Drought and Persistent Wet Spells over South America Based on Observations and the U.S. CLIVAR Drought Experiments

KINGTSE C. MO
NOAA/NWS/NCEP/Climate Prediction Center, Camp Springs, Maryland

ERNESTO H. BERBERY
Department of Atmospheric and Oceanic Science/ESSIC, University of Maryland, College Park, College Park, Maryland

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ABSTRACT

This study employs observations and the model simulations from the U.S. Climate Variability and Predictability (CLIVAR) Drought Working Group to examine extreme precipitation events like drought and wet spells that persist more than one season over South America. These events tend to persist over northeastern Brazil, the Guianas, and the west coast of Colombia, Ecuador, and Peru. They are least likely to persist over southeastern South America, which includes Uruguay, southern Brazil, and northeastern Argentina.

The U.S. CLIVAR simulations, particularly those of the National Center for Atmospheric Research (NCAR) Community Atmosphere Model, version 3.5 (CAM3.5), capture satisfactorily the impact of the El Niño–Southern Oscillation (ENSO) and the north tropical Atlantic (NTA) sea surface temperature anomaly (SSTA) signals on persistent extreme events and reproduce the mechanisms inducing the teleconnection patterns. The cold (warm) ENSO favors wetness (dryness) over Venezuela, Colombia, and northeastern Brazil and dryness (wet spells) over southeastern South America and southern Argentina. The NTA SSTAs alone tend to have a more local impact affecting mostly over northern South America in March–May.

The simulations show that when the two modes (ENSO and NTA) act in concert, the effects may become noticeable in different and remote areas of the continent, as they shift the probability of drought and persistent wet spells over different regions of South America. The impact is strong when the ENSO and the NTA are in opposite phases. For the cold (warm) Pacific and warm (cold) Atlantic, droughts (persistent wet spells) are intensified over southeastern South America, while persistent wet spells (droughts) are favored over the northern part of the continent. The changes in the patterns are regional and not as intense when both oceans are warm (or cold).

1. Introduction

Persistent extreme precipitation events such as drought and wet spells are the costliest natural disasters (Wilhite 2000). The consequences of drought reach a wide range of sectors including agriculture, commerce, hydropower, and many others; thus, a drought early warning system may mitigate their impacts. To build such a system, it is necessary to identify the preferred regions for their occurrence and understand the mechanisms associated with them. Usually, drought has favored regions to occur (Mo and Schemm 2008) as a consequence of both local and external forcing. The local factors such as the seasonal cycle, soil moisture, and moisture transport often provide the background conditions for extreme events to evolve and persist, but they do not trigger droughts or persistent wet spells. The persistent nature of drought and wet spells links them to large-scale low-frequency forcing such as sea surface temperature anomalies (SSTAs).

The focus of this paper will be on droughts and persistent wet spells over South America, where El Niño–Southern Oscillation (ENSO) and the SSTs in the northern tropical Atlantic (NTA) both have most prominent influences. The warm (cold) phase of ENSO shows dryness (wetness) over northeastern South America (Grimm 2011; Ropelewski and Halpert 1987, 1989). Aceituno (1988) reported an ENSO signal in parts of the

Corresponding author address: Kingtse C. Mo, NOAA/NWS/NCEP/Climate Prediction Center, 5200 Auth Rd., Camp Springs, MD 20746.
E-mail: kingtse.mo@noaa.gov

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Amazon. Ronchail et al. (2002) found the ENSO impact over the northeastern Amazon. The warm (cold) phase of ENSO is also in favor of wetness (dryness) over southeastern South America (SESA, identified here as 20°–35°S, 50°–60°W). The ENSO impact is seasonally dependent (Diaz et al. 1998; Cazes-Boezio et al. 2003) and is often modulated by regional forcing; Grimm (2003) suggested that soil moisture and temperature can be important factors in producing regional anomalies, while Grimm et al. (2007) reported a connection between spring soil moisture, surface temperature, orography, and summer monsoon rainfall. The ENSO effect often triggers extreme rainfall episodes in short time (Grimm and Tedeschi 2009; Pscheidt and Grimm 2009).

Even though ENSO composites of precipitation anomalies are seemingly stable, the detailed effects may vary from one episode to another owing to inter-ENSO variability (Drumond and Ambrizzi 2006).

Both observational and modeling studies (Hastenrath and Greischar 1993; Nobre and Shukla 1996; Chang et al. 2000; Taschetto and Wainer 2008) indicate that the NTA can influence precipitation in Brazil’s northeastern region. Ronchail et al. (2002) found that some sectors of Amazonia may also be impacted by the tropical Atlantic. The NTA can shift the location of the intertropical convergence zone (ITCZ), which in its southernmost (northernmost) position leads to increased (decreased) precipitation in Brazil’s northeastern region (Moura and Shukla 1981).

Because rainfall anomalies are linked to the ITCZ, the influence of the NTA tends to be local and somewhat limited. However, there are cases in which the ENSO mode and the NTA mode occur simultaneously (Uvo et al. 1998; Paegle and Mo 2002), in which case the NTA SSTAs can modify the ENSO impact on rainfall over South America. Marengo et al. (2008) showed that while some droughts (1926, 1983, and 1998) were induced by El Niño, this was not the case of the 2005 drought on the western Amazon. According to these authors, the causes of the 2005 drought were not related to El Niño but to a combination of factors, with the first one being the anomalously warm tropical North Atlantic. In addition, Zeng et al. (2008) state that the 2005 drought was caused by a combination of the 2003/04 El Niño and a dry spell attributable to a warm subtropical Atlantic. Yoon and Zeng (2009) concluded that the influence of the North Atlantic SST is stronger on the southern Amazon during the dry season, while the ENSO influence is stronger over the entire basin especially during the wet season.

In addition to the NTA and ENSO, wet and dry spells may be influenced by trends, the South Atlantic SSTAs, and the Pacific decadal oscillation (PDO). The trends of precipitation can increase the probability for drought or wet spells to occur. In the recent decades, there were reported positive precipitation trends over southeastern South America (e.g., Barros et al. 2000; Liebmann et al. 2004; Barros et al. 2008; Re and Barros 2009) and drying trends over the west coast of Peru, Colombia, Venezuela, and northern Brazil (Dai et al. 1998, 2004; Haylock et al. 2006). The influence of the south tropical Atlantic SSTAs on rainfall over northeast Brazil is similar to that of NTA, that is, by influencing the position of ITCZ (Hastenrath and Greischar 1993; Chang et al. 2000; Taschetto and Wainer 2008). Robertson et al. (2003) suggested that the South Atlantic SSTAs have no significant effect on the regional precipitation as the SSTAs may be driven by the atmospheric circulation. The relationships among different SSTAs modes such as ENSO and Atlantic SSTAs on rainfall can also be modulated by decadal forcing such as the abrupt change of the climate regime in 1976 (Kayano et al. 2009) and the Pacific decadal oscillations (Kayano and Andreoli 2007; Garcia and Kayano 2008). It is not the purpose of this article to do a full review of the subject, and more exhaustive discussions can be found in the literature (e.g., Taschetto and Wainer 2008; Kayano and Andreoli 2007; Grimm et al. 2007).

While several of the articles in the introduction discuss extremes for given regions, fewer discuss the overall South American patterns. Moreover, most of them describe the effects of SSTs on South American climate using observations and global reanalysis but very few from a modeling perspective. A recent set of carefully designed GCM model experiments by the U.S. Climate Variability and Predictability (CLIVAR) Drought Working Group has been released, and it is well suited to investigate how current models simulate the ENSO and NTA impacts on precipitation, as well as how the NTA modulates the impact of ENSO (Schubert et al. 2009). These experiments were performed by forcing several AGCMs with fixed prescribed SST anomalies in the Pacific (P) and/or in the Atlantic (A). The model simulations can also be used to analyze the mechanisms responsible for the linkages between SSTAs and droughts/persistent wet spells in South America. A drawback of these experiments is the assumption that the SST modes in different oceans are independent and do not interact, even though it is known that NTA SSTAs can be influenced by ENSO (e.g., Saravanan and Chang 2000; Mo and Hakkonen 2001). The U.S. CLIVAR simulations do not include the PDO or South Atlantic SSTAs. Therefore, these modes will not be included in the analysis of this article.

The objectives of this paper are as follows: first, to identify the preferred South American regions where drought and persistent wet spells tend to occur and to
examine the oceanic forcings associated with them; second, to evaluate whether the U.S. CLIVAR model simulations can capture the observed drought and wet spells associated with the leading SST anomalies; and last, to examine whether the models reproduce the mechanisms by which the teleconnections take place. The regional preference of drought and persistent wet spells will be discussed in section 3, followed in section 4 by an analysis on the impacts of the ENSO and the NTA modes, and their combination, on drought based on observations and the U.S. CLIVAR experiments. Conclusions will be presented in section 5.

2. Datasets and indices of extremes

a. Observations and reanalysis

The SST data are the monthly reconstructed SSTs from Smith et al. (1996) updated to 2006. The atmospheric variables at 2.5° grid spacing were obtained from the 1948–2006 National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) global reanalysis (Kalnay et al. 1996). Climatological monthly means are calculated for the base period 1948–2006, and monthly mean anomalies are defined as the departures from the monthly mean climatology for that month. The observed precipitation dataset used is the Climate Prediction Center (CPC) reconstructed 1948–2007 monthly mean precipitation based on rain gauges analyzed to a 1° × 1° grid (Chen et al. 2002).

b. The U.S. CLIVAR model experiments

The multimodel multi-institution AGCM experiments designed by the U.S. CLIVAR Drought Working Group are used to study the relationships between persistent SST forcing and droughts. The leading SST models are obtained by a rotated empirical orthogonal function (REOF) analysis. For each mode, the SST anomalies associated with the REOFs were added to the monthly mean climatology to form global SSTs, which are used to force the AGCMs (Schubert et al. 2009). The U.S. CLIVAR experiments are available online from http://gmao.gsfc.nasa.gov/research/clivar_drought_wg/index.html.

The first three REOF modes of annual mean SST variability explain 27.2%, 20.5%, and 5.8% of the variance, respectively (Schubert et al. 2009). These modes are reproduced in Fig. 1 based on the SST data from Smith et al. (1996). The first mode (Fig. 1a) is the trend mode that shows positive loadings in the three southern oceans. The associated rotated principal component (RPC) (Fig. 1b) indicates that the trends are positive but nonlinear. The second mode is the ENSO mode (Figs. 1c and 1d), with positive SSTAs extending from the central to the eastern Pacific and weak negative SSTAs in the North and South Pacific. The third mode is the North Atlantic mode (Fig. 1e), which shows a horseshoe-shaped pattern with positive SSTAs over the North Atlantic, north of 45°N, and over the tropical North Atlantic. This mode resembles the first non-ENSO mode described by Mestas-Nuñez and Enfield (1999) and also the first REOF for SSTAs in the Atlantic identified by Mo et al. (2009). The RPC 3 (Fig. 1f) and the time series of SSTAs averaged over the north tropical Atlantic (NTA; 5°–15°N, 20°–40°W) have a correlation of 0.9 for unfiltered monthly mean data.

The U.S. CLIVAR simulations included experiments with the three modes. In the same way that experiments forced with REOF 2 reflect the ENSO forcing, experiments forced with REOF 3 pattern reflect the NTA forcing. However, for the REOF 1 experiments (trend mode) the forcing was repeated each year, thus there is no real trend in the simulations. For this reason that set of experiments was not examined here. The experiments with the combinations of the Pacific (P) and/or the Atlantic (A) anomalies are labeled with subindices to denote c for cold, w for warm, and n for neutral SSTAs. For example, PcAw reflects a case in which the Pacific Ocean is cold and the Atlantic is warm. Details of the experiments and mean statistics are given in Schubert et al. (2009). In our study, we analyze 9 experiments: PnAn, PwAn, PcAn, PnAc, PnAw, PcAc, PwAc, PwAw, and PcAw. The climatological run PnAn is the control experiment with neutral conditions in both oceans.

Experiments from four models were analyzed here. The models are the NCEP Global Forecasting System (GFS) (Campana and Caplan 2005), the NASA Seasonal-to-Interannual Prediction Project (NSIPP-1) model (Bacmeister et al. 2000; Schubert et al. 2004), the Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model, version 2.1 (AM2.1; Delworth et al. 2006), and the NCAR Community Atmosphere Model, version 3.5 (CAM3.5). These experiments are labeled as GFS, NSIPP, GFDL, and CAM3.5, respectively. The GCM simulations cover a 50–51-yr period, except the GFS experiments that last 36 yr.

The climatological monthly means of all experiments (grand means) can be calculated for each calendar month by averaging over the 9 experiments together, and the results are similar to the climatological monthly means calculated from the PnAn experiment. For each experiment, the monthly mean anomaly is the departure from the grand monthly mean for that month, and the ensemble mean of any variable was obtained by taking the equally weighted mean of that variable from the GFS, NSIPP, CAM3.5, and GFDL model runs.
c. Indices of droughts and wet spells

The Palmer drought severity index (PDSI; Palmer 1965) and the standardized precipitation index (SPI; Hayes et al. 1999) are used to identify droughts and persistent wet spells. The PDSI was obtained from the NCAR data library (available online at http://www.cgd.ucar.edu/cas/catalog/climind/pdsi.html). The dataset was derived using historical precipitation and surface temperature data over landmasses at 2.5° resolution from 1870 to 2003 (Dai et al. 2004), but only the period from 1948–2003 was used in this study. The SPI index was computed with running 6-month precipitation totals (SPI6) from 1948 to 2007 for South America following the method outlined by McKee et al. (1993, 1995). A persistent drought (wet) episode is classified when the SPI6 index is less than −0.8 (greater than 0.8) or the PDSI is less than −2 (greater than 2) (Svoboda et al. 2002).

The procedure to calculate the frequency of drought or persistent wet spell occurrence in the simulations is the same as in Mo et al. (2009). For example, for the 36-yr GFS runs, the precipitation from 9 experiments was pooled to form a time series of $36 \times 12 \times 9$ months. The SPI6 index was calculated from the pooled time

FIG. 1. First three modes of SST variability. (a) The long-term trend SSTA pattern REOF 1 and (b) its associated RPC 1; (c) the Pacific SSTA pattern REOF 2 and (d) its associated RPC 2; (e) the North Atlantic SSTA pattern REOF 3 and (f) associated RPC 3. The contour interval is 0.4 nondimensional unit. Positive anomalies are shaded.
series of monthly mean precipitation, and the index values for the first year of each experiment were discarded. For each experiment, the frequency of drought occurrence was determined by counting the number of months \( N \) that the SPI6 is below \(-0.8\) at each grid point. Because the experiments for the different models have different lengths of integration, the frequency of drought (or persistent wet spell) occurrence is given by the ratio between \( N \) and the total length of the experiment. Large (small) percentage means that droughts are more (less) likely to occur. Finally, the statistical significance was assessed by the Monte Carlo method (Mo et al. 2009).

3. Regional preference of drought and persistent wet spells

In this section, we establish the preferred regions of drought or persistent wet spells occurrence based on the PDSI index. A dry (wet) event was selected when the PDSI is below (above) or equal to \(-2\) (+2). At each grid point, the total number of months \( N \) under wet or dry conditions was determined. From these extreme events, we selected a subset of cases that the positive or negative PDSI persisted more than one year. Figure 2a shows the number of months under persistent extreme conditions divided by \( N \) (expressed as a percentage).

Extreme events over South America have preferred regions to occur. Figure 2 shows that these preferred regions are northeastern Brazil including the northeastern Amazon, the Guianas, and the west coast of Colombia, Ecuador, and northern Peru, where 50% of the extreme events persist more than a year. Over SESA, including southern Brazil, Uruguay, and northeastern Argentina, drought or persistent wet spells are least likely to persist on interannual time scales.

The regional preference is in part modulated by the seasonal cycle of precipitation and moisture transport. For areas where the seasonal cycle is strong, drought occurring at the end of the rainy season will not get relief until the next rainy season, increasing the possibility for dryness to persist. For areas where the seasonal cycle is weak, rainfall from all seasons can contribute to the long-term measures of drought. Drought occurring in winter may get relief in the following seasons, and therefore wet or dry events are less likely to persist.

Figure 2b shows the difference between maximum and minimum precipitation from the monthly mean precipitation climatology calculated for each calendar month (i.e., the amplitude of the annual cycle), followed by the precipitation climatology for each season in Figs. 2c–f. In the tropics, the precipitation evolution depends on the seasonal migration of the ITCZ. Rainfall expands and shifts from Colombia and west Brazil southeastward in austral spring (September–October (SON)) to central Brazil and SESA. In austral summer (December–February (DJF)), convection moves to the Southern Hemisphere and the monsoon gets established. Rainfall occurs over western and central Brazil, the Amazon, and La Plata basin, but northeast Brazil is relatively dry. Over northeast Brazil, the wet season takes place from January to May with a maximum in March–May (MAM). The dry season occurs during austral winter (June–August (JJA)) when the ITCZ moves from the Southern Hemisphere to the north of the equator. Over the Amazon, where the seasonal cycle is largest (6 mm day\(^{-1}\)), rainfall occurs from November to April with a maximum in February. The weakest seasonal cycle is located over SESA, where the difference between \( P_{\text{max}} \) and \( P_{\text{min}} \) is 2 mm day\(^{-1}\) or less (Fig. 2b). Patagonia (south of 35°–40°S) is very dry with mean annual rainfall less than 2 mm day\(^{-1}\).

4. SSTA forcing

Before discussing the impact of ENSO on South American precipitation and the NTA’s modulation of the ENSO effects, we will discuss trends. These long-term trends do not trigger drought, but they will provide more favorable conditions for drought or persistent wet spells to occur.

a. Trends

The relation of the SSTA trend mode (Fig. 1a) on droughts over South America is examined with linear regressions of observations. The annual mean RPC 1 time series of the trend mode (Fig. 1b) was regressed against annual mean anomalies of precipitation and PDSI. The characteristic time (Trenberth 1984) for PDSI averaged over South America is about 22 months, so the statistical significance was assessed by assuming 1 degree of freedom every 2 yr. Although the general patterns of precipitation and PDSI seem to differ, there are regional similarities particularly in southeastern South America (Fig. 3). Some of the differences are caused because the PDSI not only depends on precipitation but also on surface temperature. The common features between precipitation and the PDSI are an increase of wetness over central and eastern Argentina, the border between Brazil and Bolivia and SESA, and an increase of dryness over the Guianas. Both variables also show dryness over the west coast of Ecuador and northern Peru. Overall, the regression map for PDSI (Fig. 3b) agrees with the PDSI trend from 1950–2002 reported by Dai et al. (2004).

The positive precipitation trends over SESA agree with previous observational reports of a remarkable increase of precipitation in the last few decades, with
FIG. 2. (a) Percentage of the total number of months of extreme events that persist more than one year. Contour interval is 0.1, and values greater than 0.5 are shaded. (b) The difference between the \( P_{\text{max}} \) and \( P_{\text{min}} \) of the monthly \( P \) climatology. The contour interval is 4 mm day\(^{-1}\); and \( P \) seasonal mean climatology from 1948–2006 for (c) DJF, (d) MAM, (e) JJA, and (f) SON. Contour interval is 2 mm day\(^{-1}\). Values greater than 4 (8) mm day\(^{-1}\) are shaded light (dark).
some locations reaching up to 50% increases (Barros et al. 2000; Liebmann et al. 2004; Barros et al. 2008; Krepper and Zucarelli 2009). Genta et al. (1998) detected long-term positive trends in the streamflow of the main rivers of La Plata basin in SESA after the mid-1960s and linked them to the increase of the eastern Pacific SSTAs. The general pattern with positive trends over SESA and dryness over the west coast of Ecuador and Peru also agrees with the trends reported by Haylock et al. (2006), who linked the trends to positive SSTAs in the eastern Pacific near the west coast of South America and positive SSTAs in the Atlantic centered near 35°S. Both are features of the REOF 1.

b. El Niño–Southern Oscillation

Seasonal composites of the observed SPI6 and PDSI were prepared for the cold and warm ENSO phases. Then, the composites’ differences between the cold and warm ENSO events were averaged over all seasons and presented in Figs. 4a and 4b. Consistent with the results reported by Ropelewski and Halpert (1987, 1989), Grimm et al. (1998), Grimm et al. (2000), Dai et al. (1998), Ronchail et al. (2002), and others, the two indices consistently show that during cold (warm) ENSO events with cold (warm) SSTAs over the tropical Pacific, wetness (dryness) is more likely to occur over Venezuela, Colombia, and northern Brazil including the northeastern Amazon. Dryness (wet spells) is more likely to occur over southeastern South America and southern Argentina.

After verifying the impact of ENSO from observations, we turn to analyze the U.S. CLIVAR experiments. The ensemble frequency of drought occurrence maps for PcAn (cold Pacific, neutral Atlantic) and PwAn (warm Pacific, neutral Atlantic) averaged over the four models (GFS, NSIPP, GFDL, and CAM3.5) is presented in Figs. 4c and 4d. They exhibit roughly the same patterns as observations (Figs. 4a and 4b), indicating that large-scale droughts and floods are captured satisfactorily in the U.S. CLIVAR experiments for the two phases of ENSO.

Figure 5 shows that, not only is the ensemble similar to observations, but that there is also a reasonable agreement among most models. Models perform more consistently for the warm ENSO experiment (PwAn; Figs. 5a–d) than for the cold ENSO experiment (PcAn; Figs. 5e–h). All models, but to a lesser degree NSIPP, are able to simulate the impact of ENSO on drought and persistent wet spells reasonably well. Still, there are differences in the details, particularly for the NSIPP simulation where the signal is limited to the northern tier of South America with a weaker signal over southern Argentina.

The CAM3.5 has the best performance in capturing the impact of ENSO on drought probably because of its higher horizontal resolution (1.4°×1.4°) that can reasonably resolve regional features like the low-level jet east of the Andes (Berri and Inzunza 1993; Berbery and Collini 2000; Silva and Berbery 2006; Nicolini and Saulo 2006; Silva et al. 2009). This low-level jet is a mesoscale feature that cannot be resolved by the coarse resolution of global reanalysis or equivalent resolution GCMs (Berbery and Barros 2002). Instead, global models tend to represent low-level jets as regions of intense winds as a result of a deflection of the Andes (Campetella and
Given the higher resolution of the CAM3.5 simulations, they are used to examine whether the model is able to capture the mechanisms relating droughts and wet spells in South America to the ENSO phenomenon.

The difference of annual mean vertically integrated moisture flux and its divergence $[D(Q)]$ between cold (PcAn) and warm (PwAn) experiments from CAM3.5 are presented in Fig. 6a, while the annual mean model
precipitation difference is given in Fig. 6b. The first 10 yr of the simulations are not included in the mean to avoid potential spinup biases. The mechanisms of the impact of ENSO are well known. The purpose here is to evaluate whether the model experiments are able to capture them.

The large-scale structure of the vertically integrated moisture flux convergence during the cold phase (Fig. 6a) shows larger divergence anomalies (reduced total convergence) in the Pacific, and thus the moisture flux arrows reflect a reduction in moisture convergence. The pattern is also consistent with increased anomalous fluxes from the Pacific to northern South America (Fig. 6a). The moisture flux convergence anomalies also show an increase (or reduced total divergence) over northern Brazil, again in agreement with the positive precipitation anomalies (Fig. 6b). Interestingly, there is also a weakening of moisture fluxes from the North Atlantic to northern Brazil. Both the reduced moisture flux convergence and sinking motions (Fig. 7) are associated with suppressed convection over the tropical Pacific. Over southern South America, the negative anomalies of precipitation relate to the (weak) anomalies of moisture flux divergence and appear directly related to the reduction of intensity of the low-level jet east of the Andes, as inferred from Fig. 6a. The strengthening (weakening) of the low-level jet during warm (cold) ENSO has also been discussed by Silva et al. (2009).

Figure 7 presents the circulation changes associated with ENSO as represented by the CAM3.5 model. The mean velocity potential (Fig. 7a) depicts the known wavenumber-1 pattern (e.g., Hoskins et al. 1989), with upper-level divergent winds over the western Pacific and wind convergence elsewhere. The longitudinal cross section of the circulation at the equator (Fig. 7b) indicates that most of the ascending (descending) motions are collocated with the upper levels of wind divergence (convergence). This circulation structure is representative of the Walker circulation as described, for example, by Webster (1983). Also in agreement with Webster (1983), a second and more localized region of ascending motions is observed at 60°W, over northern South America. Figures 7c and 7d depict the CAM3.5 model differences PcAn – PwAn, which show an ENSO influence on the global pattern of velocity potential and
corresponding ascending motions associated with suppressed convection in the Tropical Pacific where the divergence anomalies are located. Again, this simulated pattern is consistent with previous observational results (Mo and Rasmusson 1993). The cold phase of ENSO exhibits a westward shift and a reduction of the intensity of the divergent wind and an increase of the wind convergence elsewhere. Moreover, there is a reduction of the ascending motions (or increase of the descending motions) over the Pacific, while over northern South America the cold phase exhibits stronger rising motions than the warm phase. This structure of reduced ascending motions over the Pacific and increased over South America is indicative of a weakened Walker circulation (Bjerknes 1969; Julian and Chervin 1978).

During cold (warm) ENSO events, positive (negative) 500-hPa omega values (Fig. 7e) are indicative of sinking (rising) motions over the midtropical Pacific region and the South Pacific convergence zone (SPCZ) region. The simulated pattern suggests modulations to the local Hadley circulation that are similar to those described by, for example, Oort and Yienger (1996) and Garcia and Kayano (2008). Figure 7e shows sinking (rising) motions over the tropical Pacific have concurrent rising (sinking) motions in the subtropics at the same longitudes, particularly on the winter hemisphere (not seen here because of averaging over the whole year). In addition, the opposite sign of the omega anomalies over tropical South America is yet another indication of the Walker circulation that has its compensating branch collocated with positive (negative) rainfall anomalies (Fig. 6b).

The ENSO phases are known for modulating the Southern Hemisphere storm track (Solman and Menendez 2002), which can be represented by the standard deviation of the unfiltered meridional wind (Chang 1993). Figure 7f presents the storm track at 200 hPa (contours) for the PnAn simulations. The pattern is similar to that found from global reanalysis for austral winter by Berbery and Vera (1996), but in this case it is more uniform and smoother because it is the annual mean pattern. The storm-track changes between the cold ENSO phase and the warm ENSO phase (PcAn − PwAn) are color shaded. Most of these changes are located over the Pacific Ocean and South America. The negative values over the continent indicate a weaker and southward-shifted storm track during the cold phase and a more intense and northward-shifted storm track during the warm phase in agreement with Solman and Menendez (2002). The changes in precipitation (Fig. 6b) are located precisely where the corresponding storm-track changes are observed. Overall, the CAM3.5 not only captures the influence of ENSO on drought, but it also captures the physical mechanisms for such connection.

c. The north tropical Atlantic mode

The major influence of the NTA SSTAs is over northern South America (Uvo et al. 1998). The PDSI composite based on the warm minus cold NTA (Fig. 8a) shows negative values (dryness) extending from north-eastern Brazil to central western Brazil including the Amazon. Because of the seasonality of rainfall (Fig. 2), the major contribution to annual drought indices comes from MAM. As expected, the PDSI composite for MAM (Fig. 8b) shows a similar pattern as for the whole year (Fig. 8a), but the signal is stronger. The MAM observed precipitation difference between the warm and cold NTA phases (Fig. 8c) reveals a strong dry signal over northeastern Brazil consistent with the
Fig. 7. (a) Velocity potential at 200 hPa for the PnAn simulations. The unit is $5 \times 10^6$ m$^2$ s$^{-1}$; (b) height–longitude cross section showing the longitudinal circulation along the equator (streamlines). The shades correspond to vertical motions, with pink indicating ascending motions and blue descending motions. (c),(d) As in (a),(b), but for the difference PcAn − PwAn; (e) vertical velocity (omega) at 500 hPa for the difference PcAn − PwAn. Contour interval is 0.01 Pa s$^{-1}$; (f) the annual mean storm track at 200 hPa. Contour intervals are 2 m s$^{-1}$ starting at 16 m s$^{-1}$, and storm-track differences for PcAn − PwAn are shaded.
PDSI but adding a weak wet signal over Colombia, Ecuador, and SESA. In general, the influence of NTA on drought is more regional and weaker than the ENSO influence.

The models' simulations for the Atlantic variability exhibit large spread. The signals in the frequency of occurrence for PnAw and PnAc are weak and do not pass the field significance test, and consequently they are not...
shown. As with observations, the influence of NTA is limited to MAM, and the signal becomes weaker if all seasons are pooled together. Overall, the MAM ensemble precipitation differences between the PnAw and PnAc phases (Fig. 8d) somewhat resemble the general pattern of the rainfall response to NTA such as strong wetness over Colombia and dryness over northeastern Brazil, but there are also large errors. For example, the ensemble does not capture rainfall anomalies in the transition region over the Guianas, where the patterns in Figs. 8c and 8d have opposite signs in comparison with observations. Neither does the ensemble pick up the weak positive anomalies over SESA. Among all models, the CAM3.5 performs best (Fig. 9b). The CAM3.5 captures positive anomalies (wetness) over Ecuador, Colombia, Venezuela, and the Guianas, with negative anomalies over central and eastern Brazil and weak wetness over SESA. Despite the spatial patterns being similar, the magnitudes are clearly not. Observations (Fig. 8c) indicate strong dryness over eastern Brazil and weak wetness toward the northwestern sector. Instead, the CAM3.5 model, and as seen the ensemble simulations as well, produces strong wet anomalies and somewhat weak dryness; moreover, the centers are not strictly collocated. With the model errors in mind, we will use the CAM3.5 simulations to study the mechanisms responsible for the connection between precipitation and the NTA. The discussion will be focused on MAM, when the signal is stronger and statistically significant. This will also serve as a reference for the next section when we discuss the combined influence of ENSO and NTA on precipitation.

During MAM, SSTAs in the NTA have an important role in modulating the northeast and southeast trade winds and the ITCZ, thus their location will influence the rainfall in northeastern Brazil and northern South America in general. When the NTA is warm, the moisture flux anomalies are northward over northern South America, indicating increased moisture flux convergence over the ITCZ; however, these anomalies also indicate that there is less moisture transported from the NTA through northeastern Brazil to the Amazon. Therefore, warm NTA is in favor of drought in northeastern Brazil and the Amazon (Fig. 9b). The increased rainfall over northern South America seems to respond to the larger moisture transported from the warm North Pacific to northwestern South America just north of the equator (Fig. 9b). These model results agree with the observational study of Ronchail et al. (2002).

The increased rainfall over warm SSTAs in the NTA corresponds with the increased rising motion indicated by negative 500-hPa omega (Fig. 9c). The slightly positive omega near the coastline of northeastern Brazil also suggests a downward branch of the regional Hadley circulation that would favor the regional dryness. This is confirmed by the cross section showing the changes in meridional circulation averaged for the band 65°S–35°W (Fig. 9d). The two main features are the increase of ascending motions at approximately 0°–20°N and increased descending motions approximately between 25°S–5°N. The ascending and descending branches reflect the north–south shift on the regional Hadley cell associated with the changes in the location of the SSTAs. The CCM3.5 is thus able to simulate the mechanisms responsible for the influence of the NTA on rainfall reported by Moura and Shukla (1981).

d. The combined effect of the Pacific and tropical North Atlantic Oceans

Observational datasets are not long enough to study the combined ENSO and NTA on drought over South America, so we turn to the U.S. CLIVAR experiments. Figure 10 presents the ensemble mean frequency of drought occurrence for combinations of the Pacific and Atlantic modes (PwAc, PcAw, PcAc, and PwAw). As seen earlier, when the Pacific is warm alone, droughts are more likely to occur over northern South America except Venezuela and Colombia, while wetness is favored over SESA and southern Argentina (Fig. 4d). When the Pacific is warm and the North Atlantic is cold (PwAc, Fig. 10a), the wetter region over southeastern South America expands toward the north and now includes parts of southern Brazil, while over the northern part of the continent dryness becomes more localized and intense (in comparison to the ENSO effect alone). The combination of the cold Pacific and warm Atlantic (PcAw) in Fig. 10c affects not only the wetter region over the northeastern part of the continent but also the drought region over Argentina, which becomes better defined in shape and intensity, with an additional intensification of dryness over the western Amazon and Peru/Ecuador.

If the two modes are in their warm phase (PwAw, Fig. 10b), the wettest south of 30°S (central Argentina) remains about the same as in the case of the warm Pacific with neutral Atlantic (PwAn) discussed in relation to Fig. 4d, but now the main effect is a shift and increased intensity of the drought toward southeastern Amazonia and northeastern Brazil. The reverse pattern (Fig. 10d) is seen for the case in which both oceans are cold (PcAc).

All these results indicate that the influence of ENSO on drought changes according to the sign of the NTA SSTAs. The changes in the signal of the cold phase of ENSO due to modulations by the NTA SSTAs (experiments PcAc and PcAw) are further analyzed from the CAM3.5 simulations to complement the cold Pacific with neutral Atlantic (PcAn) case that had been
discussed in section 3b. The case of the warm Pacific is basically the reverse of what will be shown here.

Starting with the modulations in the regional Hadley cell, Fig. 11 presents a cross section of the meridional circulation averaged over the 65°–35°W band. When both oceans are cold (Fig. 11a), there is a slight increase of the ascending motions around 0°–10°S (with respect to PcAn) and anomalous descending motions around...
Figure 12b shows a consistent pattern of increased precipitation over northeastern Brazil and slight dryness farther south.

If the cold Pacific is combined with warm SSTAs over the NTA (PcAw, Fig. 11b), the two modes lead to a markedly stronger ascending branch of the Hadley cell between 0°–20°N, with negative (descending) anomalies approximately between 5°–40°S. Once again the precipitation anomalies (Fig. 12d) are consistent with this picture, indicating not only positive precipitation anomalies.
collocated with the ascending motions, but also the marked dryness at the same latitudes of the anomalies of the descending motions.

The combination of the two modes, ENSO and NTA, does not only affect the meridional circulations but also has an influence on the transports of moisture over large areas. Figures 12a and 12c show that the PcAw case has a stronger moisture flux convergence than the case PcAc, just like it occurred with the strength of the meridional circulation. Moreover, the stronger tropical convergence has associated a weaker low-level jet east of the Andes with reduced moisture transport from Amazonia into southeastern South America (in agreement with Silva et al. 2009), thus being another element in the reduction of the rainfall magnitude in southeastern South America.

In summary, a cold Pacific together with a warm Atlantic reinforce the signals in such a manner that extratropical/subtropical droughts become stronger and more distinct, while the tropical regions experience stronger wet spells. The opposite effects are noted for the case of warm Pacific with a cold Atlantic (not shown). The changes in the patterns when both oceans are simultaneously warm (or cold) are regional and not as intense as the cases of opposite signs.

5. Concluding remarks

The drought and persistent wet spells over South America are analyzed using both observations and the U.S. CLIVAR multi-institution multimodel AGCM experiments. Drought and persistent wet spells over South America have preferred regions to occur and persist on interannual time scales. They are most likely to persist over northeastern Brazil, the Guianas, and the west coast of Colombia, Ecuador, and Peru. Over SESA, extreme events are least likely to persist.

Overall, the ensemble mean model simulations are able to capture satisfactorily the signals over South America, even though they differ in details. For weaker or regional signals, the models have large spreads, but stronger signals (e.g., ENSO) are more consistently reproduced by all models. Among the four sets of model simulations, the CAM3.5 has a more realistic performance, probably due to the higher resolution, so these model outputs are used to evaluate the mechanisms related to the extreme wet and dry episodes.

The U.S. CLIVAR simulations agree with observations in that at interannual time scales, ENSO has the largest impact on drought and persistent wet spells over South America. The cold (warm) ENSO phase favors wetness (dryness) over Venezuela, Colombia, and northern Brazil including the northeastern Amazon and drought (wetness) over southeastern South America and southern Argentina.

In agreement with previous observational studies, CAM3.5 simulations show that the influence on the extreme precipitation events over the tropical continent is due to changes in the intensity of the Walker circulation in response to changes in convection over the tropical Pacific. In addition, the modifications of the general circulation result in a weakening of the storm track over the South American continent. Regionally, the changes of circulation over South America in response to ENSO also affect the moisture transport from the tropics into the subtropics and extratropics through modulations of the low-level jet east of the Andes that favors or suppresses occurrence of extreme events. The U.S. CLIVAR
experiments are able to capture both the ENSO impact on rainfall and the mechanisms responsible for such impact.

The North Tropical Atlantic SSTAs have a regional and weaker influence limited to the subtropics. The annual signal is dominated by the intensity of the MAM anomalies, particularly for northeast Brazil, where rainfall is determined by the location of the ITCZ in direct response to the NTA SSTAs. The NTA warm (cold) SSTAs decrease (increase) the moisture transports into the Amazon so the occurrence of drought (wet spells) increases. A more relevant aspect of the NTA is that it modulates the ENSO influence by shifting the probability of extreme events over different regions of South America. For cold or warm concurrent SSTAs in both the Pacific and the Atlantic (experiments PcAc and PwAw), rainfall changes tend to be at smaller scales. However, when the Pacific and the Atlantic SSTAs are opposite in phase, the area of impact shifts northwestward to Colombia, Venezuela, and the Guianas. More importantly, when in opposite phases, the NTA and ENSO may act to intensify the precipitation anomalies and better define the affected regions over southeastern South America and southern Argentina.

**FIG. 12.** (a) Annual mean differences of moisture flux divergence (in color) and vertically integrated moisture flux (qflux) (vector) between the PcAc and PnAn from the CAM3.5 experiments. The units of D(Q) are indicated by the color bar in mm day$^{-1}$. The unit vector is 45 g (cm × s)$^{-1}$. (b) As in (a), but for the P difference. Areas where positive (negative) values are statistically significant at the 5% level are colored according to the color bar. The units are mm day$^{-1}$. (c) As in (a), but for the PcAw experiment. (d) As in (b), but for the PcAw experiment.
The areas where drought and wet spells are most likely to persist are also regions dominated by the influence of ENSO and NTA. Rainfall anomalies over these areas have a seasonal cycle greater than 2 mm day\(^{-1}\). One interesting point is that SESA is also influenced strongly by ENSO. However, wet or dry precipitation anomalies over that region do not seem to persist on interannual time scales. This region has a very weak seasonal cycle, and the SSTAs from different oceans can influence rainfall during different seasons (as suggested by Diaz et al. 1998 and Liebmann et al. 2004).

Several components to build an early warning system for drought have been addressed in this paper. We have presented the regions over South America where droughts and wet spells are more likely to develop and persist. The ENSO is a large contributor to drought with its impacts modulated by the Atlantic SSTAs, which even in some regions can dominate. We have also shown that current models like those used for the U.S. CLIVAR experiments can reproduce the patterns and mechanisms resulting in extremes. Currently, most coupled model systems are able to forecast ENSO a few months in advance (Coelho et al. 2003; Palmer et al. 2004; Saha et al. 2006). Those forecasts, combined with the monitoring of the Pacific and Atlantic Oceans SSTs, and with drought indices such as the SPI and PDSI over the particular regions discussed here, could serve as the basis to build such an early warning system to mitigate the impact of drought over South America.

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