found five years with positive IOD events (1961, 1963, 1967, 2012, 2017), 10 negative IOD events (1901, 1903, 1906, 1912, 1917, 1947, 1958, 1960, 1974, 1996), and 63 cases of IOD neutral in observations. During ENSO neutral, we found 290 model years with negative IOD, 260 model years with positive IOD, and 1094 model years with neutral IOD during ENSO neutral in the 40 ensemble members of LENS from 1920 and 2005.

b. General circulation anomalies in South America

In this section, we describe the circulation response over South America during IOD events in rainfall observations, in LENS, and in the forced experiments. In the first two cases, the response of each variable to IOD is defined as the difference between the composite of the pure IOD events in each phase and the composite of the cases with both IOD and ENSO neutral. In the forced experiments, the response is defined as the difference between the ensemble mean in each phase and the ensemble mean of the control experiment. Statistical significance of the response is determined by Student's *t* test at or

above the 95% confidence interval. Because of how few IODp and IODn events there are in the observational record the responses are not statistically significant.

Figure 2 shows the response to each IOD phase for the September–October average in terms of precipitation, horizontal wind at 850 hPa and zonal wind at 200 hPa in the forced experiments (Figs. 2a,b), in LENS (Figs. 2c,d), and in observations (only rainfall, Figs. 2e,f). In Fig. 3 we depict the same fields for the November–December average. During the positive IOD, there is widespread drying of the central and southeastern Brazil in observations in austral spring (Fig. 2f). These anomalies are also found both in LENS and the forced experiments (Figs. 2b,d) and are maintained through December (Fig. 3), suggesting a drying of the area where the SACZ is commonly observed during the austral summer. This response is consistent with the anomalies found by Chan et al. (2008).

During IODp events, the northern portions of the Amazon basin get wetter. In observations, this signal appears over northern Brazil during November and December (Fig. 3f). Although the results in observations are not statistically significant, a similar precipitation response in the Amazon appears earlier in LENS and lasts throughout the period (Figs. 2d and 3d). A positive rainfall anomaly begins to develop in September–October (Fig. 2d) and strengthens in November–December (Fig. 3d). The wetting appears in similar geographic locations in observations and LENS, especially for November–December. The forced experiments also feature a significant wetting over the Amazon in September–October (Fig. 2b), in a similar geographic location as the response found in LENS (Fig. 2d). However, this signal dissipates during November–December in the forced experiments (Fig. 3b). The difference in the seasonality of this signal may be due to differences between the models. LENS consists of an OAGCM, whereas our experiments are run in an atmospheric global climate model with prescribed SST and therefore the ocean feedback that is important in the tropics is not represented. It is also important to remember that although our method of selecting pure IOD cases in LENS and observations reduces the influence of ENSO in the final results, significant SST anomalies remain in the Pacific.

In the forced experiments, the precipitation response to IODp is associated with a weakening of the low-level winds directed from the Amazon to southeastern South America and, although the precipitation anomalies are weak, there is indication of increased convection over Amazon. These signals are examined closer in Fig. 4. Figure 4 shows the response of the horizontal wind, vertical velocity (ω) and specific humidity to IODp in September for a cross section between the Amazon and southeastern South America. There is anomalous upward motion over the Amazon in the forced experiment, while downstream areas in central and southeastern Brazil show more subsidence and drier conditions as the low-level winds weaken (Fig. 4). This dipole signal is also found during the early phases of the South American monsoon and is associated with the compensating subsidence from an increase in latent heat release over the Amazon (Gandu and Silva Dias 1998). Meanwhile, the South Atlantic subtropical high strengthens (Fig. 2b), increasing the moisture flux from the Atlantic Ocean to southeastern Brazil and contributing to the southeasterly anomalies in the low-level flux off the southeastern Brazil. During November and December,

FIG. 2. Response of horizontal wind at 850 hPa (vectors), zonal wind at 200 hPa (contours), and precipitation (color shading) for (a) IODn and (b) IODp, averaged over September and October in the forced experiments. The same anomalies are shown in LENS for (c) IODn and (d) IODp, compared to model years with both IOD and ENSO neutral. Precipitation anomalies in observations during years with (e) IODn and (f) IODp are shown. They are compared with 63 cases with both IOD and ENSO neutral. Stippling indicates areas where precipitation anomalies are significant above the 95% confidence level. Only significant anomalies of wind are shown. The extent of the Amazon basin is shown in magenta in (e).

the upper-level jet stream strengthens and narrows over South America and the Atlantic Ocean (Fig. 3b).

The precipitation response to IODn is weaker than the one found for IODp. In observations, it is characterized by increased rainfall over the northern part of the La Plata basin (around

15°S) especially during November–December and negative anomalies over southern Brazil and Uruguay (Fig. 3e). This response is also found in LENS, where it develops around September and is maintained through December (Figs. 2c and 3c). In the forced experiments, a similar wetting appears during

FIG. 3. As in Fig. 2, but averaged over November and December.

IODn (Fig. 2a), although the response is southward shifted compared to the anomalies in observations and LENS. In both LENS and our experiments there is a small, but significant strengthening of the northwesterly winds, increasing the moisture flux to the region where we found the positive precipitation response (Figs. 2a,c). In the following sections, we will explore the mechanisms that may generate these anomalies.

c. Extratropical wave train

Both IOD phases are associated with anomalous convection over the Indian Ocean. The anomalous tropical convection can

generate a Rossby wave train (e.g., Sardeshmukh and Hoskins 1988) that propagates from the subtropics across the Pacific and reaches South America. Figure 5 shows the November–December Rossby wave source response to IODn and to IODp over the Indian Ocean in the forced experiments. A significant response is found south of the imposed SST anomalies in both IODp and IODn.

The November–December response of the meridional wind at 200 hPa to IODn and IODp in the forced experiments is shown in Fig. 6. In both phases of the IOD, although more robust for IODp, a wave train arches across the South Pacific

FIG. 4. Vertical profile of vertical wind response (color shading), specific humidity anomalies (contours; in g kg^{-1}), and horizontal wind anomalies (barbs; in cm s^{-1}) for IODp in a cross section between the points 0° , 75.0°W and 28°S , 35°W , in September in the forced experiments. The insert map shows the path of the cross section in red. Areas where the vertical wind response is statistically significant at or above the 95% confidence level are hatched. Stippling represents areas where the specific humidity response is statistically significant at or above the 95% confidence level.

from the west to east, reaching South America (Figs. 6a,b). This signal is similar to the results found for September and October and in LENS (not shown). This wave train has also been linked to rainfall anomalies in southern Australia and temperature anomalies in Antarctica (McIntosh and Hendon 2018; Li et al. 2015).

In both IODn and IODp, the wave trains reach South America, disturbing the atmospheric circulation in the subtropics, for example, as shown by a strengthening of the South

Atlantic subtropical high (SASH) during the IODp, associated with an increased sea level pressure in the southern Atlantic (Fig. S4). The persistence of an anticyclonic circulation in the South Atlantic and southeast Brazil inhibits the northward propagation of low pressure systems and is associated with extreme drought events in southeast Brazil (Seth et al. 2015; Rodrigues et al. 2019). In addition, the onset of the South American monsoon is usually preceded by a weakening of the South Atlantic subtropical high (Raia and Cavalcanti 2008).

FIG. 5. Rossby wave source response to (a) IODn and (b) IODp, averaged over November and December in the forced experiments. Stippling represents responses that are statistically significant above the 95% confidence level.

FIG. 6. Meridional wind response at 200 hPa, averaged over November and December for (a) IODn and (b) IODp in the forced experiments. Stippling indicates areas where the anomalies are statistically significant above the 95% confidence level.

However, no significant response in the onset date of the South American monsoon was found in the forced experiments (Fig. S5).

Note that a Rossby wave train directed to the Northern Hemisphere is also excited during both phases of the IOD. The positive IOD has been associated with delayed circulation anomalies in the Asian summer monsoon (Yuan et al. 2008) and to the forcing of the North Atlantic Oscillation (Hardiman et al. 2020) in a case study of a recent extreme IOD event.

d. Tropical mechanism

The IOD can influence South America through a weakening of the local Walker circulation. During IODp, the magnitude of the effective wind [Eq. (1)] in the tropics over South America and the Atlantic Ocean (10°S–10°N, 80°–10°W) significantly weakens, on average, by about 6% in the forced experiments (Fig. S6).

While the Walker circulation weakens, ascending motion is strengthened over northwestern South America, favoring rainfall over the Amazon. Figure 7 shows the profile of vertical velocity response averaged between 5°S and 5°N during September–October. Even though the response is dwarfed by the local response close to the Indian Ocean, ascending motion over tropical South America (around 60°W) is significantly increased during IODp. This may be leading to the increase in rainfall over equatorial South America (Fig. 2).

We can better understand the effect of IODp on the Walker circulation by examining the evolution of the low-level seasonal wind response, using partially overlapping months. Figure 8 shows the equatorial 925 hPa zonal wind response, averaged between 10°S and 10°N, as a function of longitude in a Hovmöller diagram (for partially overlapping months) for IODp in the forced experiments. Consistent with a weakening of the Walker circulation, the trade winds weaken over the

FIG. 7. Response of the vertical component of the wind (ω) averaged between 5°S and 5°N for IODp in the forced experiments (shading) and streamlines of the climatological vertical and zonal wind for September and October. Only anomalies statistically significant above the 95% confidence level are shown.

FIG. 8. Response of the mean equatorial zonal wind averaged between 10°S and 10°N for IODp at 925 hPa is depicted with overlapping months on the y axis and longitude on the x axis, for the forced experiments. Hatched areas represent anomalies that are statistically significant at or above the 95% confidence level.

western portion of South America. The low-level wind response diverges in the eastern Indian Ocean, around 100°E, throughout the season. The trade winds weaken, starting around western Africa (0° longitude) in June–August and this signal spreads westward to the Atlantic Ocean up to 20°W in December. Meanwhile, the trade winds in the Pacific strengthen starting in August and this response spreads eastward with time.

This supports the hypothesis that the response of the Walker circulation to the IOD may be mediated by Kelvin waves, as suggested by Saji and Yamagata (2003). Figure 9 shows the space–time spectral analysis of the 850 hPa wind response to IODp. The strongest response is found in the Kelvin-wave frequencies. The westward propagation of the weakening of the trade winds may suggest that the increased latent heat released in the Indian Ocean excites a Gill–Matsuno-type response (Gill 1980; Matsuno 1966). Indeed, part of the tropical response is associated with a Rossby-wave response, but only in the longest wavelengths.

During IODp, the weakening of the low-level easterly winds is stronger in the Amazon region than in the adjacent areas (Figs. 8 and 2b). This, associated with the weakening of the flux from the Amazon to subtropical South America, results in increased moisture content in the Amazon region (Fig. 4). Meanwhile, the ascending branch of the Walker circulation is displaced to the east during IODp and is located anomalously in the central Amazon region (Fig. 7). As a combination of more moisture and ascending movement, convection is favored in the area.

Note that the IOD has a weak influence in the position of the ITCZ in the Atlantic Ocean. During IODn, the ITCZ is shifted northward from its climatological position in the forced experiments during September–October (Fig. 2a). No significant impact is found in LENS or in November–December. However, the response of the Atlantic ITCZ to IODp is different in LENS and

in the forced experiments. In the IODp experiments, there are dry anomalies in the center of the ITCZ in November–December, suggesting a drying of the area (Fig. 3b). Meanwhile, in LENS there is a significant wetting of the equatorial Atlantic during IODp in November–December (Fig. 3d), suggesting a strengthening of the ITCZ. The lack of an interactive ocean in the CAM5 experiments may account for these differences.

4. Concluding remarks

In this study, we analyzed how the Indian Ocean dipole (IOD) can influence the large-scale circulation over South America in observations, a large ensemble of simulations of an OAGCM and forced experiments where we directly force the SST anomalies corresponding to the positive and negative IOD in an atmospheric global climate model.

In observations of precipitation, the response to the positive phase of the IOD is characterized by drying over central and southeastern Brazil in SO and ND, accompanied by a wetting over the Amazon in ND. However, the observational record is short and since we want to isolate the effect of the IOD from the effects of ENSO, we only analyze ENSO neutral cases. This results in 5 IODp events and 10 IODn events in the observational record. The response in terms of rainfall is therefore not statistically significant. Nevertheless, this response is reproduced in both LENS, where we have 260 IODp cases and 290 IODn cases and thus statistical significance, as well as in forced experiments where we have an ensemble of 100 members of each sign of the IOD.

Two mechanisms contribute to the large-scale circulation anomalies found over South America associated with the positive IOD: a weakening of the tropical Walker circulation and an extratropical Rossby wave train that is forced by anomalous convection over the Indian Ocean.

FIG. 9. Symmetric component of the power spectra divided by the background power between 10°N and 10°S, based on the zonal wind response at 850 hPa for IODp.

During the positive phase of the IOD, a wave train is excited in the Indian Ocean that propagates across the Pacific to the South Atlantic. This anomaly strengthens the South Atlantic subtropical high, weakens the northwesterly low-level winds and reduces moisture transport from the Amazon to central and southeastern Brazil, generating dry anomalies in the areas where the SACZ is commonly observed. Meanwhile, the weakening of the Atlantic portion of the Walker circulation displaces its ascending branch to the east, increasing convection over the central portions of the Amazon basin. This results in a precipitation anomaly dipole between these two areas, similar to what is found during the early stages of the South American monsoon. This dipole is commonly associated with the onset date of the South American monsoon in observations (Raia and Cavalcanti 2008) and LENS (Sena and Magnusdottir 2020). However, we do not find any significant changes in the onset date of the South American monsoon in the forced experiments.

The signal found over the Amazon during IODp events is the opposite to the effect of El Niño, the most common phase of ENSO when IOD is in its positive phase. While El Niño is associated with drought over the Amazon, we see an anomalous wetting of the region during IODp. Future studies should address the effect of concurrent phases of positive ENSO and IOD on rainfall over the region.

The circulation anomalies in South America are weaker during the negative IOD. The precipitation response to IODn is characterized by a wetting over the northern portion of the La Plata basin, especially during November–December, in observations.

This response is reproduced in LENS, where it is observed as early as September and remains active until December. In the forced experiments (Fig. 2a), this response is shifted southward and is stronger during September–October and these anomalies are associated with a strengthening of the low-level winds from the Amazon to subtropical South America, resulting in an increase of moisture flux and a wetting downstream, specifically over southern Brazil. The absence of an active ocean in the forced experiments or external influences not due to the IOD in LENS may account for the difference in location of these responses.

Extreme events of IODp are projected to become more frequent by the end of the twenty-first century (Cai et al. 2013). It is therefore important to have a better understanding of how the IOD can affect global rainfall patterns. A better understanding can lead to increased predictive skill, which motivates studies on this subject.

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