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Madden-Julian Oscillation influence on sub-seasonal rainfall variability on the west of South America

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Abstract

The Madden-Julian Oscillation (MJO) is the leading driver of intraseasonal rainfall variability in the global tropics. However, the influence of MJO on western tropical South America (WTSA) has not been a focus of research. This is not surprising since the MJO convective core becomes nondescript as it propagates across the eastern Pacific, such that its influence on the Pacific coast of tropical South America is not obvious in global analyses. In this study, we examine MJO impacts on subseasonal rainfall variability in the rainiest season for WTSA (February–April). In order to avoid confusion with El Niño Southern Oscillation (ENSO) signals, only ENSO-neutral years are included in the analysis. We found that the MJO convective core reemerges when it propagates onto land in WTSA, and that it is associated with subseasonal precipitation anomalies of 20–50% relative to climatology. The MJO signal is evident in the Real-Time Multivariate MJO (RMM) index; however, the signal is clearer when a custom subseasonal index for the region based on WTSA outgoing longwave radiation is employed. Dynamically, the MJO influence is consistent with a modulation of the Pacific Ocean Walker Circulation descending branch, which is climatologically located in or near WTSA. Furthermore, MJO drives zonal and vertical motions on moisture and wind fields that influence precipitation in the region. We found that the timing of deep convection on subseasonal timescales captured by the regional index is consistent with a dominant role of the MJO convective core, rather than propagation of equatorial Rossby or Kelvin waves. However, there is evidence that equatorial Rossby waves that emerge over the tropical Atlantic Ocean also influence precipitation in WTSA on MJO timescales.

Keywords

Madden-Julian Oscillation; South America; intraseasonal variability; precipitation; western Amazon

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1. Introduction

Western Tropical South America (WTSA) is a notably complex climatic region. Bounded by the tropical Pacific Ocean to the west and the Amazon basin to the East, the area as defined for this paper includes portions of Colombia, territorial Ecuador and Peru (Figure 1). Topographically, this includes the lowland coastal regions that grade from arid in Peru to humid tropical in Colombia, tropical Andes Mountains, branching mountain chains and valleys of Colombia, and the eastern slopes of the mountains that descend into the Amazon lowlands. Climatically, conditions range from arid to humid tropical. Over a large part of the region, rainfall variability is high and can have significant impacts on ecosystems and the economy. Given the underlying geographic and climatic diversity, this variability has to be understood in the context of both global and local factors.

The dominant global-scale control on interannual climate variability in WTSA is the El Niño-Southern Oscillation (ENSO) (Bjerknes 1969). ENSO impacts on the region are well-documented and have been the subject of extensive research (Garreaud et al. 2009b; and references therein). Seasonally, the timing of rainy seasons in the region generally follows the migration of the Inter-Tropical Convergence Zone (ITCZ) over South America (Marengo et al. 2004; Garreaud 2009a) and fluctuations of coastal ocean surface currents (e.g., equatorial currents, Humboldt currents). In addition, the ascending Walker Circulation branch that sits in tropical South America modulates precipitation distribution over WTSA on both seasonal and intraseasonal timescales (Grimm and Ambrizzi 2009; and references therein).

The Madden-Julian oscillation (MJO; Madden and Julian 1971, 1972) is the most prominent mode of intraseasonal variability in the tropics (Madden and Julian 1994). For this reason, it is reasonable to expect that the MJO might influence precipitation on intraseasonal timescales in WTSA. MJO is a broad region of deep convection that originates over the Indian Ocean and then propagates eastward across the tropics, passing through the Maritime Continent, the western Pacific, South America, and finally over Africa. It takes approximately 30–60 days to complete a cycle around the globe (Madden and Julian 1994; Kiladis and Weickmann 1992) at about 5 m s^{-1} (Weickmann et al. 1985; Zhang 2005). The MJO involves variations in a variety of fields such as wind, sea surface temperature (SST), cloudiness, rainfall and Outgoing Longwave Radiation (OLR) (Zhang 2005).

Although the greatest variance in convection associated with MJO occurs over the Indo-Pacific warm pool (Madden and Julian 1994; Roundy 2014), its influence extends across the tropics and into the extra-tropics (Shimizu and Ambrizzi 2016; Jones et al. 2004, Roundy 2014; and for worldwide analysis, see Donald et al. 2006). For example, MJO impacts on precipitation events have been documented in Asia (Lawrence and Webster 2002), Australia (Hendon and Liebmann 1990; Wheeler et al. 2009), North America (Mo and Higgins 1998; Whitaker and Weickmann 2001; Zhou et al. 2012), and Africa (Matthews 2004, Berhane et al. 2015). In addition, MJO is known to modulate the genesis of tropical cyclones in the Pacific Ocean (Frank and Roundy 2006).

In South America, MJO modulates the South American Monsoon System (SAMS) (Nogués-Paegle et al. 2000; Liebmann et al. 2004; Carvalho et al. 2011), variations in which are related to anomalies in the position and intensity of the South Atlantic Convergence Zone (SACZ) (Nogués-Paegle and Mo 1997; Nogués-Paegle et al. 2000; Carvalho et al. 2004). These disturbances in the SACZ originate from the extratropical wave train that is the result of the interaction between the South Pacific Convergence Zone (SPCZ) and MJO (Grimm and Silva Dias 1995a, b; Ferranti et al. 1990; Grimm and Ambrizzi 2009). Additional variability of the SACZ can also be traced to the equatorward propagation of midlatitude Rossby wave train (Liebmann et al. 1999; Kiladis and Weickmann 1992; Carvalho et al. 2004; Carvalho et al. 2011; Vera et al. 2018; Boers et al., 2019). Furthermore, the MJO is the main atmospheric mechanism for intraseasonal modulation of precipitation over tropical Brazil (De Souza and Ambrizzi 2006), including the Amazon basin (Mayta et al. 2019; Espinoza et al. 2012; Paccini et al. 2018). Anomalies in the Walker cell over northern South America on MJO timescales drive considerable precipitation variability over the tropics, particularly over central-east Brazil, eastern and southeastern Amazon, and northeast Brazil (Grimm and Ambrizzi 2009; Ferranti et al. 1990; De Souza et al. 2005), suggesting that the passage of MJO over South America contributes to a considerable fraction of the total precipitation in the region (De Souza and Ambrizzi 2006), despite South America's considerable distance from the regions of strongest MJO convection in the Indian Ocean and Western Pacific Ocean.

These previous studies of MJO impacts on South America have elucidated mechanisms through which MJO affects different portions of the continent. Most of the studies, however, have focused on extratropical regions, and the remainder has examined the Amazon basin and Brazil. To our knowledge, no study has observed the influence of MJO on WTSA. As climate conditions and climate variability in this region are quite different from other portions of South America, this paper aims to quantify the MJO impact on precipitation variability during the rainiest season (February–April) in WTSA and to describe mechanisms through which MJO exerts influence on the region. We approach this problem with a suite of diagnostic tools applied to historical climate records and reanalysis data. Given WTSA's many challenges in agriculture, water management, health, energy, and disaster reduction (e.g., floods and droughts), increased knowledge of the physical and dynamical mechanisms associated with MJO impacts has the potential to contribute to the improvement of climate services on subseasonal timescales and longer.

2. Data and Methodology

2.1. Datasets

We draw on a range of reanalysis and satellite-derived products to characterize MJO and its impacts on WTSA.

OLR estimates are used as a proxy for deep convection and precipitation variability in tropical regions; thus, negative OLR anomalies indicate deep convection and positive OLR anomalies indicate clear skies (e.g., Nogués-Paegle et al. 2000). We used daily OLR values from the National Oceanic and Atmospheric Administration (NOAA) Climate Data Record (CDR), version 1.2 (Lee et al. 2011). The OLR data is estimated directly from the high-

resolution infrared radiation sounder (HIRS), has a resolution of 1-degree $\times$ 1-degree equal-angle gridded maps spanning from January 1, 1979 to date.

Daily precipitation estimates are drawn from the U.S. Geological Survey (USGS) Climate Hazards Group Infrared Precipitation with Station data (CHIRPS) merged gauge and satellite product (Funk et al. 2015). CHIRPS is a quasi-global rainfall dataset that spans from 50°S-50°N (and all longitudes) with a resolution of 0.05 degrees and is available from 1981 to date. The CHIRPS product is effective for investigating precipitation variability because of its high resolution, 37-year record, easy access, and overall good performance in the tropics (e.g., Dinku et al. 2018). In South America, validation studies have shown that CHIRPS estimates adequately reproduce several characteristics of precipitation when compared to ground stations and that it tends to outperform other high-resolution satellite-based datasets (e.g., Ceccherini et al. 2015; Paredes-Trejo et al. 2017; Rivera et al. 2018). Nevertheless, since CHIRPS relies heavily on satellite proxies for precipitation, it must be applied with caution.

For this reason, we used the CHIRPS dataset as our primary source to describe precipitation variability over WTSA. However, to compare the representation of CHIRPS precipitation in the region, we used two additional rainfall datasets. The first is the Tropical Rainfall Measuring Mission (TRMM) Multi-satellite Precipitation Analysis product (TMPA, Huffman et al. 2007). One of its products is the TRMM-3B42 (Huffman and Bolvin 2018) daily accumulated precipitation version 7 (V7) which is generated from the research-quality 3-hourly TMPA (3B42) and is available at 0.25-degree $\times$ 0.25-degree from 1998 to date. The second one is the observed data from the Hydro-geodynamics of the Amazon Basin Observatory (HYBAM, www.sohybam.org). The HYBAM daily observed precipitation (HOP) is a gridded dataset with 1-degree resolution from 1980 to 2009 for the Amazon basin derived from 752 meteorological stations that belong to the national hydro-meteorological institutions of five countries (Espinoza et al. 2009; Guimberteau et al. 2012; Espinoza et al. 2016). These datasets are chosen for comparison because they have been used widely in the region for different analyses, including studies of intraseasonal variation (e.g., Paccini et al. 2018, Mayta et al. 2019).

Atmospheric fields used in this study [e.g., geopotential height, wind vector data, vertical pressure velocity (omega) and humidity] come from the National Center for Environmental Prediction (NCEP)-National Center for Atmospheric Research (NCAR) Reanalysis-1 (R-1; Kalnay et al. 1996) dataset. The NCEP-NCAR dataset has a resolution of 2.5° at eight standard pressure levels (1000, 925, 850, 700, 600, 500, 400, and 300 hPa) and is available from 1948 to the present.

Daily Sea Surface Temperature (SST) data derives from the NOAA 1/4° daily Optimum Interpolation Sea Surface Temperature (OISST, Reynolds et al. 2007), which provides complete ocean temperature fields constructed by combining bias-adjusted observations from different platforms (satellite, ships, buoys) on a regular global grid, with gaps filled by interpolation. This dataset is available from 09/01/1981 to the present.

In addition, this study uses the Real-Time Multivariate MJO (RMM) index for MJO developed by Wheeler and Hendon (2004). The index describes the atmospheric variability directly related with MJO, and it is calculated as the principal component (PC) time series of the two leading empirical orthogonal functions (EOFs) of combined daily mean fields of OLR, 850 hPa zonal wind, and 200-hPa zonal wind data averaged over the tropics (15°S–15°N). The index is used to describe the temporal evolution of the amplitude and phase of the MJO and is partitioned into eight phases, where each phase corresponds to the geographical position of the active convective center of the MJO. The eight MJO phases are categorized as “strong” when the amplitude of the MJO is greater than 1 or “weak” when amplitude is less than 1, irrespective of the phase of the MJO (Wheeler and Hendon 2004; Zhang 2013). To ensure that our regional index is consistent with the RMM MJO index, we compared the analyses based on RMM to the analyses performed using the regionally customized MJO index (described below) developed in this study.

2.2. Data Analysis

For each dataset, we removed the ENSO cycle by limiting the analysis only to ENSO-neutral years. Neutral years, based on the NOAA Oceanic Niño Index (ONI, <https://catalog.data.gov/dataset/climate-prediction-center-cpc-oceanic-nino-index>), yielded a population of 18 ENSO-neutral years from the original period dataset (1981–2016). Then, from the neutral years, we computed the daily anomaly field for each dataset by subtracting the climatological day-of-year average from each day, on a gridded basis. In this way, we avoid convection signals in the dataset associated with both the seasonal cycle and ENSO phenomenon that can be confused with similar structures to specific phases of MJO (Lo and Hendon 2000; Wheeler and Hendon 2004). These daily anomalies fields for ENSO-neutral years are used to construct the regional index and composite schemes, as is described below.

While RMM and other global MJO indexing systems provide a useful framework for exploring global MJO impacts, these systems are not tuned to capture intraseasonal variability in specific target regions. For that reason, following the guidance of the Climate Variability and Predictability (CLIVAR) MJO Working Group (Waliser et al. 2009), we derived a regionally optimized index of intraseasonal variability based on EOF analysis. To extract the intra-seasonal time and space scales associated with MJO we filtered the daily OLR anomalies using a 20–100-day Lanczos bandpass filter (Duchon 1979) with 201 weights in the WTSA region (15°S–2.5°N, –82.5°W, –65°E). Then, the leading EOFs were calculated from the covariance matrix on these data for the February–April season (Figure 2). As seen in Figure 2, the leading EOFs present distinct spatial patterns associated with the variance explained by each EOF. The first EOF is well separated from the remaining EOFs, as confirmed by a North et al. (1982) test, and accounts for 39.9% of the variance. The second EOF explains 14.9% of the variance, and the third EOF accounts for 10.8% of the variance. The second and third EOFs will not be considered further because their expansion coefficient time series do not exhibit significant associations with that of the first EOF. Therefore, the first EOF is applied as an index of intraseasonal variability that we find captures the timing and character of the MJO impact on WTSA. From here on, we refer to this index as the Western Intraseasonal Index (WISI).

Next, intraseasonal variability at MJO timescales in WTSA is studied through composites of various fields using the RMM index and our regional WISI, based only on ENSO-neutral years. RMM composites were made by averaging the unfiltered daily anomaly fields on the dates when the RMM amplitude is greater than 1 in the phase of interest, a thresholding approach that is consistent with that used in previous RMM index studies (Wheeler and Hendon 2004). Meanwhile, we identified the positive and negative phases of MJO using the WISI when PC1 was higher or lower than one standard deviation (1σ) of the PC1, respectively. Results based on this restrictive definition have been commonly used when an EOF-based MJO index is used to describe the MJO (Zhang and Zhang 2018; Berhane et al. 2015). Thus, WISI composites for the different fields correspond to the average of unfiltered daily anomalies that fall within the dates of the positive or negative phase given by the WISI.

We used the non-parametric Mann-Whitney test, also called the Wilcoxon test (Mann and Whitney 1947; Wilcoxon 1945) to assess the significance of differences in composite analyses. The use of the non-parametric test is more appropriate when testing the significance of composite means since the data belonging to each group are not randomly selected but instead, have been selected based on the value of the index time series (Brown and Hall 1999). Consequently, the assumption of normality is also difficult to verify and inappropriate since the data distribution in each composite is unknown even if the original data distribution is assumed to be Gaussian (Terray et al. 2003). The Mann-Whitney U test overcomes the weaknesses of the normality assumption made by parametric tests (e.g., Dezfuli et al. 2015; Terray et al. 2003). The two-sided p-values for the composites are calculated using the test Mann-Whitney U-statistic:

$$U = R - \frac{n_2}{2}(n_2 + 1)$$

Here R is calculated as the sum of the ranks of the values, where first all the values from both groups are ranked, and then the ranks only from the first group are summed. And the term n_2 is the number of observations in the second group. We reject the null hypothesis when the p-value is less than 0.05, which is equivalent to the confidence level of 95% or higher.

Finally, the last part of section 3 is devoted to confirming that the convective signal captured by the regional WISI is associated mainly with the eastward propagation of the MJO convective core. As described by Wheeler and Kiladis (1999), the propagation characteristics of convectively-couple equatorial waves can be described with a wavenumber-frequency spectrum analysis of daily OLR anomalies. The spectrum analysis acts as a filter that retains the signal desired within the wavenumber-frequency domain applied to the daily OLR data, yielding a new OLR dataset in which the chosen mode of interest (Kelvin, Rossby, and MJO) is isolated (Wheeler and Kiladis, 1999). To retain the spectral peak associated with Kelvin waves, we used wavenumbers from 1 to 14 and frequencies from 1/30 cpd through to 1/2.5 cpd. Meanwhile, for Rossby waves ($n = 1$ ER) wavenumbers correspond from -10 to -1 and frequencies from 1/48 cpd through to 1/9.7 cpd. Kelvin and Rossby waves have the same equivalent depth ranges, from 8 to 90 m. Finally, for the propagating MJO convective core, wavenumbers and frequency domains

range from 1 to 5 and 1/96 cpd to 1/30 cpd, respectively (see Wheeler and Kiladis (1999) for details). The filter was applied to the daily OLR anomalies for all longitudes in the domain 20°S–20°N. Afterward, we regressed the WISI against the filtered OLR fields at a range of time lags to obtain sequences that capture each mode of influence. The statistical significance of regression coefficients for each lag composite is assessed through a two-tailed Student's *t*-test. The degrees of freedom for each grid was calculated based on the sample size for each filtered mode minus 2.

3. Results and Discussion

3.1 The impact of MJO on precipitation

This section presents the evaluation of MJO impacts on precipitation in WTSA using the RMM index and the regional WISI. In order to reduce the number of panels in the RMM index figures, phases 2 and 3, 4 and 5, 6 and 7, and 8 and 1 were merged, as each pair of merged phases showed similar connections with WTSA. Phases 8 and 1 are the period when MJO enhanced convection is found over the Western Hemisphere and Africa. Phases 2 and 3 correspond to when enhanced convection is over the Indian Ocean. Phases 4 and 5 occur when the convective center is located over the Maritime Continent. Finally, phases 6 and 7 correspond to when the convection core is over the western Pacific Ocean (Wheeler and Hendon 2004).

Figure 3 shows composites of daily OLR anomalies and daily precipitation anomalies for the merged phases 8+1, 2+3, 4+5 and 6+7 of MJO according to the RMM index for February–April. During phases 8+1 there is a coherent and statistically significant reduction in OLR in WTSA (Fig. 3a) and associated enhancement of precipitation (Fig. 3e). Meanwhile, during phases 4+5 suppressed convection (Fig. 3c) is associated with anomalous dryness in WTSA (Fig. 3g). These anomalies are strongest over southwest Colombia, Ecuador and northern of Peru (5°N to 15°S). There was no significant relation between precipitation and deep convection over most of WTSA during phases 2+3 and 6+7 (Fig. 3b and 3f; Fig. 3d and Fig. 3h, respectively), with the exception of some association in northern Colombia. The differences shown in Figure 3 are consistent with previous studies of MJO influence on other regions in South America, which have found precipitation increase in South America during phases 8+1, when the MJO convective core is located over Western Hemisphere and Africa, and precipitation reductions during phases 4+5 when the MJO convective core is centered over the Maritime Continent (e.g., Alvarez et al. 2016; Mayta et al. 2019; Juliá et al. 2012). TRMM-B42 composites (Fig. S3 in the supplementary material) show similar patterns to those obtained with CHIRPS precipitation product, supporting MJO rainfall impact analysis on WTSA.

Figure 4 shows the composites of daily unfiltered OLR anomalies and daily unfiltered precipitation anomalies based on WISI for February–April. Panels (a) and (b) show OLR anomalies in the context of the global tropics. They demonstrate that, when the regional WISI is greater than 1σ , suppressed convection (positive OLR anomalies) are found over Maritime Continent and Western Pacific (Fig. 4a), which is consistent with MJO phases 8 and 1 in the RMM system. Conversely, WISI less than 1σ is associated with strong convection in the Maritime Continent and Western Pacific (Fig. 4b), which is characteristic

of RMM MJO phases 4 and 5. These results provide confidence that WISI is capturing variability that is consistent with conventional MJO analysis and with known MJO convective features (Tian and Waliser, 2014). These convection patterns are associated with the expected precipitation impacts on WTSA (Fig. 4c, d; Fig. S4 in the supplemental material), with a pattern that is consistent with those found using RMM indices (Fig. 3) but with a slightly stronger and more extensive signal, reflecting the regionally customized nature of the index. For this reason, we use WISI for all further analyses.

The precipitation impact captured by WISI is both statistically significant (Fig. 4) and meteorologically meaningful in the context of background precipitation rates (Fig. 5). The percentage changes are computed as follows:

$$\Delta MJO\ phase_{p,n}(\%) = \left[\frac{\overline{FMA\ phase_{p,n}\ precip} - \overline{FMA\ precip}}{\overline{FMA\ precip}} \right] * 100$$

Here, $MJO\ phase_{p,n}(\%)$ is the percentage change of precipitation for a specific phase (positive or negative) due to MJO. $\overline{FMA\ phase_{p,n}\ precip}$ is the daily average precipitation anomaly calculated for the days that fall within each MJO phase given by the WISI, and $\overline{FMA\ precip}$ is the daily average precipitation anomaly for all days during the FMA season. Figure 5 shows positive excursions of WISI ($> +1\sigma$) are associated with an increase in precipitation on the order of 20% to 50% over southwest Colombia, Ecuador, and Peru. For the negative phase (when $WISI < -1\sigma$) there are precipitation decreases greater than 10% and ranging above 50% over much of the region. Again, this signal is not uniform, as over the northern Peru coastline there is a ~20% increase in precipitation, and this positive signal is seen on other portions of the coast as well. Percentage change of rainfall obtained using TRMM-3B42 and HOP dataset (Fig. S5 in the supplementary material) are consistent with CHIRPS; however, TRMM-3B42 shows a greater increase in precipitation to the east of the Andes and higher negative values in the Peruvian coast during an MJO positive phase.

3.2 MJO influence on atmospheric circulations

In this section, we investigate the MJO remote atmospheric connections that trigger precipitation changes in WTSA diagnosed using composite analysis. Composites are the average of unfiltered daily anomaly fields during ENSO-neutral years that fall on the dates for each phase defined by the regional WISI.

First, we consider vertical velocity anomalies at 500 hPa (Fig. 6). These show that the anomalous ascent (descent) in WTSA for $+1\sigma$ (-1σ) WISI conditions are associated with opposite-signed anomalies over the Maritime Continent. The positive phase is associated with an inverse overturning Walker-like circulation, with the core of the upward motion at the equatorial eastern Pacific and anomalous weak rising motion (positive pressure velocity anomaly) over the Maritime Continent (Fig. 6a). The weakening of the Pacific Ocean Walker Circulation is consistent with positive OLR anomaly observed in regions associated with subsidence (Fig. 4a). This reduction in convection over the Maritime Continent (Fig. 6a) affects Walker Circulation across the tropics, increasing convection and rainfall in WTSA (Fig. 4a, c). In the absolute vertical velocity values (not shown), there is still average rising

motion over the Maritime Continent during this phase, but it is weaker, while the descending branch of Walker cell that sits in WTSA is overcome by rising motion due to deep convection in the region as the result of the MJO convective core in tropical South America. Reverse patterns appear during the negative phase (Fig. 6b), where there is an anomalous intensification of the ascending Walker-like circulation in the western Pacific. The intensification that is related to stronger sinking motion over WTSA triggers negative precipitation anomalies in the region (Fig. 4d). Interestingly, during the positive phase, vertical velocity anomalies extend across much of tropical South America, including significant anomalous ascent over northeast Brazil, whereas in the negative phase anomalies are concentrated over WTSA.

These vertical velocity anomalies are consistent with the upper-level (250 hPa) winds anomalies shown in Figure 7. The positive phase of WISI shows easterly wind anomalies across the equatorial Pacific Ocean (Fig. 7a, vectors), accompanied by upper-tropospheric wind divergence over tropical South America and upper-level wind convergence over the Maritime Continent (Fig. 7a, contours), as would be expected for a weakened Pacific Ocean Walker Circulation. But there are also significant patterns in other portions of the tropics, including a wind convergence center in the eastern tropical Pacific Ocean at 100°W, near South America (Fig. 7a, contours). Moreover, the presence of upper-level westerlies on the west side and upper-level easterlies on the east side of equatorial South America implies zonal surface convergence in WTSA which is consistent with stronger negative OLR anomalies in the region observed during this phase (Fig. 4a). Additionally, the upper-level divergence in South America is accompanied by a pair of anticyclones to the west and a pair of cyclones to the east of the continent located in the subtropics, and the equatorial zonal wind anomalies are coupled to these subtropical wind systems (Fig. 6a). Rui and Wang (1990) noticed similar patterns in their composite analysis, and they associated these structures with Walker circulation and Gill's stationary solution which led them to conclude that these wind anomaly features are associated with a coupling of equatorial Kelvin and Rossby waves. In the negative phase, there is a similar proximal pattern, with a wind convergence anomaly over South America and anomalous wind divergence in the neighboring tropical Pacific Ocean.

These patterns are consistent with a model of anomalous zonal circulations forming in response to the MJO core, but they may also reflect interactions with other dynamical processes. Notwithstanding this complexity, the pattern of upper-level wind convergence and wind divergence over WTSA seen here is consistent with previous studies that show the SAMS (e.g., De Souza and Ambrizzi 2006; Nogués-Paegle et al. 2000) and the western Africa monsoon (e.g., Berhane et al. 2015; Lavender and Matthews 2009) are strengthened or weakened due to upper-level divergent anomalies circulation in its zonal (Walker Circulation) and meridional (Hadley Circulation) components (Grimm and Ambrizzi 2009).

Closer to the surface, circulation anomalies over WTSA associated with WISI are complicated by the extreme topography of the region. During the positive phase (Fig. 8a) the low-level easterly winds that prevail in FMA during ENSO-neutral years (see Fig. 1) are weakened. This weakening increases the probability of onshore transport of moist air from the neighboring ocean—which is also in a positive SST anomaly in this phase (Fig. 9a)—

and leads to convergence in equatorial WTSA along the Pacific Coast and along the Andes Mountains. This wind anomaly also intensifies the blocking of the predominant easterly moisture flux by diminishing the moisture that crosses over the Andes (Hoyos et al. 2017). Convergence is also seen south of the equator and east of the Andes Mountains in Peru and western Brazil, while divergence is observed to the north in the Panama Bight as well as over the equatorial Amazon. In the negative phase, there is divergence over the coastal equatorial region, associated with strengthened easterlies that reduce the intrusion of moist Pacific Ocean air into the region (Fig. 8b). There is also significant convergence in the equatorial Amazon in the negative phase, though this dynamic does not lead to significant rainfall anomalies in the Amazon.

Previous studies have shown that warmer SST temperatures coincide with the propagation of MJO core convection in the tropics (e.g., Zhang 2005). Anomalous clear skies and low precipitation prior to the arrival of the MJO convective core can enhance warming of surface waters, leading to a positive SST anomaly as the convective core arrives in the region. However, it is not clear how this mechanism would apply in WTSA, since there is little OLR signal associated with MJO's propagation across the eastern Pacific Ocean. Instead, it is possible that strengthened easterly winds during the negative phase of the WISI (Fig. 8b) encourage coastal upwelling and cool SST along the coast (Fig. 9b). During the positive phase the presence of onshore wind anomalies along the coast work against coastal upwelling, allowing warm SSTs to pool along the equator (Fig. 9a). The presence of warm SST during the positive phase and cool SST during the negative phase can, in turn, reinforce precipitation anomalies through ocean evaporation rates and near-surface humidity immediately off the coast.

These atmospheric circulation composites demonstrate that intraseasonal precipitation variability in WTSA is significantly attributable to MJO, and are consistent with a mechanism of divergence aloft, rising motion, and localized moisture convergence below. One can view the enhanced (suppressed) convection in the Maritime Continent and Western Pacific, and the suppressed (enhanced) convection over WTSA as an expression of MJO-induced modulation of the Walker Circulation (Grimm and Ambrizzi 2009). Thus, enhanced (inhibited) convection in WTSA corresponds to the ascending (descending) branch of an anomalous Walker Circulation tendency, as described by Madden and Julian (1994). At the same time, the MJO influence is mediated by its interaction with regional dynamics associated with topography, moisture transport from the Pacific Ocean and the Amazon basin, and coastal ocean upwelling and associated SSTs.

3.3 The role of Equatorial Waves

While the mechanisms described above focus on the propagation of the MJO convective core, previous work has suggested that equatorial Kelvin and Rossby waves interact in the propagation of MJO (e.g., Masunaga 2007; Straub and Kiladis 2003; Roundy and Frank 2004). To explore these interactions, we examine the temporal and spatial evolution of daily unfiltered OLR and the isolated signals of daily MJO-, Rossby- and Kelvin- filtered OLR regressed against the WISI at time lags from -15 days through +15 days, with an interval of 5 days, across the tropics.

Figure 10 shows the daily unfiltered OLR propagation associated with MJO development captured by the WISI index. The composite shows that during Lag -15 and Lag -10, negative OLR anomalies begin to develop in the western Pacific, just south of the equator, while a suppressed phase MJO signal emerges in the western Indian Ocean. During this period convection is suppressed in WTSA. From time Lag -5 to time Lag +5, atmospheric instabilities propagate eastward in the tropics. Propagation across the Pacific, however, is not clear in the OLR field; as is known, the MJO convective core becomes indistinct in this sector, so it is not possible to track a convectively coupled signal through to WTSA. By Lag 0, however, the convective signal does emerge over WTSA, as MJO stimulates upward motions over the region with a timing that is consistent with the arrival of the convective core. At the same time, suppressed convection is observed over the Maritime Continent and Western Pacific. Beyond Lag 0 the MJO continues its eastward propagation. Deep convection is observed over the Indo-West Pacific Ocean (from lag +10 to lag +15) and relatively clear skies are observed in WTSA. The sequence of linear regressions displayed in Fig. 10 is consistent with the results shown in Fig. 3 and Fig. 4, and it further confirms that WISI captures MJO-related variability.

Figure 11 highlights the MJO variance captured by the WISI as the MJO-filtered signal emerges in the western Pacific and propagates eastward from Lag -15 to Lag +15. The MJO-filtered OLR composites retain much of the signal (~50%) seen in Fig. 10, indicating that a considerable amount of deep convection, and thus changes in precipitation over WTSA are directly attributable to the migration of the MJO center of convection. Similar negative OLR anomalies patterns that extend from the western Amazon to the Atlantic have been related to SAMS intraseasonal modulation (Jones and Carvalho 2002) and SACZ intensification associated with MJO (De Souza and Ambrizzi 2006), offering additional mechanisms through which MJO can affect the region.

Meanwhile, the daily Rossby-filtered OLR westward signal shown in Fig. 12 is considerably weaker than the MJO-filtered signal (Fig. 11). Hence, Rossby waves are a minor factor in explaining the total intraseasonal variability (< 20%) captured by the WISI. Nevertheless, coherent Rossby structures are seen propagating westward through the time series. Most notably, a Rossby wave pair propagates from the tropical Atlantic across South America, reaching WTSA at Lag 0, such that the Rossby signal might reinforce the MJO signal in WTSA. Although additional studies are necessary to explore the convective wave-like patterns seen in Fig. 12, previous studies have suggested that MJO modulates the propagation of midlatitude Rossby wave trains (Weickmann and Berry 2007), and that these wave trains can perturb the equatorial waveguide in a manner that generates Rossby waves, which may then, in turn, influence the propagation characteristics of the MJO itself (Kiladis et al. 2009)

Kelvin waves are known to stimulate convection in tropical South America (Liebmann et al. 2009). However, a regression sequence of daily Kelvin-filtered OLR (not shown) demonstrated that propagating equatorial Kelvin waves did not significantly contribute to the convective intraseasonal variability captured by the WISI. This finding is consistent with the findings of Liebmann et al. (2009) and Straub and Kiladis (2003b), who reported that Kelvin Wave activity in South America is not strongly associated with the MJO.

Finally, when comparing the regression composites of daily MJO-, Rossby-, and Kelvin-filtered OLR summed signals for February-April season (Fig. 13) and the daily unfiltered OLR regression patterns (Fig. 10), we conclude that MJO wavenumber-frequency mode dominates the WISI signal, reinforcing the conclusion that MJO explains a good portion of intraseasonal variability in deep convection in WTSA. However, the slightly lower OLR magnitude in Figure 13 and the presence of coherent structures that resemble equatorial Rossby waves (Fig. 12), suggests that Rossby wave propagation from the tropical Atlantic, perhaps related to tropics-extratropical interactions (Kiladis and Weickmann 1992; Liebmann et al. 2004; Nogués-Paegle et al 2000; Carvalho et al. 2004; Kiladis et al. 2009), also contributes to MJO-scale deep convection variability over the region.

4. Conclusions

Western tropical South America (WTSA) is a geographically diverse and climatically complex region, in which rainy season precipitation variability can have critical impacts on ecosystem productivity, agriculture, human health, and water resources. Few studies have examined drivers of subseasonal climate variability in this region. Here we have shown a significant link between WTSA rainy season variability and the MJO, as the MJO core propagates into the region and emerges as a center of convective activity. While WTSA is remote from the regions of strongest MJO activity in the Indian Ocean and Western Pacific, the presence of MJO influence at this distance is consistent with previous studies that have found pan-tropical MJO influence (Berhane et al. 2015; Matthews 2004; Donald et al. 2006; Frank and Roundy 2006). This includes analyses that have shown MJO influence in Amazon Basin and Brazil (e.g., Carvalho et al. 2002; De Souza and Ambrizzi 2006; Paccini et al. 2018; Ferranti et al. 1990; Mayta et al. 2019). Other studies have also shown MJO influence in extratropical South America (Grim and Silva 1995a, b; Grimm and Ambrizzi 2009; Alvarez et al. 2016).

Similar to studies of the MJO influence on the Amazon Basin and Brazil, our analysis shows that MJO is the main driver of intraseasonal precipitation anomalies on the region. The mechanism of MJO influence, however, appears to be somewhat different for WTSA, on account of its geographic context. Studies of the Amazon Basin and Brazil have emphasized MJO-associated modulations in the SAMS and SACZ regions related to extra-tropical train wave propagation, along with direct influence of the passing MJO convective core. For WTSA, which is located on the edge of the descending branch of the Pacific Ocean Walker cell and includes both the Pacific coast and the Andes Mountains, we find that MJO convective core modulation of regional circulation patterns can explain much of the observed influence on precipitation. Namely, the eastward propagation of the MJO convective core into the region counteracts prevailing descent associated with the Walker circulation, while the propagation of the suppressed phase of MJO reinforces Walker circulation subsidence. This result is consistent with previous studies that have reported MJO influence on the Walker circulation (Madden and Julian 1994; Slawinska et al. 2014). Wind patterns, meanwhile, that are associated with the reemergence of MJO convection over WTSA, weaken the prevailing near-surface easterly winds, and, thereby, enhance moisture vapor convergence.

Another noteworthy finding of this study is the relation between coastal equatorial SST anomalies and MJO propagation. The SST anomalies are consistent with the observed precipitation anomalies, in that warm SSTs are found during MJO enhancement of precipitation, suggesting that warmer conditions and enhanced evaporation might contribute to the total MJO influence on the region. This relationship requires further investigation, ideally with coupled models capable of capturing SST-convection feedbacks in the eastern Pacific (Maloney and Kiehl, 2002).

We have focused on diagnosing MJO impacts on WTSA and on elucidating the dominant mechanisms of MJO influence on the region. Future work could consider secondary pathways of MJO influence, including via extratropical to tropical dynamics (Ferranti et al. 1990; Nogués-Paegle and Mo 1997; Liebmann et al. 2004), air-sea interactions, interactions with ENSO, and modulation of monsoon circulations.

Recognition of the MJO influence on WTSA also opens opportunities in subseasonal precipitation forecasts. MJO can serve as a predictor in statistical precipitation forecasts, and it can also be integrated into the evaluation of dynamically-based forecast systems. Our results indicate that a model's ability to simulate MJO propagation across the eastern Pacific and reemergence over South America could be critical for forecasting precipitation variability on extended weather to subseasonal timescales.

Supplementary Material

Refer to Web version on PubMed Central for supplementary material.

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References

- Alvarez MS, Vera CS, Kiladis GN, Liebmann B (2016) Influence of the Madden–Julian oscillation on precipitation and surface air temperature in South America. *Clim Dyn* 46:245–262. 10.1007/s00382-015-2581-6
- Berhane F, Zaitchik B, Badr HS (2015) The Madden–Julian Oscillation's influence on spring rainy season precipitation over equatorial west Africa. *J Clim* 28:8653–8672. 10.1175/JCLI-D-14-00510.1
- Boers N, Goswami B, Rheinwalt A, Bookhagen B, Hoskins B, and Kurth J (2019) Complex networks reveal global pattern of extreme-rainfall teleconnections. *Nature*, 566:373–377, 10.1038/s41586-018-0872-x [PubMed: 30700912]
- Bjerknes J (1969) Atmospheric teleconnections from the Equatorial Pacific. *Mon Weather Rev* 97:163–172. 10.1175/1520-0493(1969)097<0163:ATFTEP>2.3.CO;2
- Brown TJ, Hall BL (1999) The use of t values in composite analyses. *J Clim* 12: 2941–2944. 10.1175/1520-0442(1999)012<2941:TUOTVI>2.0.CO;2
- Carvalho LMV, Jones C, Liebmann B (2002) Extreme precipitation events in Southeastern South America and large-scale convective patterns in the South Atlantic Convergence Zone. *J Clim* 15:2377–2394. 10.1175/1520-0442(2002)015<2377:EPEISS>2.0.CO;2

- Carvalho LMV, Jones C, Liebmann B (2004) The South Atlantic Convergence Zone: intensity, form, persistence and relationships with intraseasonal to interannual activity and extreme rainfall. *J Clim* 17:88–108. 10.1175/1520-0442(2004)017<0088:TSACZI>2.0.CO;2
- Carvalho LMV, Silva AE, Jones C, Liebmann B, Silva Dias PL, Rocha HR (2011) Moisture transport and intraseasonal variability in the South America monsoon system. *Climate Dyn* 36:1865–1880. 10.1007/s00382-010-0806-2
- Ceccherini G, Amezttoy I, Hernández CPR, Moreno CC (2015) High-resolution precipitation datasets in South America and West Africa based on satellite-derived rainfall, enhanced vegetation index and digital elevation model. *Remote Sens* 7:6454–6488. 10.3390/rs70506454
- De Souza EB, Kayano MT, Ambrizzi T (2005) Intraseasonal and submonthly variability over the eastern Amazon and Northeast Brazil during the autumn rainy season. *Theor Appl Climatol* 81:177–191. 10.1007/s00704-004-0081-4
- De Souza EB, Ambrizzi T (2006) Modulation of the intraseasonal rainfall over tropical Brazil by the Madden Julian oscillation. *Int J Climatol* 26:1759–1776. 10.1002/joc.1331
- Dezfuli AK, Zaitchik BF, Gnanadesikan A (2015) Regional atmospheric circulation and rainfall variability in south equatorial Africa. *J Clim* 28:809–818. 10.1175/JCLI-D-14-00333.1
- Dinku T, Funk C, Peterson P, Maidment R, Tadesse T, Gadain H, Ceccato P (2018) Validation of the CHIRPS satellite rainfall estimates over Eastern of Africa. *Q J R Meteorol Soc*:1–21. 10.1002/qj.3244
- Donald A, Meinke H, Power B, Maia AHN, Wheeler MC, White N, Stone RC, Ribbe J (2006) Near-global impact of the Madden-Julian oscillation on rainfall. *Geophy Res Lett* 33:L09704. 10.1029/2005GL025155
- Duchon CE (1979) Lanczos Filtering in One and Two Dimensions. *J Appl Meteorol* 18:1016–1022. 10.1175/1520-0450(1979)018<1016:LFOAT>2.0.CO;2
- Espinoza JC, Ronchail J, Guyot JL, Cochonneau G, Filizola N, Lavado W, De Oliveira E, Pombosa R, Vauchel P (2009) Spatio-temporal rainfall variability in the Amazon basin countries (Brazil, Peru, Bolivia, Colombia, and Ecuador). *Int J Climatol* 29:1574–1594. 10.1002/joc.1791
- Espinoza JC, Lengaigne M, Ronchail J, Janicot S (2012) Large-scale circulation patterns and related rainfall in the Amazon Basin: a neuronal networks approach. *Clim Dyn*. 10.1007/s00382-011-1010-8
- Espinoza JC, Segura H, Ronchail J, Drapeau G, Gutierrez-Cori O (2016) Evolution of wet- and dry-day frequency in the western Amazon basin: relationship with atmospheric circulation and impacts on vegetation. *Water Resour Res* 52:8546–8560. 10.1002/2016WR019305
- Ferranti L, Palmer TN, Molteni F, Klinker E (1990) Tropical-extratropical interaction associated with the 30–60 day oscillation and its impact on medium and extended range prediction. *J Atmos Sci* 47:2177–2199. 10.1175/1520-0469(1990)047<2177:TEIAWT>2.0.CO;2
- Frank WM, Roundy PE (2006) The role of tropical waves in tropical cyclogenesis in the western North Pacific. *Mon Weather Rev* 134:2397–2417. 10.1175/MWR3204.1
- Funk C, Peterson P, Landsfeld M et al. (2015) The climate hazards infrared precipitation with stations —a new environmental record for monitoring extremes. *Sci Data* 2:150066. 10.1038/sdata.2015.66 [PubMed: 26646728]
- Garreaud RD (2009a) The Andes climate and weather. *Adv. Geosci* 22:3–1. 10.5194/adgeo-22-3-2009
- Garreaud RD, Vuille M, Compagnucci R, Marengo J (2009b) Present-day South American climate. *Palaeogeogr Palaeoclimatol Palaeoecol* 281:180–195. 10.1016/j.palaeo.2007.10.032
- Grimm AM, Silva Dias PL (1995a) Use of barotropic models in the study of the extratropical response to tropical heat sources. *J Meteorol Soc Jpn* 73:765–779. 10.2151/jmsj1965.73.4_765
- Grimm AM, Silva Dias PL (1995b) Analysis of tropical-extratropical interactions with influence functions of a barotropic model. *J Atmos Sci* 52:3538–3555. 10.1175/1520-0469(1995)052<3538:AOTIWI>2.0.CO;2
- Grimm AM, Ambrizzi T (2009) Teleconnections into South America from the tropics and extratropics on interannual and intraseasonal timescales In: Vimeux F, Sylvestre F, Khodri M (eds) *Past climate variability in South America and surrounding regions*, 1st edn Springer, Netherlands, pp 159–191. 10.1007/978-90-481-2672-9_7

- Guimberteau M, Drapeau G, Ronchail J, Sultan B, Polcher J, Martinez JM, Prigent C, Guyot JL, Cochonneau G, Espinoza JC, Filizola N, Fraizy P, Lavado W, De Oliveira E, Pombosa R, Noriega L, Vauchel P (2012) Discharge simulation in the sub-basins of the Amazon using ORCHIDEE forced by new datasets. *Hydrol Earth Syst Sci* 16:911–935. 10.5194/hess-16-911-2012
- Hendon HH, Liebmann B (1990) The intraseasonal (30–50 day) oscillation of the Australian summer monsoon. *J Atmos Sci* 47:2909–2923. 10.1175/1520-0469(1990)047<2909:TIDOOT>2.0.CO;2
- Hoyos I, Dominguez F, Cañón-Barriga J, Martínez JA, Nieto R, Gimeno L, Dirmeyer PA (2017) Moisture origin and transport processes in colombia, northern south america. *Clim Dyn*. 10.1007/s00382-017-3653-6
- Huffman GJ, Bolvin DT, Nelkin EJ, Wolff DB, Adler RF, Gu G, Hong Y, Bowman KP, Stocker EF (2007) The TRMM multisatellite precipitation analysis (TMPA): quasi-global, multiyear, combined-sensor precipitation estimates at fine scales. *J Hydrometeorol* 8(1): 38–55.
- Huffman GJ, Bolvin DT (2018) TRMM and Other Data Precipitation Data Set Documentation. NASA: Greenbelt, USA, 1–40. 10.5067/TRMM/TMPA/DAY/7
- Jones C, Carvalho LMV (2002) Active and break phases in the South American monsoon system. *J Clim* 15:905–914. 10.1175/1520-0442(2002)015<0905:AABPIT>2.0.CO;2
- Jones C, Waliser DE, Lau KM, Stern W (2004) Global occurrences of extreme precipitation and the Madden–Julian oscillation: observations and predictability. *J Clim* 17:4575–4589. 10.1175/3238.1
- Juliá C, Rahn DA, Rutllant JA (2012) Assessing the influence of the MJO on strong precipitation events in subtropical, semi-arid north-central Chile (30°S). *J Climate* 25:7003–7013. 10.1175/JCLI-D-11-00679.1
- Kalnay E, Kanamitsu M, Kistler R, Collins W, Deaven D, Gandin L, Iredell M, Saha S, White G, Woollen J, Zhu Y, Chelliah M, Ebisuzaki W, Higgins W, Janowiak J, Mo KC, Ropelewski C, Wang J, Leetmaa A, Reynolds R, Jenne R, Joseph D (1996) The NCEP/NCAR 40-year reanalysis project. *Bull Am Meteorol Soc* 77:437–471. 10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2
- Kiladis GN, Weickmann KM (1992) Circulation anomalies associated with tropical convection during northern winter. *Mon Weather Rev* 120:1900–1923. 10.1175/1520-0493(1992)120<1900:CAAWTC>2.0.CO;2
- Kiladis GN, Wheeler MC, Haertel PT, Straub KH, Roundy PE (2009) Convectively coupled equatorial waves. *Rev Geophys* 47: RG2003 10.1029/2008RG000266
- Lavender SL, Matthews AJ (2009) Response of the West African monsoon to the Madden–Julian oscillation. *J Clim* 22:4097–4116. 10.1175/2009JCLI2773.1
- Lawrence DM, Webster PJ (2002) The boreal summer intraseasonal oscillation: relationship between northward and eastward movement of convection. *J Atmos Sci* 59:1593–1606. 10.1175/1520-0469(2002)059<1593:TBSIOR>2.0.CO;2
- Lee HT and NOAA CDR Program, 2011: NOAA Climate Data Record (CDR) of Daily Outgoing Longwave Radiation (OLR), Version 1.2. NOAA National Climatic Data Center. <https://www.ncei.noaa.gov/data/outgoing-longwave-radiation-daily/access/>. Accessed 15 April 2019.
- Liebmann B, Kiladis GN, Marengo JA, Ambrizzi T, Glick JD (1999) Submonthly convective variability over South America and the South Atlantic convergence zone. *J Clim* 12:1877–1891. 10.1175/1520-0442(1999)012<1877:SCVOSA>2.0.CO;2
- Liebmann B, Vera CS, Carvalho LMV, Camilloni IA, Hoerling MP, Allured D, Barros VR, Báez J, Bidegain M (2004) An observed trend in Central South American precipitation. *J Clim* 17:4357–4367. 10.1175/3205.1
- Liebmann B, Kiladis GN, Carvalho LM, Jones C, Vera CS, Bladé I, Allured D (2009) Origin of convectively coupled Kelvin waves over South America. *J Clim* 22:300–315. 10.1175/2008JCLI2340.1
- Lo F, Hendon HH (2000) Empirical extended-range prediction of the Madden-Julian oscillation. *Mon Weather Rev* 128: 2528–2543. 10.1175/1520-0493(2000)128<2528:EERPOT>2.0.CO;2
- Madden RA, Julian PR (1971) Detection of a 40–50 day oscillation in the zonal wind in the tropical Pacific. *J Atmos Sci* 28:702–708. 10.1175/1520-0469(1971)028<0702:DOADOI>2.0.CO;2
- Madden RA, Julian PR (1972) Description of global-scale circulation cells in the tropics with a 40–50 day period. *J Atmos Sci* 29:1109–1123. 10.1175/1520-0469(1972)029<1109:DOGSCC>2.0.CO;2

- Madden RA, Julian PR (1994) Observations of the 40–50-day tropical oscillation—a review. *Mon Weath Rev* 122:814–837. 10.1175/1520-0493(1994)122<0814:OOTDIO>2.0.CO;2
- Maloney ED, Kiehl JT (2002) MJO related SST variations over the tropical eastern Pacific during northern Hemisphere summer. *J Clim* 15:675–689. 10.1175/1520-0442(2002)015<0675:MRSVOT>2.0.CO;2
- Mann HB, Whitney DR (1947) On a test of whether one of two random variables is stochastically larger than the other. *Ann. Math. Stat* 18:50–60. Retrieved from <http://www.jstor.org/stable/2236101>
- Marengo JA (2004) Interdecadal variability and trends of rainfall across the Amazon basin. *Theor Appl Climatol* 78:79–96. 10.1007/s00704-004-0045-8
- Masunaga H (2007) Seasonality and regionality of the Madden-Julian oscillation, Kelvin wave, and equatorial Rossby wave. *J Atmos Sci* 64:4400–4416. 10.1175/2007JAS2179.1
- Matthews AJ (2004) Intraseasonal variability over tropical Africa during northern summer. *J Clim* 17:2427–2440. 10.1175/1520-0442(2004)017<2427:IVOTAD>2.0.CO;2
- Mayta VC, Ambrizzi T, Espinoza JC, Silva Dias PL (2019) The role of the Madden-Julian oscillation on the Amazon Basin intraseasonal rainfall variability. *Int J Climatol* 39: 343–360. 10.1002/joc.5810
- Mo KC, Higgins RW (1998) Tropical influences on California precipitation. *J Clim* 11:412–430. 10.1175/1520-0442(1998)011<0412:TIOCP>2.0.CO;2
- Nogués-Paegle J, Mo KC (1997) Alternating wet and dry conditions over South America during summer. *Mon Weather Rev* 125: 279–291. 10.1175/1520-0493(1997)125<0279:AWADCO>2.0.CO;2
- Nogués-Paegle J, Byerle LA, Mo KC (2000) Intraseasonal modulation of South American summer precipitation. *Mon Weather Rev* 128: 837–850. 10.1175/1520-0493(2000)128<0837:IMOSAS>2.0.CO;2
- North GR, Bell TL, Cahalan RF, Moeng FJ (1982) Sampling errors in the estimation of empirical orthogonal functions. *Mon Weather Rev* 110: 699–706. 10.1175/1520-0493(1982)110<0699:SEITEO>2.0.CO;2
- Paccini L, Espinoza JC, Ronchail J, Segura H (2018) Intra-seasonal rainfall variability in the Amazon basin related to large-scale circulation patterns: a focus on western Amazon-Andes transition region. *Int J Climatol* 38, 2386–2399. 10.1002/joc.5341
- Paredes-Trejo FJ, Barbosa HA, Kumar TVL (2017) Validating CHIRPS-based satellite precipitation estimates in Northeast Brazil. *J Arid Environ* 139:26–40. 10.1016/j.jaridenv.2016.12.009
- Reynolds RW, Smith TM, Liu C, Chelton DB, Casey KS, Schlax MG (2007) Daily high-resolution blended analyses for sea surface temperature. *J Clim* 20:5473–5496. 10.1175/2007JCLI1824.1
- Rivera JA, Marianetti G, Hinrichs S (2018) Validation of CHIRPS precipitation dataset along the Central Andes of Argentina. *Atmos Res* 213:437–449. 10.1016/j.atmosres.2018.06.023
- Roundy PE, Frank WM (2004) Effects of low-frequency wave interactions on intraseasonal oscillations. *J Atmos Sci* 61:3025–3040. 10.1175/JAS-3348.1
- Roundy PE (2014) Regression analysis of zonally narrow components of the MJO. *J Atmos Sci* 71:4253–4275. 10.1175/JAS-D-13-0288.1
- Rui H, Wang B (1990) Development characteristics and dynamic structure of tropical intraseasonal convection anomalies. *J Atmos Sci* 47: 357–379. 10.1175/1520-0469(1990)047<0357:DCADSO>2.0.CO;2
- Shimizu MH, Ambrizzi T (2016) MJO influence on ENSO effects in precipitation and temperature over South America. *Theor Appl Climatol* 124:291–301. 10.1007/s00704-015-1421-2
- Slawinska J, Pauluis O, Majda A, Grabowski W (2014) Multiscale interactions in an idealized walker circulation: mean circulation and intraseasonal variability. *J Atmos Sci*. 10.1175/JAS-D-13-018.1
- Straub KH, Kiladis GN (2003) Interactions between the boreal summer intraseasonal oscillation and higher-frequency tropical wave activity. *Mon Weather Rev* 131:945–960. 10.1175/1520-0493(2003)131<0945:IBTBSI>2.0.CO;2
- Straub KH, Kiladis GN (2003b) The observed structure of convectively coupled Kelvin waves: Comparison with simple models of coupled wave instability. *J Atmos Sci* 60:1655–1668. 10.1175/1520-0469(2003)060<1655:TOSOCC>2.0.CO;2

- Terray P, Delecluse P, Labattu S, Terray L (2003) Sea surface temperature associations with the late Indian summer monsoon. *Clim Dyn* 21:593–618. 10.1007/s00382-003-0354-0
- Tian B, Waliser D (2014) Madden-Julian Oscillation (MJO). In: Njoku EG (eds) *Encyclopedia of Remote Sensing Encyclopedia of Earth Sciences Series*. Springer, New York, NY, pp. 349–358. 10.1007/978-0-387-36699-9_198
- Vera CS, Alvarez MS, Gonzalez PLM, Liebmann B, Kiladis GN (2018) Seasonal cycle of precipitation variability in South America on intraseasonal timescales. *Clim Dyn* 51: 1991 10.1007/s00382-017-3994-1
- Waliser D, Sperber K, Hendon H, Kim D, Maloney E, Wheeler M, Weickmann K, Zhang C, Donner L, Gottschalck J, Higgins W, Kang IS, Legler D, Moncrieff M, Schubert S, Stern W, Vitart F, Wang B, Wang W, Woolnough S (2009) MJO simulation diagnostics. *J Clim* 22:3006–3030. 10.1175/2008JCLI2731.1
- Weickmann KM, Lussky GL, Kutzbach JE (1985) Intraseasonal (30–60 day) fluctuations of outgoing longwave radiation and 250 mb streamfunction during northern winter. *Mon Weather Rev* 113: 941–961. 10.1175/1520-0493(1985)113<0941:IDFOOL>2.0.CO;2
- Weickmann KM, Berry E (2007) A synoptic-dynamic model of subseasonal atmospheric variability. *Mon Weather Rev* 135:449–474. doi:10.1175/MWR3293.1
- Wheeler M, Kiladis GN (1999) Convectively coupled equatorial waves: analysis of clouds and temperature in the wavenumber-frequency domain. *J Atmos Sci* 56:374–399. 10.1175/1520-0469(1999)056<0374:CCEWAO>2.0.CO;2
- Wheeler MC, Hendon HH (2004) An all-season real-time multivariate MJO index: development of an index for monitoring and prediction. *Mon Weather Rev* 132:1917–1932. 10.1175/1520-0493(2004)132<1917:AARMMI>2.0.CO;2
- Wheeler MC, Hendon HH, Cleland S, Meinke H, Donald A (2009) Impacts of the Madden-Julian oscillation on Australian rainfall and circulation. *J Clim* 22:1482–1498. 10.1175/2008JCLI2595.1
- Whitaker JS, Weickmann KM (2001) Subseasonal variations of tropical convection and week-2 prediction of wintertime western North American rainfall. *J. Climate*, 14, 3279–3288. 10.1175/1520-0442(2001)014<3279:SVOTCA>2.0.CO;2
- Wilcoxon F (1945) Individual comparisons by ranking methods. *Biometrics* 1:80–83. 10.2307/3001968
- Zhang C (2005) Madden–Julian oscillation. *Rev Geophys* 43:RG2003 10.1029/2004RG000158
- Zhang C (2013) Madden–Julian oscillation: bridging weather and climate. *Bull Am Meteorol Soc* 94:1849–1870. 10.1175/BAMS-D-12-00026.1
- Zhang C, Zhang B (2018) QBO–MJO connection. *J Geophys Res Atmos* 123: 2957–2967. 10.1002/2017JD028171
- Zhou S, L’Heureux M, Weaver S, Kumar A (2012) A composite study of the MJO influence on the surface air temperature and precipitation over the continental United States. *Clim Dyn* 38:1459–1471. 10.1007/s00382-011-1001-9

Fig. 1. February-April (FMA) average precipitation (mm day^{-1}) from the CHIRPS product, and patterns of 850 hPa horizontal winds (m s^{-1}) from the NCEP/NCAR Reanalysis for ENSO-neutral years from 1981 to 2016. The red box indicates the area of study (WTSA)

Fig. 2.

Spatial structure of the three leading EOF modes of the 20–100-day bandpass filtered OLR data over West of South America (-15°S , 2.5°N , -82.5°W , -65°E) for FMA during ENSO-neutral years from 1981 to 2016

Fig. 3.

FMA composites of daily unfiltered OLR anomalies (upper-panel, W m^{-2}) and daily unfiltered rainfall anomalies (bottom-panel, mm day^{-1}) for the merged phases of MJO according to RMM index during ENSO-neutral years from 1981 to 2016. The phases merged to obtain each composite are indicated in the bottom left of each panel. Only significant anomalies at 95% confidence level are shown. Contour interval for OLR is 2 W m^{-2} and contour interval for rainfall is 1 mm day^{-1} , blue (red) shadings areas indicate positive (negative) rainfall anomalies

Fig. 4.

FMA composites for: (a), (b) daily unfiltered OLR anomalies (W m^{-2}) and (c), (d) daily unfiltered rainfall anomalies (mm day^{-1}) during ENSO-neutral years from 1981 to 2016 based on WISI for (a), (d) $\text{WISI} > +1\sigma$ and (b), (d) $\text{WISI} < -1\sigma$. Only results significant anomalies at 95% confidence level are shown. Contour interval for OLR is 2 W m^{-2} . Rainfall contour interval is 1 mm day^{-1} , blue (red) shadings areas indicate positive (negative) rainfall anomalies

Fig. 5.

Percentage change (%) of daily unfiltered rainfall for (a) $\text{WISI} > +1\sigma$ and (b) $\text{WISI} < -1\sigma$ with respect to the mean of FMA season. Only areas in which the monthly average rainfall of FMA season is greater than the monthly average rainfall across the year are shown.

Contour interval is 10 percentage change, blue (red) shading indicates increases (decreases) in rainfall with respect to the climatology for ENSO-neutral years

Fig. 6.

FMA composites of daily unfiltered vertical pressure velocity anomalies (shading, 1×10^2 Pascals s^{-1}) and daily unfiltered horizontal wind vector anomalies (vector, $m s^{-1}$) at 500 hPa based on WISI for (a) $WISI > +1\sigma$ and (b) $WISI < -1\sigma$ during ENSO-neutral years from 1981 to 2016. The 95% confidence level is denoted by cross-hatching and thick black arrows for pressure velocity and winds, respectively. The pressure velocity contour interval is 0.4×10^2 Pascals s^{-1} . The reference vector in the top right-hand corner of each panel shows $3 m s^{-1}$.

Fig. 7.

FMA composites of daily unfiltered divergent wind anomalies (shading, $1 \times 10^6 \text{ s}^{-1}$) and daily unfiltered horizontal wind vector anomalies (vectors, m s^{-1}) at 250 hPa for (a) WISI $> 1\sigma$ and (b) WISI $< -1\sigma$ during ENSO-neutral years from 1981 to 2016. The 95% confidence level is denoted by cross-hatching and thick black arrows for divergent wind and horizontal wind, respectively. The divergent wind contour interval is $0.8 \times 10^6 \text{ s}^{-1}$. The reference vector in the top right-hand corner of each panel shows 5 m s^{-1} .

Fig. 8.

FMA composites of daily unfiltered divergence moisture anomalies (shading, $1 \times 10^5 \text{ g kg}^{-1} \text{ s}^{-1}$) and daily unfiltered horizontal winds anomalies at 850 hPa for (a) $\text{WISI} > 1\sigma$ and (b) $\text{WISI} < -1\sigma$ during ENSO-neutral years from 1981 to 2016. To calculate the daily divergence moisture anomalies, the daily divergence moisture is calculated first and then the daily climatology is removed of the daily divergence moisture. The 95% confidence level is denoted by cross-hatching and thick black arrows for divergence moisture and winds, respectively. The divergence moisture contour interval is $0.5 \times 10^5 \text{ g kg}^{-1} \text{ s}^{-1}$. The reference vector in the top right-hand corner of each panel shows 3 m s^{-1} .

Fig 9.

FMA composites of daily unfiltered anomalous SST (°C) for (a) WISI > 1σ and (b) WISI < -1σ during ENSO-neutral years from 1981 to 2016. Contours interval is 0.5 °C, red (blue) shadings areas indicate positive (negative) SST anomalies

Fig. 10.

FMA regression composites of time-lagged daily unfiltered OLR anomalies (shading, W m^{-2}) against WISI time series during ENSO-neutral years for 1981–2016. Only significant anomalies at 90% confidence level are shown. Contour interval is 2 W m^{-2} , blue (red) shading indicates positive (negative) OLR anomalies

Fig. 11.

FMA regression composites of time-lagged daily MJO-filtered OLR anomalies (shading, W m^{-2}) against WISI time series during ENSO-neutral years for 1981–2016. Only significant anomalies at 90% confidence level are shown. Contour interval is 2 W m^{-2} , blue (red) shading indicates positive (negative) OLR anomalies

Fig. 12.

FMA regression composites of time-lagged daily Rossby-filtered OLR anomalies (shading, W m^{-2}) against WISI time series during ENSO-neutral years for 1981–2016. Only significant anomalies at 90% confidence level are shown. Contour interval is 2 W m^{-2} , blue (red) shading indicates positive (negative) OLR anomalies

Fig. 13.

FMA regression composites of time-lagged sum of daily MJO-, Rossby-, and Kelvin-filtered OLR anomalies (shading, W m^{-2}) against WISI time series during ENSO-neutral years for 1981–2016. Only significant anomalies at 90% confidence level are shown. Contour interval is 2 W m^{-2} , blue (red) shading indicates positive (negative) OLR anomalies