Interdecadal Variability of the South American Precipitation in the Monsoon Season

ALICE M. GRIMM AND JOÃO P. J. SABOIA

Department of Physics, and Post Graduate Program on Water Resources and Environmental Engineering, Federal University of Paraná, Curitiba, Brazil

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ABSTRACT

Interdecadal variability modes of monsoon precipitation over South America (SA) are provided by a continental-scale rotated empirical orthogonal function analysis, and their connections to well-known climatic indices and SST anomalies are examined. The analysis, carried out for austral spring and summer, uses a comprehensive set of station data assembled and verified for the period 1950–2000. The presented modes are robust, consistent with previous regional-scale studies and with modes obtained from longer time series over smaller domains. Opposite phases of the main modes show differences around 50% in monthly precipitation. There are significant relationships between the interdecadal variability in spring and summer, indicating local and remote influences. The first modes for both seasons are dipole-like, displaying opposite anomalies in central-east and southeast SA. They tend to reverse polarity from spring to summer. Yet the summer second mode and its related spring fourth mode, which affect the core monsoon region in central Brazil and central-northwestern Argentina, show similar factor loadings, indicating persistence of anomalies from one season to the other, contrary to the first modes. The other presented modes describe the variability in different regions with great monsoon precipitation. Significant connections with different combinations of climatic indices and SST anomalies provide physical basis for the presented modes: three show the strongest connections with SST-based indices, and two have the strongest connections with atmospheric indices. However, the main modes show connections with more than one climatic index and more than one oceanic region, stressing the importance of combined influence.

1. Introduction

Decadal and interdecadal climate oscillations and their impacts around the globe have motivated several observational studies at the global scale or in specific regions (e.g., Zhang et al. 1997; Cayan et al. 1998; Mestas-Nuñez and Enfield 1999; Tourre et al. 1999; Folland et al. 1999; Power et al. 1999; Barlow et al. 2001; Chelliah and Bell 2004; Meinke et al. 2005; Knight et al. 2006; Folland et al. 2009). They show that these oscillations involve both oceanic and atmospheric fields. Climate variations in this time frame control water availability, affect ecosystems, influence farming practices (e.g., Podestá et al. 2009), and modulate higher-frequency variability and extreme events such as floods and droughts. Furthermore, they affect models’ skill (Grimm et al. 2006).

Knowledge of this kind of variability is important to medium- and long-range water resources management in South America (SA), particularly in planning hydropower generation. Since the hydroelectricity distribution networks are interconnected in Brazil and some countries in the continent share hydropower generation plants that depend on rainfall over large basins, it is useful to know the large-scale, continental modes of interdecadal precipitation variability and how they impact the precipitation regimes. Even the determination of the return period of floods, important for the design of dams, should take into account interdecadal variations. Besides, this low-frequency natural climate variability needs to be well characterized and understood in order to validate climate models and to achieve more reliable detection of anthropogenic climate change (e.g., Parker et al. 2007).

Interdecadal variability of precipitation has been reported and analyzed in particular regions of SA (e.g., Robertson and Mechoso 1998; Lucero and Rodriguez 2001; Compagnucci et al. 2002; Vargas et al. 2002; Piovano et al. 2002; Rusticucci and Penalba 2000;
Kayano and Andreoli 2004; Kayano and Sansigolo 2009; Agosta and Compagnucci 2012), but so far, no comprehensive continental-scale analysis has been carried out. Some of those previous regional results will be compared to the results of the present study.

The focus of this study is on the austral summer monsoon season, because it is the rainy season in most of SA (e.g., Fig. 1 in Grimm 2011). It undergoes strong interannual variability, which is affected by remote influence, such as the El Niño–Southern Oscillation (ENSO; e.g., Ropelewski and Halpert 1987; Rao and Hada 1990; Grimm 2003, 2004), and also by local influence of surface–atmosphere interactions that establish a relationship between spring and summer precipitation in parts of SA (Grimm et al. 2007). This relationship seems to be modulated by interdecadal variability (Grimm and Zilli 2009).

In the present study, interdecadal variability comprises periods of 8 years and greater. Precipitation variability in this band is substantial over SA in summer (Fig. 1, left panel). It amounts to more than 30% of the summer precipitation variance in extensive regions with a rainy season in the summer and a dry winter, such as central SA, northern and western Argentina, and western tropical SA, south of the equator (Grimm 2011). Besides, it contributes significantly to the variability of regions where the summer precipitation variance is very high (Fig. 1, right panel).

Further motivation for a continental-scale analysis of interdecadal oscillations is provided by the inspection and comparison of filtered series of summer precipitation in regions chosen in Fig. 1 for great contribution of interdecadal variability to total variability [such as central-northwestern Argentina, central Brazil, and the western Amazon (Peru) (Fig. 1, left)] and/or high variability in summer [such as the eastern continental SA convergence zone (SACZ) and the region north of it, the southern Paraná–La Plata basin, the northeastern Amazon, and northern SA (Fig. 1, right)]. There are interesting relationships that could be manifested in continental-scale modes (Fig. 2). The interdecadal oscillations in central-northwestern Argentina (Fig. 2a) and in the core monsoon region in central SA (Fig. 2b) show similarities: dry conditions in the 1960s and early 1970s, with a sharp variation in the mid-1970s, and wet conditions till the mid–late 1990s, when a decreasing tendency starts. On the other hand, the variations in central-northwestern Argentina (Fig. 2a) are different from those in the middle-lower Paraná–La Plata basin [eastern SA subtropics (Fig. 2c)]. While in the former region precipitation increased sharply in the mid-1970s, in the latter it started increasing in the early 1960s, besides showing differences in other periods. The eastern continental SACZ (Fig. 2d), in the upper Paraná–La
Plata basin, displays variations approximately opposite to those of the middle-lower part of the basin (Fig. 2c). These dipole-like variations are relevant, since these two very populated regions in South America contain important hydroelectric power plants, including the world’s largest hydroelectric power plant, Itaipu. There are also indications of an east–west tropical dipole of variability south of the equator, between southern northeast Brazil, north of the SACZ (Fig. 2e), and the western Amazon (Fig. 2f). Yet variations in the northeastern Amazon (Fig. 2g) and in northern SA (Fig. 2h) do not show marked similarity or opposition to those in other displayed regions.

The main difficulty in characterizing continental-scale interdecadal modes of variability has been to assemble reasonably reliable rainfall data over a sufficiently long period and large spatial coverage. Therefore, this study covers an important lacuna of information and is a first step to establish a more detailed connection between the continental-scale interdecadal climate variability over
SA and the global-scale interdecadal oceanic and atmospheric oscillations.

It is our objective to characterize the continental-scale interdecadal variability of the South American precipitation in the monsoon season, comprising austral spring (September–November) and summer [December–February (DJF)] at the beginning and the peak of the monsoon season, respectively. The analysis also includes regions not affected by the monsoon. We examine the following aspects: 1) spatial structure and temporal evolution; 2) main relationships between spring and summer; and 3) statistical connections to sea surface temperature (SST) anomalies and known climatic indices.

The datasets, their verification procedures, and the methods used are described in section 2. Modes of variability for spring and summer obtained from the continental dataset are described in section 3, while in section 4, modes obtained from a dataset with longer time series but less spatial coverage are described and related with the continental-scale modes. The relationship between spring and summer modes is disclosed in section 5. The connection of the precipitation modes to SST anomalies and well-known climatic indices is broached in section 6, while section 7 shows that the continental-scale results are consistent with previous regional or local-scale studies. Section 8 summarizes the findings and draws conclusions.

2. Data and methods

a. Data

An extensive set of monthly precipitation data from around 10 000 stations in SA has been assembled and verified for the period 1950–2000. Most of these data are from the Brazilian Water Agency (ANA) with contributions from hydrometeorological institutes from Argentina, Uruguay, Paraguay, and Peru. To obtain better coverage in northern South America, a 1° gridded set described in Liebmann and Allured (2005) is also used in that region.

The final precipitation series used in this study are obtained on a 2.5° × 2.5° grid of 176 points over South America by averaging the station data within circles of radius \(1.25\sqrt{2} = 1.77^\circ\), the centers of which are separated by 2.5° (Fig. 3a). The results obtained from this extensive dataset are compared with those obtained from the gridded dataset Climatic Research Unit (CRU) 1.0 (Hulme et al. 1998), with longer time series but less spatial coverage (Fig. 3b). It was chosen because its resolution (2.5° × 3.75°) is comparable to our grid, and its grid points have interpolated data, avoiding filling.
with climatology. The CRU data period 1900–93 was selected as the best compromise between spatial and temporal coverage.

The station precipitation data are exhaustively examined in order to find different types of detected problems: missing data recorded as zeros, unrealistic large values, same values registered for two stations far away from each other, and improbable changes in the precipitation regime. The main problem is the record of missing data as zeros, which is especially difficult to detect in regions with a well-defined dry season. The code developed to detect this problem includes many tests based on the station climatology, phases of climate variability (e.g., El Niño and La Niña), and comparison with data from neighbor stations.

In grid boxes with fewer than 10 stations available, missing station data are estimated using regression equations with data from neighbor stations, selected by criteria of proximity, minimum period of data, and significant correlation with the data of the station under focus (Tabony 1983). The estimated value is obtained from the weighted average of the regressed values and undergoes a procedure that provides continuity between the estimated and the observed values. The weights are based on the correlation coefficients between the station data and the data of the selected neighbor stations. The interpolation of missing data is only carried out for northern and central-western Brazil, because there is no need to do it in other regions.

After the average series are calculated for all grid points, a few missing values in a grid point may still be estimated from the eight neighbor grid points, following the same criteria mentioned above. Grid boxes for which the number of stations contributing to the average is too different in different periods are verified with respect to inhomogeneities and corrected. Furthermore, a visual inspection of the gridpoint series is carried out in order to detect very unusual values. The final distribution of the grid points with monthly rainfall series in the period 1950–2000 is shown in Fig. 3a. This dataset is named the continental dataset.

Although finding sources and disclosing mechanisms of the interdecadal precipitation modes is beyond the objectives of this manuscript, a preliminary analysis is included of their connection (meaning correlation of their factor scores) with well-known climatic indices [the Atlantic multidecadal oscillation (AMO), North Atlantic Oscillation (NAO), interdecadal Pacific oscillation (IPO), Pacific decadal oscillation (PDO), tropical South Atlantic (TSA), tropical North Atlantic (TNA), southern annular mode (SAM), and northern annular mode (NAM)]. Two differently calculated AMO indices (different data origin, methods, and regions used in the North Atlantic to average SST: 0°–70°N and 0°–60°N) are used to show that the definitions of the indices influence their correlation with the modes. All those indices are available online, along with their definitions (see the appendix). Their seasonal averages obtained from monthly unsmoothed values are filtered as our data (see next section) and correlated with the presented precipitation modes. Connections to SST anomalies are analyzed using the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) dataset (Rayner et al. 2003).

b. Methods

Rainfall totals for each season and each year are submitted to a 9-point Gaussian moving average filter. The weights, which follow a Gaussian curve, are calculated to retain oscillations with periods equal or greater than 8 years (Mitchell et al. 1966). This threshold was chosen to exclude the ENSO-related and other interannual variability but include the near-decadal variability that may contribute to the 9.0-yr oscillation detected by Robertson and Mechoso (1998) in streamflows of the Paraná/La Plata River basin. The filtered data are submitted to empirical orthogonal function (EOF) analysis using the correlation matrix. The covariance matrix gives very similar results.

Rotation of a certain number of modes is carried out in order to identify intrinsic modes of climate variability in particular regions, with common underlying mechanisms. The rotation used is varimax, which preserves the orthogonality of the time coefficients (factor scores) without imposing orthogonality on the spatial factor loading maps. Unrotated modes are constrained to be spatially orthogonal, which can produce nonphysical modes (Richman 1986). Besides, if unrotated EOFs were used, the data sample would not be large enough to fulfill the North et al.’s (1982) rule of thumb for considering two modes as independent or faithfully representing true EOFs (North et al. 1982; Richman 1986). However, the sampling problem (as well as other problems inherent to unrotated EOFs) is greatly reduced by using rotated solutions, even if the neighboring eigenvalues are close, as long as there is structure within the data (Richman 1986).

According to the Kaiser’s rule (Wilks 1995), the rotation includes the modes with eigenvalues equal to or greater than 1. As the North et al.’s (1982) rule does not apply to rotated modes, the robustness of the rotated EOFs (REOFs) is examined by repeating the analysis withholding different sets of years of the record: 1950–2000, 1955–2000, and 1960–2000. This produces very similar sets of five rotated leading modes, which is not true for unrotated modes.

The significance of correlation between series of filtered data (whose members are not independent) is
assessed by a Monte Carlo procedure: 10,000 random permutations of the series are generated and the correlation between them is calculated for each permutation. The number of times in which the absolute value of the correlation between the random permutations exceeds that between the original data is counted. This value, divided by 10,000, gives the level of significance.

The significance of differences between means is tested using unequal variance Student’s *t* test (Wilks 1995), which is more adequate for samples that could have different statistical characteristics in different phases of interdecadal variability.

3. Continental modes of interdecadal variability

For spring, only the first two REOFs are shown, because these have highest loadings in regions with much rain in spring (Fig. 4). For summer, the first five REOFs are presented, because they are robust, their highest loadings cover different regions in which summer is part of the rainy season, and they are differently connected to climatic indices and SST anomalies (Fig. 5).

To emphasize that the REOFs here presented are not just artifacts of the analysis but represent real oscillations in the data, the maps of REOF factor loadings are replaced with the correlation maps between the factor scores and precipitation data, including the respective statistical significance. The patterns of significant correlations and of factor loadings are completely consistent; therefore, the factor loadings maps are not shown.

a. Spring

The spring REOF1 (Fig. 4a; 18.4% of the variance), is focused on the variability in central-east and southeast SA, where the significant correlations have opposite

![Fig. 4. First two spring REOFs. (bottom) Factor scores. (top) Correlation coefficients between the factor scores and precipitation data and their significance levels. The isolines interval is 0.1, and the zero isoline is omitted. Colors indicate significance levels of positive or negative correlation coefficients. The negative signs inserted before the levels of significance indicate that they refer to negative correlation coefficients. Areas with data but no significant correlation, are shaded in gray.](image-url)
Fig. 5. As in Fig. 4, but for the first five summer REOFs.
sign, indicating a dipole-like variability. This variability is intense in the period 1950–80, with a strong change of phase in the 1960s, but weakened during the 1980s and 1990s. The REOF2 (Fig. 4b; 15.3% of the variance) concentrates on the variability in central-southeast Brazil. Although these two modes share a negative phase in the 1970s, they have different oscillations in other periods.

The next three modes (not shown) are focused on regions in which spring is relatively dry compared to summer. REOF3 (15.0% of the variance) focuses on northwestern and northeastern SA. REOF 4 (10.9% of the variance) represents variability in northern Argentina, northern Paraguay, western Uruguay, and part of central-west Brazil, covering mainly the western part of the Paraná–La Plata basin, especially the Paraguay River basin. REOF5 (7.4% of the variance) focuses on the eastern equatorial Amazon basin, around the mouth of the Amazon River.

b. Summer

The summer REOF1 (Fig. 5a; 17.9% of the variance) displays a dipole-like pattern of variability between central-east and southeast SA as the first spring mode, although more concentrated on the southern parts of these regions, particularly the continental SACZ and the lower Paraná–La Plata basin. Opposite signs in the anomalies are also shown between central-east and northwest SA. The summer (and spring) dipole of anomalies between central-east and southeast SA is a recurrent variability pattern. It also appears as the first mode of intraseasonal and interannual variability of observed precipitation (Ferraz 2004; Grimm and Ambrizzi 2009; Grimm and Zilli 2009; Grimm 2011) and among the first modes for outgoing longwave radiation (OLR; Nogués-Paegle and Mo 1997) and merged datasets (satellite estimates + rain gauge data) (Zhou and Lau 2001; Nogués-Paegle and Mo 2002). The present study shows that it is also the first mode of interdecadal variability of observed precipitation. As the spring REOF1, it shows a strong change of phase in the 1960s and weaker variability during the 1980s and 1990s.

REOF2 (Fig. 5b; 15.7% of the variance) is focused on the variability in central/northern Argentina and central SA, around the core monsoon region with the highest summer precipitation (Fig. 1 of Grimm 2011). This mode best represents the “climatic shift” in the mid-1970s. It resembles spring REOF4 (not shown).

REOF3 (Fig. 5c; 12.8% of the variance) also displays a dipole pattern between central-east and southeast SA, but, unlike the first mode, it is focused on the northern parts of these regions. Besides, an east–west dipole over tropical SA, south of the equator, is disclosed by this mode. It shows a nearly decadal variability in the analyzed period.

REOF4 (Fig. 5d; 12.3% of the variance) is strongly focused on the variability in northeast SA, where summer is the beginning of the rainy season, but also has weaker influence on part of southeast SA. It shows a nearly decadal variability in most of the analyzed period.

REOF5 (Fig. 5e; 7.7% of the variance) is concentrated on the variability in a region south of the SACZ, in the middle-upper Paraná–La Plata basin. Its temporal variability has a predominant period a little shorter than decadal.

These modes represent the variability depicted in Fig. 2, and more. REOF1 reflects the out-of-phase variability between SACZ and the southern Paraná–La Plata basin shown in Figs. 2c and 2d (with some differences, since Fig. 2c also includes the middle part of this basin). REOF2 represents the in-phase variability of the core monsoon region and central-northwestern Argentina, displayed in Figs. 2a and 2b. REOF3 incorporates the east–west dipole-like variability shown in Figs. 2e and 2f, besides the variability in the middle Paraná–La Plata basin, included in Fig. 2c. REOF4 represents the variability in northeastern Amazon and northern northeast Brazil, depicted in Fig. 2g. REOF5 represents a nearly decadal variability in the middle-upper Paraná/La Plata basin (not specifically included in Fig. 2), which has been reported by Robertson and Mechoso (1998) in river flows.

As mentioned before, unrotated modes must be spatially orthogonal, but rotated modes are not subjected to this restriction. The comparison of both types of modes emphasizes some important features disclosed by rotation and the fact that the structures are simpler in the rotated modes, which are focused on more regional oscillations. For instance, the oscillations in the central-western and eastern parts of subtropical South America (Figs. 2a, c) are neither totally opposite (as suggested by EOF1, not shown) nor the same (as could be inferred from EOF2, not shown). The eastern and western parts have different variability but could share some oscillations, as can be seen from REOF1 and REOF2 (e.g., Compagnucci et al. 2002; Agosta and Compagnucci 2012; Penalba and Vargas 2004). One great difference between those two regions is that the western part shows change of phase in the mid-1970s (Figs. 2a, 5b), while the eastern part has a prominent change of phase in the 1960s (Figs. 2c, 5a).

The importance of the summer monsoon precipitation and its contribution to the variability of the annual precipitation is shown by the first rotated interdecadal variability mode of annual precipitation (Fig. 6). This mode is similar to REOF2 (Fig. 5b), although it also contains contributions from other modes and seasons. Why is the first annual mode more similar to REOF2 rather than to REOF1? The answer is in section 5. In the
next section, the presented modes are compared with modes obtained from longer series with more limited spatial coverage to verify their consistency.

4. Modes of interdecadal variability from long series

The results shown above cover a period of 50 years and, although providing a good spatial coverage, might reflect modes present only in this limited period. To see if they are present in longer periods, at least partially, the rotated EOF analysis was also carried out with the CRU data in the period 1900–93, over the limited coverage shown in Fig. 3b. This verification is only shown for summer, since the results are similar for spring.

It is known that unrotated modes are not stable when subportions of the domain are examined. However, this problem is greatly reduced by using rotated solutions (Richman 1986). A rotated mode should remain stable if the whole domain or only part of it is used, as long as strong factor loadings are situated in overlapping regions of those two domains. As this is the case in the present study, the analysis of the CRU data does provide additional reliability to the presented modes.

Amongst the first five rotated summer modes from the continental dataset (Fig. 5), REOF1, REOF2, and REOF4 are those with the highest factor loadings in regions reasonably covered by the CRU dataset (Fig. 3b). Consistently, they show the strongest correlations with CRU modes (Table 1).

REOF1 (Fig. 5a) cannot be completely reproduced with the CRU dataset, which does not cover central-east and northwest SA, where this mode has strong components. However, the first mode with CRU data (CRU REOF1; Fig. 7a) shows the strong components displayed by REOF1 in southeast SA, especially in the southern Paraná–La Plata basin (cf. Figs. 5a, 7a), and these modes are very significantly correlated (0.00 significance level) in the overlapping period (1954–1989) (Table 1).

REOF2 (Fig. 5b) is also not completely reproducible with the CRU dataset, which does not cover central SA. However, during the overlap period, it is very significantly correlated (0.00 significance level) with CRU REOF4 and CRU REOF5 (Figs. 7d,e; Table 1), which, like REOF2, have components over central-northwestern Argentina, extending toward central SA, and also show strong change of phase in the mid-1970s. Those CRU modes reproduce regional aspects of REOF2 (cf. Fig. 5b with Figs. 7d,e) and might appear mixed in the overlap period.

REOF4 (Fig. 5d), with strongest components over northeast SA, can be reasonably reproduced with the CRU data, as is shown by CRU REOF2 (Fig. 7b), which explains why these modes have the strongest correlation between modes obtained from the two datasets (Table 1).

CRU REOF3 (Fig. 7c) does not have one obvious counterpart in the first five summer continental modes. It represents features in southern SA present in more than one continental mode, as is suggested by its significant correlation with the REOF2 and REOF4 (Figs. 5b,d), which show similar features in that region.

Table 1. Correlation coefficients between the factor scores of summer REOFs calculated with the continental dataset and the CRU dataset. The highest correlation coefficients (levels of significance better than 0.01) are in bold.

<table>
<thead>
<tr>
<th></th>
<th>REOF1</th>
<th>REOF2</th>
<th>REOF3</th>
<th>REOF4</th>
<th>REOF5</th>
</tr>
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<tbody>
<tr>
<td>CRU REOF1</td>
<td>0.81</td>
<td>0.13</td>
<td>0.17</td>
<td>0.11</td>
<td>0.26</td>
</tr>
<tr>
<td>CRU REOF2</td>
<td>0.19</td>
<td>0.18</td>
<td>-0.19</td>
<td>0.86</td>
<td>-0.41</td>
</tr>
<tr>
<td>CRU REOF3</td>
<td>0.16</td>
<td>0.47</td>
<td>0.45</td>
<td>0.47</td>
<td>0.16</td>
</tr>
<tr>
<td>CRU REOF4</td>
<td>-0.12</td>
<td>-0.60</td>
<td>0.55</td>
<td>-0.15</td>
<td>0.49</td>
</tr>
<tr>
<td>CRU REOF5</td>
<td>0.37</td>
<td>0.77</td>
<td>0.02</td>
<td>-0.05</td>
<td>-0.01</td>
</tr>
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</table>
Fig. 7. As in Fig. 4, but for the first five summer CRU REOFs.
The comparison between the continental and the CRU modes indicates that if the CRU data cover the regions in which a continental mode has highest factor loadings, there will be a CRU mode with highest factor loadings in the same regions, and the factor scores of the corresponding modes are highly correlated (100% confidence level) (e.g., REOF4 and CRU REOF2). If CRU data are present in only part of the regions with strongest factor loadings of a continental mode, the features of the related CRU mode in this part are similar to those of the continental mode, and the corresponding factor scores have very significant correlation (100% confidence level) during the overlap period (e.g., REOF1 and CRU REOF1). This consistency confirms the robustness of the presented continental modes. Different CRU modes of higher order may be significantly correlated with a same continental mode over the shorter period, since they may appear mixed in a shorter period (e.g., CRU modes REOF4 and REOF5 and continental REOF2).

Most of the short-term tendencies in continental modes (1950–2000) are not confirmed in the longer interval (1900–93), but southeast SA shows increasing precipitation and increased amplitude of the interdecadal variability in part of the second half of the twentieth century (Fig. 7a). It is not possible to assess if this is part of a tendency to greater variance or a behavior restricted to a limited time interval, such as the increased amplitude during 1940–70 in northeast SA (Fig. 7b).

5. Relationships between spring and summer

Some features of the interdecadal variability in spring and summer are highly related, either in terms of inversion of the anomalies from one season to the next, or in terms of persistence.

The anomalies in central-east SA and in parts of southeast SA tend to have their signs inverted from spring to summer, especially in central-east SA. This is indicated by the high positive correlation between the factor scores of the continental REOFs of spring and summer: 0.81, almost 100% confidence level. These modes display a dipolar pattern between central-east and southeast SA but the polarity is inverted from spring to summer (Figs. 4a, 5a). This tendency to inversion, here disclosed for interdecadal time scales, is also observed on interannual time scales (Grimm and Zilli 2009) and is likely due to processes of surface–atmosphere interactions caused by precipitation anomalies over central-east SA in spring (Grimm et al. 2007). Grimm and Zilli (2009) noted an interdecadal modulation of that inverse relationship, since it is stronger in some periods. The inversion on interdecadal time scales may be simply the rectification of random sampling of the interannual variability, or the interdecadal variability may have an active role in the spring–summer inversion on interannual time scales by favoring spring anomalies that trigger the processes leading to inversion in summer. It seems that the latter alternative is more probable than the former, since the relationship on interdecadal time scales is even more significant than on interannual time scales (see Grimm and Zilli 2009).

Contrary to the tendency to reversal of anomalies in central-east SA, there are regions in which the anomalies in spring tend to persist throughout summer. The main example is central–northwestern Argentina and central SA, where the persistence is indicated by the high positive correlation between REOF4 in spring (not shown) and REOF2 in summer (0.59, 100% confidence level). Both modes have strong factor loadings of same sign in these regions.

There are also significant correlations between other rotated modes of spring and summer precipitation, although weaker than the two mentioned above. For instance, the positive correlation between spring REOF2 and summer REOF5, both with strong factor loadings of same sign in the middle-northern part of Paraná–La Plata basin, and the positive correlation between spring REOF3 (not shown) and the summer REOF4, both with strong factor loadings of same sign in northeast SA, indicate that interdecadal anomalies in these regions tend to persist from spring to summer.

The tendency to reversal of anomalies associated with the spring and summer REOF1s (Figs. 4a, 5a) explains why these seasonal modes are not the main contributors to the first interdecadal mode of annual precipitation (Fig. 6), although both of them present similar dipole-like patterns of factor loadings. As the anomalies tend to reverse from spring to summer, their contribution to the annual precipitation anomalies is smaller than the contribution from the summer REOF2 (Fig. 5b) and its associated spring REOF4 (not shown), the anomalies of which in central/northwestern Argentina and central SA have the same sign in spring and summer.

The interdecadal variability has great impact on seasonal precipitation totals and therefore on the annual cycle of precipitation. For example, the annual cycles of the precipitation in the region (18.75°–21.25°S, 38.75°–43.25°W) of the continental SACZ over central-east Brazil, averaged over the periods 1957–67 and 1968–77 [opposite phases of spring and summer REOF1s, (Figs. 4a, 5a)], show a clear shift (Fig. 8). This implies a shift of the rainy season and therefore in its onset and demise. If in one phase the spring is drier (wetter) than normal, the summer tends to be wetter (drier) than normal, consistent with the inverse relationship described in this section. Differences between precipitations during the opposite phases of REOF1 reach 50% in October and
January, and are significant in the spring months and in January and February, but not in December, when the reversal occurs.

6. Connections with SST and climatic indices

It is beyond the scope of this paper to detail the connections of the precipitation modes with SST and/or atmospheric circulation interdecadal variability. Notwithstanding, to prompt further studies on origins and mechanisms, and to show that the presented modes are related to known climatic interdecadal oscillations, statistical connections are established with SST anomalies and well-known climatic indices. The analysis uses differences between anomaly composites for SST in positive and negative phases of the precipitation modes [respectively years with factor scores above (below) 1 standard deviation]. Therefore, the anomalous SST signs described below refer to the positive phase of the precipitation modes.

a. Spring

For the spring REOF1, warm SST anomalies extend over the central eastern Pacific and North Atlantic. They straddle the equator in the eastern Pacific and extend poleward along the subtropical North American and South American west coast. Cold anomalies predominate over the western and central North and South Pacific (Fig. 9a). The pattern bears similarity to the warm phase of ENSO, except for the weaker signal on the eastern equatorial Pacific. It resembles the third interdecadal SST mode of Folland et al. (1999), the second mode of Parker et al. (2007), and mode 4 of Mestas-Nuñez and Enfield (1999). It is generally known as interdecadal Pacific oscillation (IPO). However there are differences with respect to the IPO found in previous studies in the southern Pacific, where the present anomalies are much stronger and extensive than in the northern Pacific. There are also differences in the North Atlantic, where the warm SST anomalies are stronger than in the IPO mode, indicating that this mode is not only connected to IPO, but also to a North Atlantic–related mode. This is confirmed by a significant correlation with the AMO index (Table 2). The ENSO-like anomalies in the Pacific are expected to enhance ENSO-like precipitation anomalies over SA, which is the case, as shown by the comparison of Fig. 4a with precipitation anomalies produced by ENSO in spring (Grimm 2003, 2004). The strong SST anomalies in the subtropical southern Pacific, related to the activity in the South Pacific convergence zone, are able to modulate the spring response to ENSO over southeastern SA (Barros and Silvestri 2002; Vera et al. 2004).

Before proceeding with the analysis of the connections with SST and climatic indices, some comments are necessary, using the above example of spring REOF1. Most of the precipitation modes are significantly correlated with more than one climatic mode or SST mode, since different precipitation patterns may result from different combinations of phases of those modes. For instance, the spring REOF1 seems to be associated with the positive phases of SST modes AMO and IPO. This is confirmed by its associated SST anomalies (Fig. 9a) and the correlation with the index AMO [+0.43; significance level (SL) = 0.00], but the correlation with the index IPO is not highly significant (+0.24; SL = 0.12) (Table 2). However, correlation with calculated SST REOFs (not shown) gives higher value with the IPO mode (positive; SL = 0.01) than with the AMO mode (positive; SL = 0.08). One possible reason for this discrepancy is that the IPO index used here is based on the IPO SST mode calculated by Parker et al. (2007), which has weaker components in the subtropical southern Pacific than the IPO SST mode we calculated, probably due to different SST data and domain and the fact that our mode is rotated, while Parker et al.’s is not. As mentioned before, the subtropical southern Pacific SST anomalies are important for the spring precipitation anomalies over southeastern SA.

The above comments prompt two cautionary remarks concerning connections between precipitation modes and climatic indices. First, if different modes show a strong correlation with an index, it does not mean that the SST anomaly patterns influencing these modes are the same. For instance, they might differ in the North Atlantic but, even so, contribute to high values of AMO_70 (SST averaged over the North Atlantic,
Besides, different definitions of AMO indices (such as AMO_70 and AMO_60) yield different correlation coefficients (Table 2). Second, the definition of an index associated with an SST mode (e.g., AMO, IPO, and PDO) does not always include the characteristics of this mode that are responsible for the teleconnection with the anomalous precipitation over South America. This is why the correlation of a precipitation mode with a climatic index may be different from the correlation with the associated SST mode, although the sign is always consistent. Therefore, and because some modes are predominantly associated with atmospheric climatic indices, the correlation analysis with both climatic indices and with SST is advisable.

The strongest SST connection with the spring REOF2 (Fig. 9b) looks IPO-like, although shifted eastward, and of opposite phase to that associated with REOF1 (Fig. 9a). The differences between these IPO-like patterns resemble the differences between the third and fourth SST REOFs of Mestas-Nuñez and Enfield (1999), describing interdecadal variability in the eastern and central tropical Pacific, respectively. In the Atlantic, the SST anomalies are of the same sign as for REOF1 (but stronger in the tropical North Atlantic), so there is the same (opposite) sign of SST anomalies in the eastern tropical Pacific and southwestern (northern) Atlantic, contrary to REOF1. Consistent with these SST anomalies, the largest correlations are with the indices AMO_60, AMO_70, TNA (positive), and IPO (negative) (Table 2). According to Mo and Berbery (2011), the positive SST anomalies in the northern tropical

![Fig. 9. Differences between anomaly composites for SST in positive and negative phases of the spring precipitation REOFs 1 and 2. Colors indicate the levels of confidence of the differences, with signs indicating positive or negative differences.](image-url)

**Table 2.** Correlation coefficients (and respective levels of significance within parentheses) between the factor scores of spring REOFs and climatic indices. Values significant at 0.05 or better are in bold.

<table>
<thead>
<tr>
<th>Spring Indices</th>
<th>REOF1</th>
<th>REOF2</th>
</tr>
</thead>
<tbody>
<tr>
<td>AMO_70</td>
<td>0.43 (0.00)</td>
<td>0.39 (0.01)</td>
</tr>
<tr>
<td>AMO_60</td>
<td>0.23 (0.13)</td>
<td>0.37 (0.01)</td>
</tr>
<tr>
<td>NAO</td>
<td>0.13 (0.41)</td>
<td>–0.14 (0.39)</td>
</tr>
<tr>
<td>IPO</td>
<td>0.24 (0.12)</td>
<td>–0.28 (0.07)</td>
</tr>
<tr>
<td>PDO</td>
<td>0.16 (0.31)</td>
<td>–0.19 (0.22)</td>
</tr>
<tr>
<td>TSA</td>
<td>0.21 (0.18)</td>
<td>–0.04 (0.88)</td>
</tr>
<tr>
<td>TNA</td>
<td>0.25 (0.11)</td>
<td>0.42 (0.01)</td>
</tr>
<tr>
<td>SAM</td>
<td>0.29 (0.06)</td>
<td>–0.26 (0.10)</td>
</tr>
<tr>
<td>NAM</td>
<td>–0.20 (0.21)</td>
<td>0.22 (0.16)</td>
</tr>
</tbody>
</table>
Atlantic during La Niña–like anomalies in the Pacific contribute to enhance negative precipitation anomalies northward of their usual La Niña–related position in southeast SA and reduce the associated positive anomalies in northern-northeastern SA. This is confirmed by comparison between REOF2 (Fig. 4b) and REOF1 (Fig. 4a, with opposite sign, as the IPO phase is opposite to that associated with REOF2).

### Summer

The SST anomaly patterns associated with the summer REOF1 (Fig. 10a) and with the spring REOF1 (Fig. 9a) are very similar, exhibiting connections to AMO and IPO with stronger (weaker) SST anomalies in the subtropical southern (northern) Pacific. This is coherent with the strong positive correlation between spring and summer REOF1s (section 5). The precipitation dipole between central-east and southeast SA is inverted from spring to summer (Figs. 4a, 5a) because of surface–atmosphere interactions and different teleconnections in spring and summer (Grimm et al. 2007). In spring, there are both tropics–extratropics and tropics–tropics teleconnection from the Pacific into SA, while in summer only tropics–tropics teleconnection is visible (Fig. 11). The same behavior is apparent during ENSO events (Grimm 2003, 2004; Cazes-Boezio et al. 2003).

The connection with AMO and IPO is confirmed by the significant positive correlation with the corresponding indices (and also TNA) (Table 3). The negative correlation with NAO and NAM, which are themselves related (Hurrell and Deser 2010), may be due to the nonstationary connection between NAO and AMO (Walter and Graf 2002).

The SST anomaly pattern for summer REOF2 (Fig. 10b) resembles that for spring REOF4 (not shown). Both modes are correlated (section 5) and show associated IPO-like patterns. Although both the summer REOF1 and REOF2 are related to IPO, there are differences between their associated SST patterns. The SST anomalies for REOF2 are weaker (stronger) in the subtropical southern (northern) Pacific, which is consistent with its stronger correlation with indices IPO and PDO (SL = 0.00; Table 3). In the Atlantic, they are of opposite sign, as shown by Figs. 10a,b and by the correlations with AMO, TNA, NAO, and NAM (Table 3). Besides, there are significant differences in the Indian Ocean. REOF2 corresponds to the tropical multidecadal mode (TMM) of Chelliah and Bell (2004), which represents variability of tropical convective rainfall and surface temperature and shows similar temporal evolution in the same period [with negative (positive) phase before (after) the mid-1970s] and, as REOF2, exhibits connections with SST anomalies in the North Atlantic, tropical Pacific, and Indian Ocean, as well as NAO and PDO.

The summer REOF3 does not show very significant associated SST anomalies over extensive areas (Fig. 10c) and has low correlation with the main SST-based climatic indices (AMO, IPO, and PDO; Table 3). It is not significantly correlated with any of the first six rotated SST modes (not shown). It is significantly correlated with NAO, which is consistent with the associated SST tripole pattern in the North Atlantic (Grossmann and Klotzbach 2009; Hurrell and Deser 2010) and with strong SLP anomalies in the North Atlantic (not shown). Its positive significant correlation with TSA (Table 3) reflects the usual relationship between below- (above-) normal precipitation in subtropical east Brazil and above- (below-) normal SST anomalies in the adjacent ocean during summer. In the Atlantic, near-equatorial enhanced (suppressed) convective activity occurs over positive (negative) SST anomalies, while off-equatorial cloudiness anomalies are negatively correlated with local SST anomaly (Tanimoto and Xie 2002).

For the summer REOF4, strongest connections resemble those of the spring REOF5 (not shown), as both have highest factor loadings in northeastern SA and are significantly correlated (section 5). These connections are with a strong meridional Atlantic SST gradient across the equator and with the IPO (with weak subtropical SST anomalies) (Fig. 10d). They are consistent with the significant correlations with indices TNA, TSA, AMO, and IPO (Table 3). As REOF 1, it has the same sign of correlation with IPO and AMO indices, although weaker (Table 3), but there are important differences between REOF4 and REOF1. REOF4 shows weaker associated SST anomalies in the subtropical southern Pacific (which explains its weaker factor loadings in southeastern SA), and, most importantly, it displays a dipole of significant SST anomalies in the northern and southern tropical Atlantic near the equator. It is the only mode with significant correlation with both TNA and TSA (Table 3). The negative (positive) precipitation anomalies in southeastern (northeastern) SA in the positive phase of REOF4 (Fig. 5d) are consistent with La Niña–like anomalies in the central east tropical Pacific and negative meridional SST gradient across the tropical Atlantic.

The SST anomalies associated with the summer REOF5 are relatively weak and scattered (Fig. 10e), and there is no significant correlation with the main SST-based climatic indices (AMO, IPO, and PDO; Table 3) or with any of the first six rotated SST modes (not shown). The highest correlation is with SAM (it is the strongest connection with this index; Table 3), which is coherent with its SLP distribution in high and mid-latitudes (not shown). The SST anomalies resemble those associated with spring REOF2 (Fig. 9b), though
FIG. 10. As in Fig. 9, but for the summer precipitation REOFs 1–5.
weaker. These two modes are positively correlated (section 5), and the spring REOF2 also holds negative correlation with SAM (although barely significant; SL = 0.10). The precipitation anomalies of summer REOF5 over the middle/upper Paraná–La Plata basin (Fig. 5e) are consistent with the late spring–summer impact of SAM on the precipitation in southeast SA (Silvestri and Vera 2003). Besides, the most significant SST anomalies (negative in the midlatitudes of the southwest Atlantic and Indian Oceans) are coherent with the negative phase of SAM, and the near-decadal time scale of variability of this mode (Fig. 5e) is coherent with a quasi-decadal variability in the SAM (8–16 years), reported by Yuan and Yonekura (2011). The two summer modes connected to SAM (REOF5 and REOF2) have associated SST anomalies in the midlatitudes of the southwest Atlantic and Indian Oceans of opposite signs, consistent with their opposite correlations with SAM.

The presented first five modes, with strongest factor loadings in different regions of high summer precipitation, are connected with different combinations of SST-based modes or atmospheric modes. There are not two modes with the same combination of signs of significant correlation coefficients.

7. Connections with previous regional results

Previous studies have reported interdecadal variability in some regions of SA. Some of these regional results are compared with the information of the continental-scale modes of interdecadal variability determined in the present study.

An important indicator of interdecadal oscillations of precipitation in western subtropical SA is the level of Laguna Mar Chiquita, whose catchment area is located in northwest Argentina (Fig. 12, left), where the rainy season is austral summer, with some contribution from spring and early autumn (Fig. 1 from Grimm 2011). Piovano et al. (2002) reported the lake-level variations for the period 1890–2000 by combining data such as salinities, shoreline positions and photographs with instrumental data of monthly lake-level elevation for the

| Table 3. Correlation coefficients (and respective levels of significance within parentheses) between summer REOF factor scores and climatic indices. Values significant at 0.05 or better are in bold. |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| Summer Indices | Precipitation modes | REOF1 | REOF2 | REOF3 | REOF4 | REOF5 |
| AMO_70 | 0.71 (0.00) | -0.24 (0.12) | 0.14 (0.37) | -0.31 (0.05) | 0.15 (0.36) |
| AMO_60 | 0.45 (0.00) | -0.50 (0.00) | 0.02 (0.89) | -0.23 (0.12) | 0.15 (0.36) |
| NAO | -0.35 (0.02) | 0.65 (0.00) | 0.34 (0.03) | -0.01 (0.94) | -0.21 (0.17) |
| IPO | 0.34 (0.02) | 0.70 (0.00) | -0.05 (0.76) | -0.32 (0.04) | -0.20 (0.21) |
| PDO | 0.30 (0.05) | 0.68 (0.00) | 0.22 (0.15) | 0.06 (0.72) | -0.10 (0.53) |
| TSA | -0.08 (0.61) | 0.25 (0.11) | 0.34 (0.03) | 0.30 (0.05) | -0.37 (0.02) |
| TNA | 0.43 (0.01) | -0.32 (0.04) | -0.18 (0.25) | -0.56 (0.00) | -0.08 (0.61) |
| SAM | 0.28 (0.08) | 0.40 (0.01) | 0.27 (0.08) | 0.16 (0.32) | -0.48 (0.00) |
| NAM | -0.42 (0.01) | 0.49 (0.00) | 0.12 (0.42) | 0.11 (0.46) | -0.26 (0.10) |

Fig. 11. Differences between anomaly composites for zonally asymmetric streamfunction at 200 hPa in positive and negative phases of the (a) spring and (b) summer precipitation REOF 1. Colors indicate the levels of confidence of the differences, with signs indicating positive or negative differences.
1967–2000 interval. Those variations are shown here for the period overlapping with our data (1950–2000) for comparison with the main rotated modes that show significant components over the catchment area (Figs. 12a–c): spring REOF1 (Fig. 4a), summer REOF1 (Fig. 5a), and summer REOF2 (Fig. 5b). Although the relationship between the first modes in spring and summer involves tendency to inversion of the anomalies in certain regions (as in central-east and southern Brazil), in the northernmost part of Argentina, there is persistence from spring to summer. The lake level underwent a strong increase in the 1970s (Figs. 12a–c, lower panels), and all these modes changed phase in this decade (Figs. 12a–c, upper panels), although in different years: spring REOF1 and summer REOF1 in the late 1970s (Figs. 12a,b), and summer REOF2 in mid-1970s (Fig. 12c). The main contribution to the variability of the lake level seems to come from the summer REOF2 for several reasons: i) it has strongest positive factor loadings all over the basin (Fig. 5b) and shows increase starting in the early 1970s, like the lake-level ascent (Fig. 12c); ii) in spring it rains less than in summer; iii) the summer REOF1 has opposite factor loadings in the northern part and southern part of the lake basin (Fig. 5a), and important changes of the lake level, such as its ascent in the early 1970s, do not coincide with REOF1 changes (Fig. 12b). Overall, there is good agreement between the modes of precipitation variability relevant to the Laguna Mar Chiquita catchment area and the lake-level oscillations.

Vargas et al. (2002) reported interdecadal oscillations and relationships between summer precipitation in northern Argentina, the upper Paraná River streamflow, and a circulation index associated with moisture advection to this region in spring–summer. The oscillations of these variables follow closely those of the first modes for spring and summer (Figs. 4a, 5a), with minima in the early 1950s and 1970s, and maxima around 1960 and 1980. Similar features were indicated for spring–summer precipitation in central-west Argentina by Compagnucci et al. (2002), but in this case there is also a noticeable contribution from the summer REOF2 (Fig. 5b). The great contribution of REOF2 in the region defined as central-west Argentina is also indicated by the comparison of Fig. 5b to the figure of the cumulative summer rainfall series of Agosta and Compagnucci (2012) for that region (their Fig. 2) and the same period and by their report of the great change of behavior of the precipitation in that region in the mid-1970s. This second mode is also related with the changes in the onset, demise, and duration of SA monsoon reported by Carvalho et al. (2011).

Minetti et al. (2003) pointed out a negative tendency of precipitation in southwestern SA from 1930 to the 1990s, which is coherent with the summer CRU REOF3 (Fig. 7c) as well as the continental summer REOF2 (Fig. 5b and also other continental-scale modes with factor loadings in that region).

The extremely dry conditions of the 1930s in the Argentinean pampas, which caused a large dust-bowl episode, as well as cattle mortality, crop failure, farmer bankruptcy, and rural migration (Viglizzo and Frank 2006), are visible in the CRU REOF1 (Fig. 7a). Also, the improved rainfall conditions, which favored the conversion of abandoned lands into grazing lands and croplands during the second half of the last century (Magrin et al. 2005), are clearly depicted by this CRU mode, which corresponds to the summer REOF1 in continental scale (Table 1).

Robertson and Mechoso (1998) reported a near-decadal oscillatory component (period ~9 years) in
the 1911–93 annual streamflows of the Paraguay and Paraná Rivers, measured at 27°S, which means that this component is contributed by rainfall in the middle and upper Paraná–La Plata River basin. This oscillation is strongest in the austral summer and cannot be compared with the CRU rainfall data, which are not available for such a long period in that catchment area. However, the continental-scale summer REOF5 (Fig. 5e), with strongest components over parts of the middle and upper Paraná–La Plata River basin, shows a clear oscillation with a period of about 9 years.

According to Hamilton et al. (1996, 2002), the level and extent of the Pantanal South American wetlands in central-west Brazil near Paraguay and Bolivia showed anomalously low values during the decade from 1960 to 1973. The mode with strongest factor loadings in this region in summer (rainy season) is REOF2 (Fig. 5b), which shows very low factor scores values during this period. Also the first mode of annual precipitation (Fig. 6) shows clearly that this was a very dry period in that region. Marengo (2004) reported that from the mid-1970s, the precipitation over the north (south) Amazon region decreased (increased) with respect to the pre-1970s, the precipitation over the north (south) Amazon (fourth); and the northeast Amazon (fifth). The first five rotated modes for spring, which jointly explain 64.0% of the variance, are focused on the variability shown in the second summer mode (Fig. 5b) and the first annual mode (Fig. 6), which display opposite variations in these regions, with a clear change of phase in the mid-1970s.

8. Summary and conclusions

Interdecadal variability is a substantial part of the summer precipitation variance over SA, amounting to more than 30% in extensive regions for which summer is the rainy season and winter is very dry. A comprehensive precipitation dataset has been prepared for the period 1950–2000, covering most of SA, in order to determine continental-scale modes of interdecadal variability. Rotated modes are presented for austral spring and summer seasons.

The first five rotated modes for spring, which jointly explain 64.0% of the variance, are focused on the variability over the following: central-east and southeast SA (first); northern part of the Paraná–La Plata basin (second); northeast and northwest SA (third); western part of the Paraná–La Plata basin and the northern Amazon (fourth); and the northeast Amazon (fifth). The first five rotated modes for summer jointly explain 66.4% of the variance. As the first spring mode, the summer REOF1 also displays a strong change of phase in the 1960s and a dipole-like pattern between central-east and southeast SA, but focused on the southern parts of these regions (SACZ and the southern Paraná–La Plata basin). It also features a dipole-like variability between central-east and northwest SA not apparent in the first spring mode. REOF2 is focused on central-northwest Argentina and the SA monsoon core region in central Brazil, where it displays anomalies opposite to those in northern Amazon and southern SA. It shows the clearest change of phase in the mid-1970s, sometimes called climatic shift. REOF3 displays a dipole-like pattern between central-east and southeast SA but, unlike the first mode, is focused on the northern parts of these regions. It also features a dipole-like variability between tropical east and west SA south of the equator and exhibits a nearly decadal time period. REOF4 concentrates on the variability over the northern Amazon and northeast SA, where summer is the beginning of the rainy season and has also a nearly decadal predominate period. REOF5 is strongly concentrated on a region south of the SACZ in the middle/northern Paraná–La Plata basin and has been responsible for a decadal-scale oscillation with a ~9-yr period in the streamflow of the Paraná River.

The REOF analysis carried out on CRU data with a longer period (1900–93) but less spatial coverage shows that if these data cover the regions with highest factor loadings of a continental mode, there is also a CRU mode with highest factor loadings in the same regions, and the two modes are highly correlated. If CRU data are present in only part of the regions with highest factor loadings of a continental mode, the related CRU mode is similar to the continental mode in this part, and they are also very significantly correlated. This shows the consistency of the modes obtained by different datasets and confirms the robustness of the continental modes. It is not possible to characterize continental interdecadal variability in SA with the long data series available, since the highest factor loadings of the main modes are in regions not covered by these long series (such as central SA and the SACZ). Therefore, to characterize continental-scale modes, it is necessary to assemble a comprehensive dataset, even if it spans only 50 years, and then validate these modes with longer time series and smaller spatial coverage.

The obtained continental modes show complete consistency with the results of previous regional studies and also represent adequately the interdecadal variability and covariability between interdecadal oscillations detected in different regions of SA with a great contribution of interdecadal variability and/or high precipitation variability in summer.

A noteworthy aspect of the results is the relationship between the interdecadal variability in spring and summer. The strongest correlation is between the first modes of spring and summer (Figs. 4a, 5a), indicating
that these two dipole-like modes tend to invert their polarity from spring to summer. Associated teleconnections to subtropical SA are strong in spring but weak in summer, and the SST anomalies that might produce remote forcing over the region do not generally change sign from spring to summer in interdecadal time scales. Therefore, it is probable that the remotely forced spring anomalies influence the summer anomalies through regional processes of surface–atmosphere interaction (Grimm et al. 2007) and that the remote forcing of the spring anomalies suitable to trigger the inversion process undergo interdecadal modulation.

Besides this inverse relationship between central-east and southeast SA, there are other relationships indicating persistence of anomalies from spring to summer, such as that between the spring REOF4 and the summer REOF2, both with similar factor loadings in central-northwest Argentina and central SA. This persistence explains why their contribution to the first mode of annual precipitation is greater than that from spring and summer REOF1s, which have almost opposite anomalies. These first modes impact strongly on precipitation regimes, with differences in monthly precipitation reaching 50% between opposite phases.

A preliminary analysis of the connections between the presented modes and well-known climatic indices and global SST anomalies indicates the physical basis for the presented modes. The spring and summer REOF1s are the result of the influence of both Pacific and Atlantic oceans, although further analysis is needed to identify the impact of each. Both modes display associated SST anomalies in the tropical central-eastern Pacific, with opposite anomalies in the subtropics, especially strong in the southern Pacific. In the North Atlantic, the anomalies have the same sign as in the tropical central-eastern Pacific, while in the South Atlantic they are predominantly opposite, especially in higher latitudes.

The spring REOF2 is also related to Pacific and Atlantic SST anomalies, but while the Atlantic anomalies have the same sign as the spring REOF1, the Pacific anomalies are opposite. Besides, there are differences in the patterns.

The summer REOF2 also displays connections to the Pacific and Atlantic, like the summer REOF1, but while the Pacific SST anomalies show the same sign for both modes, in the Atlantic they are opposite. Besides, it also displays connections with the Indian Ocean. Furthermore, contrary to REOF1, there are no strong SST anomalies in the subtropical southern Pacific. It has the strongest connection to the IPO index, while the REOF1 shows the strongest connection to the AMO index. It also best represents the climatic shift of the mid-1970s, suggesting that this shift might be due to a combination of changes of phase of several climatic modes, since it has the strongest correlations with most of the indices, especially with IPO and PDO (Table 3).

The summer REOF3 and REOF5 are not connected with any of the main SST modes, which is coherent with the weak SST anomalies associated with them. They have strongest connections with atmospheric circulation indices: NAO and SAM.

The summer REOF4, as the first mode, is connected to indices IPO and AMO with the same sign of correlation. However, unlike the summer REOF1, it does not show associated strong SST anomalies in the subtropical southern Pacific, which accounts for the weaker influence on southeastern SA. Most importantly, it shows significant opposite correlation with TSA and TNA and, therefore, with a strong meridional SST gradient across the equatorial Atlantic, which accounts for its strongest factor loadings over northeast SA.

The displayed connections with climatic indices and anomalous SST do not detail the physical mechanisms behind each mode and need to be cautiously analyzed, but they at least indicate that the modes have different dynamical origins and are, therefore, not simply artifacts of the rotated EOF analysis. Modeling tests will help with separating the influence of different oceans, but they need directions from observed relationships, such as those presented here.

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APPENDIX

Sources of Climatic Indices

AMO_70: [http://www.esrl.noaa.gov/psd/data/timeseries/AMO/](http://www.esrl.noaa.gov/psd/data/timeseries/AMO/)
AMO_60: [http://climexp.knmi.nl/data/amo_hadsst_ts.dat](http://climexp.knmi.nl/data/amo_hadsst_ts.dat)
IPO: [www.iges.org/c20c/IPO_v2.doc](http://www.iges.org/c20c/IPO_v2.doc)
REFERENCES


