LPB hydro-climate variability as simulated by GCM experiments: Role of remote SST forcing

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Manuscript submitted to

Climate Dynamics

July 20, 2012

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Abstract

A set of AMIP-type experiments is computed and analyzed to study springtime hydro-climate variability in the region of La Plata Basin (LPB). In particular, an ensemble of nine experiments with same interannually varying SST, as boundary forcing, and different initial conditions is used to investigate the relative role of the Pacific, Indian and Atlantic tropical oceans on modulating the local precipitation. The AMIP-type ensemble results have been compared with a coupled model experiment (using the same atmospheric component). The comparison reveals that the model has a good performance in the simulation of precipitation over LPB and South America, with a slight overestimation of the seasonal mean and an underestimation of the variability. Nevertheless, an EOF analysis of South America precipitation shows that the model is able to realistically reproduce the dominant modes of variability in spring. Further, its principal component (PC1) when correlated with global SST and atmospheric fields identifies the pattern related to ENSO and the large-scale connections. Overall the teleconnection pattern in the tropical and Southern Pacific Ocean is well captured by the SST-forced ensemble, but it is absent or too weak in other oceanic areas. In the subtropical South Atlantic the correlation is more realistic in the coupled model experiment suggesting the importance of air-sea feedbacks for that region, even at lower than interannual timescales. When the composite analysis of SST and atmospheric fields is done only over the ensemble members having a PC1 in agreement with the observations, both in terms of sign and intensity, then the correspondence between model and data is much improved. The improvement relies on avoiding climate noise by averaging only over members that are statistically similar and it suggests a high level of uncertainty due to internal atmospheric variability. Some individual springs have been analyzed as well. In particular, 1982 represents a clean case with a clear wave train propagating from the central Pacific and merging with a secondary one from eastern tropical South Indian Ocean, and it corresponds to a strong El Nino. Another case, 2003, corresponds to a rainy spring for SESA but in this case the en-
semble mean does not exhibit any teleconnection through the South Pacific and it is not able to reproduce the correct local precipitation pattern, suggesting that in this case regional effects are more important than remote forcing.
La Plata basin (LPB) is a region in South Eastern South America (SESA) comprising southern Brazil, Uruguay, northeastern Argentina, southern Paraguay and southern Bolivia that strongly relies on agriculture and hydro-electricity power. LPB region is a key area for the variability of the precipitation over South America having high values in all the seasons (see Zamboni et al., 2010). River discharge anomalies in SESA and analysis of precipitation regime over South America evidence low frequency variability (Robertson and Mechoso, 2000; Berbery and Barros, 2002; Rusticucci and Penalba, 2000), but its nature is not fully understood yet. Different hypothesis have been discussed in recent decades, like the decadal changes in the ENSO-SAM correlation (Fogt and Bromwich, 2006), a possible influence of Pacific decadal variability (Barreiro, 2010), including the role of the 1976/77 North Pacific climate shift (Huang et al., 2005), the impact of tropical Atlantic SSTA, as the tropical component of the Atlantic Multidecadal Oscillation (AMO Seager et al., 2010). Throughout most of the last century, SESA experienced a trend toward increased precipitation (e.g. Barros et al., 2008) but it is likely that anthropogenic climate forcing may explain only part of the wetting trend, as IPCC AR4 model simulations predict a weak increase in SESA precipitation over the last century (Seager et al., 2010).

At interannual timescales LPB precipitation has been linked to El Nino Southern Oscillation (ENSO) with a clear seasonality in the connection (Aceituno, 1988; Grimm et al., 2000; Paegle and Mo, 2002; Grimm, 2003; Cazes-Boezio et al., 2003; Vera et al., 2006; Barreiro, 2010, among others). El Nino influences SESA involving both upper and lower levels circulation anomalies: increased seasonal precipitation develops over LPB, while the northeast South America experiences drier conditions, and during La Nina the sign of the anomalies is reversed (Grimm et al., 2000). In the upper levels, Rossby wave trains propagating from the equatorial Pacific influence baroclinicity and advection of cyclonic vorticity over SESA (Yulaeva and Wallace, 1994; Grimm et al., 2000). In the lower levels, anomalous intensity and direction of the South American Low Level Jet (SALLJ) may change the moisture variability (Liebmann et al., 2004; Silvestri, 2004, i.e.). The season with the best established teleconnection between ENSO and LPB hydro-climate is austral spring (Cazes-Boezio et al., 2003; Barreiro, 2010; Zamboni et al., 2011). Spring is also
the season corresponding to the largest influence of the Southern Annular Mode (SAM) on LPB precipitation (Silvestri and Vera, 2009). SAM is the leading mode of variability in the Southern Hemisphere on low frequency. Positive (negative) SAM phase is associated with negative (positive) pressure anomalies over Antarctica and positive (negative) anomalies at middle latitudes (Thompson and Wallace, 2000). As a consequence of the SAM phase, decreased/increased precipitation over SESA is linked with weakened/enhanced moisture convergence associated with the anomaly of the upper level circulation over the southeastern Pacific Ocean (Silvestri and Vera, 2003).

The efforts to explain SESA precipitation variability have been based on both observations and model analysis. The comparison of the IPCC AR4 coupled model performance in simulating SESA precipitation and its variability reveal that they have problems in representing accurately the variability associated with the South Atlantic Convergence Zone (SACZ), but models having a good ENSO tend also to have a good teleconnection in South America (e.g. Silvestri and Vera, 2008; Vera and Silvestri, 2009). AGCM with forced SST have been tested as well analyzing precipitation and circulation biases over SA (Zhou and Lau, 2002), investigating the remote forcing from different ocean basins for predictability issues (Barreiro, 2010), or assessing their ability to reproduce the past history of SESA precipitation (Seager et al., 2010). Nevertheless the causes of the present interannual and lower timescale variability of SESA precipitation remain not fully explained.

Tropical oceans SST could affect the climate of South America through different mechanisms like Rossby wave trains that propagate into the extra-tropics and then into South America, affecting its eastern regions (i.e. Paegle and Mo, 2002; Vera et al., 2004); shift and alteration of the Walker circulation (Cazes-Boezio et al., 2003); influence of subtropical jets and inflow of humidity southward (Byerle and Paegle, 2002). In the present study we intend to investigate the influence from remote forcing (i.e. mainly SST) following the mechanisms just described. In particular, the model performance of atmosphere-only and atmosphere-ocean coupled experiments is investigated and compared in terms of hydro-climate variability over SESA at interannual and lower timescales. The analysis is mostly focused on austral spring when the connection between ENSO
and LPB precipitation is known and well established. A large ensemble of AMIP-type experiments with same boundary forcing (i.e. interannually varying SST) and different initial conditions is analyzed in terms of the correspondence between its PC1 and the data one. Some specific case studies have been considered as well.

The study is organized as follows: section 2 describes the model used and the experiments performed, including a list of the datasets and reanalysis used to verify the model performance. Section 3 is dedicated to the analysis of the hydro-climate variability (mainly in terms of precipitation variability) and its relationship with remote forcing. Section 4 investigates in more details the characteristics of the remote SST forcing over LPB precipitation, classifying years according to EOFs model performances. Section 5 lists and analyzes specific test cases. Finally, section 6 summarizes the main conclusions of the study.

2 Description of model experiments and datasets

Two kinds of experiments have been used for the present study: an ensemble of AMIP-type experiments and a 20th century coupled model simulation. The ensemble of AMIP-type experiments consists of 9 members with same boundary conditions, which are interannually varying SST taken from the HadISST dataset (Rayner et al., 2003), and different initial conditions. The period analyzed is 1948-2003. The experiments have been performed with the ECHAM4 atmospheric general circulation model (Roedeker et al., 1996) at T106 horizontal resolution (corresponding to a grid of $1^\circ \times 1^\circ$) and 19 sigma vertical levels.

The twentieth century coupled model simulation (hereafter SSXX) has been performed with the fully coupled atmosphere-ocean general circulation model SINTEXG (Gualdi et al., 2008). This includes prescribed concentration of greenhouse gases (i.e. CO$_2$, CH$_4$, N$_2$O and chlorine-fluorocarbons) and sulfate aerosols, as specified for the 20C3M experiment defined for the IPCC AR4 simulations (see http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php for more details). The characteristics of both atmospheric and oceanic model components are described in previous publications (Cherchi et al., 2008; Gualdi et al., 2008). The atmospheric component is the same used for the AMIP-type simulations. The oceanic component is OPA (Madec et al., 1998), which is
spatially distributed over a three-dimensional Arakawa-C-type grid (about $2^\circ \times 2^\circ$ horizontal resolution, with a meridional refinement of $0.5^\circ$ at the Equator, and 31 prescribed vertical levels).

The model outputs have been compared with observations and re-analysis data. In particular, the global distribution of sea surface temperature has been taken from the HadISST dataset (Rayner et al., 2003), atmospheric fields come from the NCEP reanalysis (Kalnay et al., 1996), and the global precipitation over land is taken from the CRU dataset (Mitchell and Jones, 2005). Satellite globally distributed precipitation for the period 1979-2003 from the CMAP dataset (Xie and Arkin, 1997) has been used for comparison with and validation of the land-precipitation dataset.

3 Simulated hydro-climate variability over South America

The analysis in this section focuses on the hydro-climate variability (mostly based on precipitation) over South America with emphasis on its southeastern part during austral spring (October, November, December mean; hereafter OND mean). Spring is chosen for our analysis because it has the largest teleconnection with ENSO (Grimm et al., 2000; Barreiro, 2010) and the largest correlation between observed and modeled LPB precipitation (not shown). Table 1 shows OND mean precipitation and its standard deviation averaged over South America and over LPB for the CRU dataset and for the model outputs. In the AMIP-type ensemble, the computation is applied to all the members as a single timeserie. In the LPB region the model simulates a larger than observed amount of precipitation, but its standard deviation is smaller (Table 1). That is to say that the model tends to underestimate the variability of the precipitation over SESA, even if it tends to overestimates its total amount.

In the literature precipitation variability over SESA, and over the LPB region in particular, has been measured by rainfall indices defined as averages over specific region (Boulanger et al., 2005; Vera and Silvestri, 2009; Barreiro, 2010, among others). Following Barreiro (2010), we define an LPB index as the averaged precipitation in the region $37^\circ S-19^\circ S$, $65^\circ W-47^\circ W$ (over land points only). The region corresponds to the same area used to compute the LPB values in Table 1. In the AMIP-type ensemble mean that index, averaged in austral spring (OND), is significantly correlated with the analogue computed from CRU data (the correlation coefficient is 0.56), suggesting that
in this season the role of the forcing from oceanic SST is large. When the LPB index is filtered
using a 7-years filter to keep the frequencies lower than 7 years (i.e. lower than interannual)
the correlation is still significant and large (i.e. the value is 0.46). Even if in both cases a large
component of internal variability remains (see Zamboni et al., 2011, for the interannual timescale)
our study intends to focus on the forced one.

Because of the model weakness in simulating the intensity of precipitation standard deviation,
we decided to identify an index based on the EOFs of the precipitation anomalies over South
America. EOFs and PCs allow identifying the dominant modes of variability avoiding the in-
consistencies between model and observations in the geographical differences. In past literature
the dominant mode of variability of the precipitation over South America have been investigated
using different datasets. Due to the sparse distribution of the observations in many regions of
South America, gridded datasets, as e.g. the CRU dataset, cannot represent correctly its precip-
itation (Stuck et al., 2006; Carril et al., 2012), and the global precipitation coverage taken from
satellite measurements after 1979 (CMAP) is eventually more reliable. Nevertheless, over LPB
region mean fields and variances of CRU and CMAP are very similar (Boulanger et al., 2005). We
strengthen this result by comparing EOF patterns obtained from the two datasets.

Fig. 1 shows the first three EOFs of land precipitation over South America (between 45S and
the Equator) during spring for the CRU and CMAP dataset during the overlapping period (1979-
2005). It is possible to identify similar spatial patterns between them (fig. 1), in particular the
north-south dipole depicted by the first mode and the triple pattern defining the third mode. In
terms of temporal variability the PCs corresponding to the three leading EOFs are compared via
correlation analysis (see Table 2): the correlation coefficients are large and statistical significant
for the same PC (the diagonal in Table 2). These results suggest that in both datasets the modes
are well separated. The results just discussed give us confidence in continuing the investigation
using the CRU dataset, using the 50 years available to validate and compare the model results in
the period 1948-2003.

Fig. 2 shows the first three modes of variability of South America precipitation during OND, and
its principal components for the CRU dataset, but considering the long time record (1948-2003).
The dominant mode of variability is a north-south dipole with centers at 15S and 30S (fig. 2a) in the eastern part of the continent. Its first principal component (PC1, fig. 2b) corresponds to the variability of ENSO, as the correlation coefficient between PC1 and NINO3.4 (monthly mean SST anomalies averaged in the box 5S-5N, 170W-120W) is 0.59. The second pattern has three poles centered at 5S, 20S and 35S (fig. 2c), and the associated PC2 (fig. 2d) corresponds to decadal timescale variability. Finally, the third mode is an east-west dipole with centers between 15-20S (fig. 2e) and its PC (fig. 2f) corresponds to a trend, at least in the last part of the record.

The comparison between the spatial patterns in fig. 2 and fig. 1 evidences that changing the time record length the first mode is unchanged, while the second and third one seem to be inverted. The difference could be related with the modulation of the decadal variability of LPB precipitation associated with the Southern Hemisphere climate, as discussed by Silvestri and Vera (2009).

The performance of the model in reproducing the dominant modes of variability of the precipitation over South America is shown in fig. 3. In terms of spatial patterns the model is able to realistically represent the dominant modes of variability of the precipitation over South America. In fact, the first mode produced by the AMIP-type experiments is a north-south dipole with centers over the LPB region and over the northeast part of the continent (fig. 3a). As second and third mode the patterns have centers forming a triple (fig. 3b) and an east-west dipole (not shown), respectively. In the AMIP-type ensemble the EOFs are computed over all the members concatenated to form a long record. Even in the coupled model the spatial patterns of the first two modes of variability are realistic (not shown). Table 3 summarizes the relationship between the principal components in the AMIP-type ensemble and ENSO: both PC1 and PC2 are significantly correlated with ENSO (i.e. NINO3.4 index). The latter is found for the AMIP-type experiment and not in the corresponding analysis using the coupled model experiment data.

On the base of these results we select the OND PC1 as index of precipitation variability over LPB. In particular, positive (negative) values correspond to wet (dry) conditions over LPB and dry (wet) ones to the north, following the intensity of the SACZ (Paegle and Mo, 2002; Silva et al., 2009; Liebmann et al., 2004). The correspondence between precipitation variability in LPB and remote SST during spring is shown in figure 4. The correlation coefficients between PC1 and
global SST identify the patterns related to ENSO and its teleconnections (fig. 4a). In fact, wet (dry) LPB years are related to positive (negative) SST anomalies in the Tropical Pacific and Indian Oceans, eastern equatorial Atlantic, subtropical South Atlantic near the South American coast and south eastern Pacific (with a max at 50S), and with negative (positive) SST anomalies in the central North Pacific, subtropical south western Pacific and southwestern South Atlantic (see also Paegle and Mo, 2002; Seager et al., 2010). The ENSO-LPB precipitation teleconnection is strong and robust in the AMIP-type ensemble, even if with some biases. In particular, the pattern in the Pacific Ocean (from the tropical sector to its southern part) is well represented in the ensemble in agreement with the idea of its strong forced influence (fig. 4b). Concerning the other oceanic sectors, the teleconnection is absent in the North Pacific and in the sub-tropical South Atlantic and it is weaker in the Indian sector (fig. 4b).

It is now well recognized that some patterns of SST variability result from a combination of atmospheric and oceanic processes (Deser et al., 2010). In the Indian Ocean the weakness in the model may be ascribed to the missed coupling in this set of experiments that is known to be important for the region (Bracco et al., 2007). The lack of the ocean-atmosphere coupling may be an explanation also for the performance in the subtropical South Atlantic, in agreement with Barreiro (2010). In fact, when the same analysis is conducted using the coupled model data the correlation in that region is more realistic (fig. 4c). Over the Indian Ocean the bias is not improved despite the presence of the ocean-atmosphere feedback (the values are in fact weaker than observed even in fig. 4c). Moreover, in the tropical Pacific Ocean the correlation tends to extend for the whole basin (up to the western edge) and this is consistent with the well-known biases of the model in the ENSO representation (Navarra et al., 2008). In the North Pacific the model misses the right connection with ENSO as it is wrongly triggered by model biases and air-sea coupling influences (Cherchi et al., 2011).

Considering timescales lower than interannual, we applied a 7 years low pass filter to the PC1 and then we computed the time-correlation with the SST. The results indicate that at low frequencies there are no indication of relevant patterns identifying a link between SST and precipitation in LPB (fig. 5). In the observations, positive significant values are found in the south Pacific sector.
(50S, 150W) and in the tropical north Atlantic (fig. 5a). In the model the positive correlation in the south Pacific sector is captured, while the signal in the north Atlantic sector is missing (fig. 5b). Additionally, in the AMIP-type ensemble there is a spurious negative correlation in the western sectors of both Pacific and Equatorial Atlantic Oceans (fig. 5b). On the other hand in the coupled model the correlation tends to be weaker everywhere (fig. 5c). Overall, the model has difficulties in capturing variability at frequency lower than interannual. In the case of the AMIP-type ensemble this may suggest that the internal variability could represent an important component of these modes.

Another remote forcing for the precipitation in the LPB region comes from the Southern Annular Mode (SAM; Thompson and Wallace, 2000). SAM depends on the interactions between the tropospheric jet stream and extratropical weather systems, being a source of uncertainty for the simulated climate at mid- to high southern latitudes (Deser et al., 2012). It has signatures in the tropospheric circulation and midlatitudes storm tracks and therefore could affect weather in southern South America (Menéndez and Carril, 2010), but its influence on SESA precipitation was not stable over recent decades (Silvestri and Vera, 2009). For the observations, our SAM index is based on in situ sea level pressure (Marshall, 2003) and it is taken from http://www.nerc-bas.ac.uk/icd/gjma/sam.html as OND mean. For the model it is defined as difference of zonal mean sea level pressure between 40S and 65S, following Marshall (2003). Fig. 6 shows a sliding correlation on 19 years between this index and the LPB precipitation index defined before (i.e. in Table 1). The magenta solid line of fig. 6 indicates that the correlation between the two indices is non-significant up to mid-70s, and then it becomes significant and negative. In fact, the time regression of the SAM index and the precipitation has a strong dipole signature only after 1979, while before the signal is weak (not shown).

In the AMIP-type ensemble the same analysis is conducted but considering the correlation of the ensemble mean (fig. 6, black solid line) and the mean of the correlation in each member (fig. 6, dashed black line). The distinction is motivated by the idea of identifying and discussing the forced and internal variability components. The correlation of the ensemble mean shows a change to a negative and significant correlation after mid-70s, suggesting that the change contain a component
that is forced by the SST. On the other hand, the mean of the correlation for each member is never significant suggesting that either the model is not able to reproduce the correct internal variability, or that its role is not dominant in this case. Considering the ensemble mean, a weakening of the time regression pattern between SAM and precipitation in the LPB region is simulated even if it is not as pronounced as observed (not shown).

4 Remote forcing on LPB precipitation variability

We classified wet (dry) LPB years using the first principal component (PC1) of precipitation anomalies over South America (as defined in the previous section), choosing 1 (-1) standard deviation as threshold. Wet (dry) LPB years correspond to precipitation dipoles with excess (deficit) of precipitation over LPB and the reverse north of it, respectively (fig. 3a).

Fig. 7 shows composite of SST and 200 mb eddy streamfunction for wet and dry LPB years computed for the HadISST/NCEP datasets and for the AMIP-type ensemble. In the observations, the SST anomalies are almost symmetric between the two phases in the tropical and south extratropical Pacific, but not in the extra-tropical North Pacific, in the Indian Ocean and in the subtropical South Atlantic sectors (fig. 7a,b). In particular, wet LPB years are associated with positive SST anomalies remarkably large over the eastern tropical Pacific (Zhou and Lau, 2001; Paegle and Mo, 2002; Seager et al., 2010), but also significant over the western tropical Indian Ocean and the eastern tropical Atlantic Ocean (i.e. the Gulf of Guinea). Furthermore, wet LPB years have negative SST anomalies in the western subtropical south Pacific, and partially in the eastern side of the Indian Ocean. The SST patterns with warm tropical Pacific and cold subtropical South Pacific enhances the convection in the SPCZ southeastward into the subtropical regions, and intensifies localized overturning cells associated with an anomalous Rossby wave source in the central south Pacific convergence zone (Vera et al., 2004). During dry LPB years SST anomalies of opposite sign are found over tropical southern Pacific and in the Indian Ocean, although the latter are less pronounced. Similar anomalies to those corresponding to the wet years are observed over the extratropical North Pacific while negative anomalies exist in the subtropical South Atlantic, off the South American coast (fig. 7a,b).
In the AMIP-type ensemble, when the composite is built considering the PC1 of all members, SST anomalies are realistic in the tropical and southern Pacific Ocean but not in the North Pacific, Tropical Indian and Atlantic sectors (fig. 7c,d). In particular, the asymmetry between strong and weak LPB years in the North Pacific is not simulated, and the anomalies in the Indian sector are largely weaker than observed.

In the observations, in both wet and dry LPB years, there is a wave train propagating from the western Pacific/Indian sector, as depicted by the 200 mb eddy streamfunction (fig. 7a,b contours), it recalls the Pacific South American Mode (Mo and Paegle, 2001; Chan et al., 2008; Taschetto and Ambrizzi, 2012). These patterns are remarkably symmetric in the Southern Hemisphere, while the signal north of the Equator in the Asian sector is larger during wet LPB years (fig. 7a,b). During wet LPB years the intensities are larger in the starting propagating phase, but in the dry LPB years the positive streamfunction anomalies over SESA and adjacent Atlantic sector are more intense. Vera et al. (2004) and Zamboni et al. (2012) proposed that these may originate from the central Pacific, but the amplification may also arise from local processes over SESA.

In the model anomalies recalling the PSA can be identified (fig. 7c,d) with the contribution from the central Pacific being more evident. During dry LPB years, in the coupled model SST anomalies in the subtropical south Atlantic, off the South American coast, are negative as in the observations, suggesting that the ocean-atmosphere coupling for this sector is important (not shown).

In the AMIP-type ensemble, the model PC1 for each member may peak in the same years as in the observations but also in others. We found that composing years according to the correspondence between AMIP-type PC1 and observed PC1 provides some extra information on the origin of the remote SST forcing for the LPB hydro-climate. We have classified years as "In Phase" when the model PC1 exceed 1 standard deviation (std) as in the observations, as "Out of Phase" when the model PC1 is of opposite sign as the observed one and as "Partial In Phase" when the model PC1 has the same sign as the observed one but the former does not exceed 1 std. During the "Out of Phase" years, the precipitation composites have excess (deficit) of precipitation in the region north of the La Plata Basin without any clear signal in the southern part of the dipole (not shown). During "In Phase" years the dipole structure is the strongest, while it is almost destroyed during
"Partial in Phase" years. However, in the latter case the anomalies over LPB, even if weaker, are of the same sign as in the "In Phase" group (not shown).

It is instructive to discuss the composite of SST and 200 mb eddy streamfunction built using the classification just introduced (fig. 8). The "In Phase" composite of SST (fig. 8a,b) reflects the observed anomalies (fig. 7a,b). In fact, during wet LPB years positive anomalies in the central eastern Pacific are associated with positive anomalies in the Indian sector and negative anomalies in the subtropical south Pacific. On the other hand, during dry LPB years negative anomalies in the central eastern Pacific are associated with positive anomalies in the North Pacific and negative anomalies in the South Atlantic (around 30S), off the SA coast. A main difference exists in the subtropical Indian sector where the In Phase composite has anomalies near zero and not negative as in the observations.

The comparison between fig. 7 and fig. 8 suggest that the forcing from SSTA in the tropical Pacific Ocean may provide both wet and dry conditions in the LPB region and that the forcing from other basins, like the Atlantic and the Indian Oceans, may trigger the teleconnection with the Pacific. In fact, in the "Out of Phase" composite (fig. 8c,d) the SST anomalies in the tropical Pacific and Indian Ocean sectors corresponds to the "In Phase" composite, while the anomalies in the North and subtropical South Pacific and in the Atlantic region largely differ. On the other hand, the main differences between "In Phase" and "Partial In Phase" composites in terms of SST are localized in the Indian Ocean (fig. 8e,f). The comparison between "In Phase" and "Out of Phase" composites suggests that an SST pattern with negative (positive) anomalies south west of large positive (negative) anomalies in the tropical Pacific Ocean may have a dominant role in terms of the wave propagation of the atmospheric teleconnection from the Indian-Pacific sector to South America (Vera et al., 2004; Zamboni et al., 2012).

Similarly to the SST, the wave propagation in the "In Phase" composite corresponds to the observations (fig. 8a,b) for both wet and dry LPB years with a clear wave train propagating from west to east. In the positive case, positive SST anomalies in the central eastern Tropical Pacific are associated with positive anomalies in the Indian sector and with negative anomalies in the subtropical south Pacific in correspondence of the dateline. Both the conditions are present only
when the "In Phase" positive cases are considered, and they seem both responsible of the wave
train. In the negative case, the HadISST composite has an SST pattern just opposite, but with
the negative anomalies in the Indian Ocean weaker (considering their absolute values) than in
the positive case. Further, in the "In Phase" negative composite the values in the Indian Ocean
are near zero. Actually in this last case the wave train seems to propagate from Indonesia rather
than from the eastern Indian sector as in the positive case. However we may not either exclude
the possibility that they correspond to a positive interference of two wave trains (Zamboni et al.,
2012). Further comparing fig. 7 and fig. 8 evidences that the "Out of Phase" composite does not
have any propagating signals (fig. 8c,d), while the "Partial in Phase" one has a wave train weaker.
Further, it seems to propagate from the Indian (central Pacific) sector in the positive (negative)
case (fig. 8e,f).

Shift and alteration of the Walker circulation associated with SSTA in the tropical oceans di-
rectly affect tropical South America (Cazes-Boezio et al., 2003). During warm ENSO events the
Walker circulation shifts eastward and its subsiding branch occurs over South America. When the
dipole is in its positive phase, i.e. rainy SESA and dry Amazon, subsidence over the Amazon is
particularly evident for the "In Phase" composite, as shown from its mean vertical velocity (see
Table 4). Conversely, when the dipole is in its negative phase the vertical velocity anomaly has
the opposite sign, favoring convection over northern South America (Table 4). In terms of local
processes over South America, the "In Phase" composite of vertically (from the surface to about
200 mb) integrated moisture shows a well-defined dipole between SESA and Amazon (fig. 9a,b).
The anomalies are symmetric comparing positive and negative phases, and the moisture fluxes are
directed north-easterly (south-westerly) in correspondence of positive (negative) moisture anoma-
lies (fig. 9a,b). The "Out of Phase" composite shows fluxes directed in the opposite direction and
the moisture anomalies over SESA are absent or extremely weak (fig. 9c,d). In the latter case,
anomalies of sign opposite to the "In Phase" composite are large in the northern part of South
America (fig. 9c,d).

During warm ENSO events, stronger upper tropospheric subtropical westerlies over the Andes
correlate with an eastward and southward humidity flow emanating from the Amazon basin to-
ward LPB (Byerle and Paegle, 2002). Regarding the anomalies in the westerlies over the Andes near the position of the subtropical jet (last column of Table 4), stronger upper tropospheric, subtropical westerlies over the Andes (e.g. during warm ENSO events) correlate with an eastward and southward humidity flow emanating from the Amazon basin toward LPB (Byerle and Paegle, 2002). Indeed, the positive phase of the dipole (rainy SESA) coincides with an intensification of the westerlies, suggesting a larger moisture supply from the northwest through the low-level jet. The weakening of the subtropical jet in the negative phase suggests opposing processes.

5 Some case studies

Table 5 summarizes the results from the AMIP-type ensemble in terms of the classification into "In Phase" and "Out of Phase" groups. In particular it contains the number of members of the ensemble in agreement with CRU results, as a measure of the inter-ensemble spread. For positive PC1 cases (i.e. wet LPB years), 1982 and 2003 represent two interesting cases worth of further investigation. In particular, 1982 is the only case having all the nine members reproducing the observed result: 100% of the members have a positive PC1 exceeding 1 standard deviation as in the observations. This case could be interpreted as the clean example of remote SST influence; moreover it corresponds to one of the strongest El Nino years in the analyzed record. Year 2003 is characterized by having 8 members over 9 with a negative large PC1 (exceeding -1 standard deviation) rather than a positive large one as in the CRU dataset.

As we mentioned in section 3, in the observations the correlation between PC1 and NINO3.4 is significant. When we consider OND SA precipitation PC1 years exceeding 1 (-1) std, 3 over 7 (2 over 8) wet (dry) LPB years correspond to El Nino (La Nina) event. This means that only 3(2) over 7(8) extreme wet (dry) LPB years occurred in correspondence of an El Nino (La Nina) year. However, as the teleconnection from the Pacific to the South America is almost simultaneous, the SSTA in October-November may have large impact even if they may not develop into ENSO years (Zamboni et al., 2012).

For negative PC1 cases (i.e. dry LPB years) it is hard to identify a net common behavior among the members. We decided to focus on 1999 because in the model it has 4 members "In Phase" and
5 members "Partial In Phase" with CRU results, and it corresponds to a La Nina year.

5.1 1982 case study

Year 1982 is the only case in our record having all the members with a PC1 larger than 1 std as in the observations: as all the members agree, we expect that the LPB precipitation pattern is completely forced from remote SST distribution. Fig. 10b shows the precipitation pattern in the AMIP-type ensemble composite (merging all 1982 years) with a well-defined dipole with excess of precipitation over LPB and deficit north of it, associated with a low-level convergence. In terms of SST, 1982 represents one of the strongest El Nino in the recent record with large positive SSTA in the eastern tropical Pacific Ocean (fig. 10a). During that year, positive SSTA in the tropical Pacific region are associated with negative SSTA in the subtropical Pacific around the dateline (both north and South of the Equator), weak positive anomalies in the Western Indian Ocean and negative anomalies in the Equatorial and subtropical South Atlantic. In the model, this SST pattern produces a clean wave train propagating from the central Pacific and merging with a secondary one from eastern tropical South Indian Ocean (fig. 10a). Over South America, the 200 mb streamfunction north of 20S is positive, in agreement with the observations (not shown), while it is negative (positive) south of it in the east (west), differently from the observations (not shown). In this case, the simulated precipitation anomalies over LPB can be interpreted as a net consequence of the teleconnection from the Pacific-Indian Ocean sectors.

5.2 2003 case study

During 2003, considering the reanalysis datasets, wet conditions over LPB are basically explained by a wave 3 configuration (fig. 11a), rather than from a teleconnection from the Pacific. 2003 is not an El Nino year and it experienced SST anomalies generally warmer than the mean climatology in all the tropical basins (fig. 11a,c). In this case the dipole precipitation pattern over South America seems to be mostly related to local effects rather than to remote SST forcing.

In the AMIP-type ensemble, the wave-3 pattern is not simulated (fig. 11c, contours) and the 200 mb streamfunction standard deviation ((fig. 11d) shows large spread among members in the Southern part of South America. Over LPB, as indicated by the sign of PC1, precipitation is
mostly negative (fig. 11d). Large positive SSTA, mainly in the Atlantic, are associated with intense precipitation north of 30S (fig. 11d): large positive precipitation anomalies over Amazon and slightly negative values over LPB provide a negative PC1, i.e. in this case "Out of Phase" than observed. Eddy streamfunction at 200 mb is highly variable among the members and only one over nine is able to reproduce the wave-3 pattern (fig. 12). In this case the internal variability and the associated spread among the ensemble members dominate the model simulation of hydro-climate over LPB and the remote SST forcing is less efficient.

5.3 1999 case study

Year 1999 experienced a strong La Nina with large negative SSTA in the tropical central and eastern Pacific Ocean (fig. 13a,c). Over South America dry (wet) conditions occur over LPB (north of it) (fig. 13b). In the model ensemble, four and five members have a PC1 "In Phase" and "Partial In Phase", respectively, with the CRU PC1. The model composite of these members shows negative precipitation over LPB and positive anomalies north of 20S (fig. 13d), but the values are weaker than observed (fig. 13b). In terms of teleconnection patterns, positive streamfunction anomalies in the upper troposphere are related with a quadruple between western Pacific and American continent sectors (fig. 13a). In the model the anomalies over SESA are largely weaker than observed (fig. 13c) and to verify the inter-ensemble performance we consider all the members separated.

Fig. 14 shows precipitation and 200 mb eddy streamfunction anomalies during OND for each member and for the ensemble mean. The members with a clear dipole with negative precipitation anomalies over LPB have positive 200 mb eddy streamfunction anomalies over SESA associated with a quadruple between western Pacific and American continent sectors (fig. 14b, g,h,i). In these cases an internal variability component dominates (fig. 13e), as the SST pattern is the same also for the other members.
6 Conclusions

We studied the influence from tropical SST anomalies on the precipitation variability over SESA at interannual and at lower than interannual timescales. The focus has been placed on the evaluation and analysis of ensemble continuous 1948-2003 integrations performed with an atmospheric GCM with relatively high horizontal resolution (1x1). It was also used a coupled global model to explore the potential importance of ocean-atmosphere interaction. We focus on the study of the austral spring motivated by the fact that during this season the signal from the tropical Pacific is more robust (e.g. Grimm et al., 2000; Barreiro, 2010).

Relatively high level of uncertainty in the observations characterizes large areas of South America (Carril et al., 2012). However, our results indicate that the regional climate modes of variability calculated from two independent precipitation databases (CRU and CMAP) for 1979-2003 are similar. Therefore we assume that CRU precipitation for the period 1948-2003 is realistic enough and we can use it for the purpose of model evaluation of broad scale regional modes of variability during austral spring.

In terms of seasonal area mean precipitation over SESA and its standard deviation, both atmospheric and coupled models made a good performance, with the seasonal mean slightly overestimated and its variability somewhat underestimated. The analogue computed from observations is significantly correlated with the values obtained from the SST-forced ensemble, suggesting that oceans influence the precipitation over SESA.

Both the atmospheric and the coupled models realistically capture the spatial patterns of the two dominant precipitation modes of variability over South America. The first mode is a south-north precipitation anomaly dipole with centers over SESA and central-northern Brazil and its principal component is used as a climate index of precipitation interannual variability (PC1). Its correlation with global SST identifies the patterns related to ENSO and its teleconnections. The teleconnection pattern in the southern Pacific Ocean is well captured in the SST-forced ensemble, but it is absent or too weak in other oceanic areas. The correlation in the subtropical South Atlantic is more realistic in the coupled model experiment, suggesting that air-sea feedbacks would be important there. These correlations with the SST were also calculated for time scales lower than interan-
nual for which the filtered (7-year low-pass filter) first PC of precipitation over South America has
been used as an index. The atmospheric model tends to capture qualitatively well the pattern of
lower-frequency teleconnections over vast areas of the Pacific, but tends to fail over the Atlantic.
Correlations tend to be too weak in the coupled model. However, its performance is qualitatively
acceptable in the south Atlantic again suggesting the potential importance of ocean-atmosphere
interactions in this sector.

The analysis of the relationship between SAM and LPB hydro-climate reveals that even though
the annular mode is associated with internal atmospheric variability, there is hint to a possible
oceanic influence on SAM variability on decadal timescales. The long-term time evolution of the
correlation between the SAM index and the precipitation in SESA was evaluated in a way to dis-
tinguish the forced variability (correlation of the ensemble mean) from the internal atmospheric
variability or climate noise (the average correlation of each member). During the last decades
the positive phase of the SAM is associated with decreased precipitation over SESA for the ob-
servations (consistent with Silvestri and Vera, 2009). In particular, we found that the SST-forced
variability resembles the evolution of the observed correlations, while the internal variability does
not, suggesting a potential for the SST anomalies to influence on the spatial circulation anomaly
patterns typically associated with the SAM.

Composite fields of upper-tropospheric streamfunction anomalies averaged over all the wet
springs in SESA consist in wave trains extending southeastward from eastern Indian Ocean and
Indonesia before they turn equatorward into South America. The dry composite is almost sym-
metric in the tropical and southern Pacific. These wave trains share some elements of the second
and third leading modes of SH circulation variability on interannual time scales (e.g. Mo, 2000,
the first leading mode of SH circulation variability is the SAM). The SST-forced ensemble cap-
tures the circulation anomalies and also those of SST in the tropical and southern Pacific, but the
anomalies are of lower amplitude than observed. Interestingly, the atmospheric model does not
capture the cold anomaly in the subtropical South Atlantic for the dry composite, but the coupled
model does, further confirming the importance of ocean-atmosphere interaction in this sector.

If the composite for wet/dry events is done by averaging only over those ensemble members
for whom the model and observations agree regarding the occurrence of strong positive/negative precipitation anomalies ("In Phase" composite), then the structure of teleconnections corresponds better with the observed. This improvement arises from avoiding the climate noise by averaging only over members that are statistically similar on the basis of the principal component of the leading precipitation mode over South America. The SST anomaly in the Indian Ocean (correctly captured in the "In Phase" composite) seems to be a factor to take into account since it is in this sector where the wave train influencing the precipitation dipole in South America originates. The pathway from the tropical Indian Ocean would be particularly important in spring as in this season there is strong covariation of ENSO and the Indian Ocean dipole (Cai et al., 2011).

We analyzed some individual springs for which the number of ensemble members in agreement with observations was very different (regarding the first leading mode of precipitation variability in South America). In the 1982 spring all members coincide in the rainfall anomaly dipole as observed. In this case the enhanced northerly low-level flow and moisture transport to the east of the Andes feed convection over SESA. The associated SST pattern produces a clean wave train propagating from the central Pacific and merging with a secondary one from eastern tropical South Indian Ocean. In 1982 occurred a strong El Nino event. However, for other El Nino or La Nina events (e.g. 1987 or 1999) the agreement between the ensemble members for simulating the rainfall dipole was not as good as in 1982, at least in part associated with unclear teleconnections (i.e. large dispersion between simulated members). It is also worth noting that only some "extreme" springs (i.e. too rainy or too dry in SESA in terms of PC1 index) are associated with the occurrence of El Nino or La Nina.

A rainy spring for SESA was 2003, but in this case almost all the ensemble members exhibit a precipitation dipole out of phase with respect to the observations (from this point of view is an opposite case to 1982). This year exhibits a zonally symmetric pattern of moderately positive SST anomalies throughout the tropics. In particular, near neutral conditions dominate across the equatorial Pacific. In this case, the ensemble mean does not exhibit any teleconnection through the South Pacific. Not having sectors with high temperature anomalies in the tropics is a source of additional uncertainty in the simulation of the SH extratropics since wave trains propagating
through the Southern Oceans are not excited uniformly in the different ensemble members (large inter-member spread in the circulation). Regional effects would be more important than remote forcing in this case.

In terms of intensity it is hard to separate from the analysis we did the influence of the strongest cases chosen. Further about the conclusions of the Indian Ocean we may not exclude that its variability is also induced by ENSO.

Acknowledgements. The research leading to these results has received funding from the European Community’s Seventh Framework Programme (FP7/2007-2013) under Grant Agreement N° 212492 (CLARIS LPB. A Europe-South America Network for Climate Change Assessment and Impact Studies in La Plata Basin). A. F. Carril and C. G. Menéndez were partially supported by PIP 112-200801-01788 (CONICET, Argentina) and PICT 2008-00237 (FONCYT, Argentina). The support of Italian Ministry of Education, University and Research, and Ministry for Environment, Land and Sea through the project GEMINA is gratefully acknowledged as well. Dr. L. Zamboni was partially supported by American Recovery and Reinvestment Act (ARRA) funding through the Office of Advanced Scientific Computing Research, Office of Science, U.S. Dept. of Energy, under Contract DE-AC02-06CH11357 and by the U.S. Department of Energy, Basic Energy Sciences, Office of Science, under contract # DE-AC02-06CH11357.
References


Carril AF, and co-authors (2012) Performance of a multi-RCM ensemble for South Eastern South America. Clim Dyn (under revision)


Mo KC (2000)


Vera C, Silvestri G, Liebmann B, Gonzalez P (2006) Climate change scenarios for seasonal pre-


Zamboni L, Cherchi A, Barreiro M, Kucharski F (2012) Dynamics of the impact of ENSO flavors on precipitation over La Plata Basin. Under review on Climate Dynamics


### Table 1
OND mean precipitation and its standard deviation (mm/d) averaged over the South American continent (SA) and over the La Plata Basin (LPB) region (65-47W, 37-19S) for CRU dataset (first row), AMIP-type ensemble (second row) and SSXX experiment (third row).

<table>
<thead>
<tr>
<th></th>
<th>CRU</th>
<th>AMIP-type</th>
<th>SSXX</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mean</td>
<td>std</td>
<td>mean</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SA OND</td>
<td>4.52</td>
<td>0.44</td>
<td>4.12</td>
</tr>
<tr>
<td>LPB OND</td>
<td>4.72</td>
<td>1.02</td>
<td>4.68</td>
</tr>
<tr>
<td></td>
<td>4.26</td>
<td>0.77</td>
<td>4.44</td>
</tr>
</tbody>
</table>

### Table 2
Correlation coefficients of South America OND precipitation anomalies first three principal components (PC1, PC2 and PC3) for the period 1979-2005 between CRU and CMAP datasets. An asterisk marks the values that are statistical significant at 95%.

<table>
<thead>
<tr>
<th></th>
<th>PC1(CRU)</th>
<th>PC2(CRU)</th>
<th>PC3(CRU)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PC1(CMAP)</td>
<td>-0.94*</td>
<td>-0.23</td>
<td>0.00</td>
</tr>
<tr>
<td>PC2(CMAP)</td>
<td>-0.17</td>
<td>0.67*</td>
<td>0.42</td>
</tr>
<tr>
<td>PC3(CMAP)</td>
<td>-0.11</td>
<td>0.33</td>
<td>-0.80*</td>
</tr>
</tbody>
</table>
Table 3. Correlation coefficients between NINO3.4 index and OND SA precipitation principal components for the AMIP-type ensemble. Values are reported for each member of the ensemble, including the mean in the bottom.

<table>
<thead>
<tr>
<th></th>
<th>PC1</th>
<th>PC2</th>
<th>PC3</th>
</tr>
</thead>
<tbody>
<tr>
<td>EXP01</td>
<td>0.60</td>
<td>-0.50</td>
<td>-0.36</td>
</tr>
<tr>
<td>EXP02</td>
<td>0.32</td>
<td>-0.57</td>
<td>-0.38</td>
</tr>
<tr>
<td>EXP03</td>
<td>0.40</td>
<td>-0.63</td>
<td>-0.09</td>
</tr>
<tr>
<td>EXP04</td>
<td>0.63</td>
<td>-0.38</td>
<td>-0.13</td>
</tr>
<tr>
<td>EXP05</td>
<td>0.44</td>
<td>-0.55</td>
<td>-0.06</td>
</tr>
<tr>
<td>EXP06</td>
<td>0.44</td>
<td>-0.42</td>
<td>-0.23</td>
</tr>
<tr>
<td>EXP07</td>
<td>0.52</td>
<td>-0.53</td>
<td>0.07</td>
</tr>
<tr>
<td>EXP08</td>
<td>0.54</td>
<td>-0.48</td>
<td>-0.29</td>
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<tr>
<td>EXP09</td>
<td>0.66</td>
<td>-0.41</td>
<td>-0.19</td>
</tr>
<tr>
<td>Mean of corr</td>
<td>0.55</td>
<td>-0.50</td>
<td>-0.18</td>
</tr>
</tbody>
</table>

Table 4. Averages of vertical velocity ($\omega$, mb/s) at 500 mb in the region Eq-20S, 75W-55W (2nd column) and of zonal velocity (m/s) at 200 mb in the region 20S-40S, 90W-60W (3rd column) for the In Phase Positive, Out of Phase Positive, Partial in Phase Positive, In Phase Negative, Out of Phase Negative and Partial in Phase Negative (from top to bottom) composites.

<table>
<thead>
<tr>
<th></th>
<th>$\omega$ 500 (mb/s)</th>
<th>u200 (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>In Phase Pos</td>
<td>0.65</td>
<td>2.82</td>
</tr>
<tr>
<td>Out Phase Pos</td>
<td>0.09</td>
<td>1.35</td>
</tr>
<tr>
<td>Partial Phase Pos</td>
<td>0.19</td>
<td>2.27</td>
</tr>
<tr>
<td>In Phase Neg</td>
<td>-0.26</td>
<td>-0.84</td>
</tr>
<tr>
<td>Out Phase Neg</td>
<td>-0.08</td>
<td>-0.17</td>
</tr>
<tr>
<td>Partial Phase Neg</td>
<td>-0.21</td>
<td>-0.75</td>
</tr>
</tbody>
</table>
Table 5. List of years where the model PC1 is "In Phase" (exceeding 1 standard deviation in the same direction) or "Out of Phase" (exceeding 1 standard deviation in the opposite direction) with the CRU PC1. Years are separated for wet LPB years (i.e. positive PC1 values) and for dry LPB years (i.e. negative PC1 values). Within each year the number of members having the same behavior is indicated in parentheses.

<table>
<thead>
<tr>
<th></th>
<th>Wet LPB yrs</th>
<th>Dry LPB yrs</th>
</tr>
</thead>
</table>
**Figure Captions**

**Fig. 1.** First three EOFs of OND mean South American precipitation (in the box shown but considering land-points only) for the period 1979-2005 for CRU (upper panels) and XieArkin (lower panels) dataset, respectively.

**Fig. 2.** First three EOFs modes and principal components (PCs) of OND South American precipitation (in the box shown but considering land-points only) for CRU dataset from 1948 to 2003.

**Fig. 3.** EOFs (1st and 2nd modes) of OND South American precipitation (in the box shown considering land-points only) for AMIP-type ensemble.

**Fig. 4.** Time-correlation coefficients of OND South America precipitation PC1 and OND SST for (a) HadISST/CRU datasets, (b) AMIP-type ensemble and (c) coupled model experiment (SSXX).

**Fig. 5.** Same as fig. 4 but a 7-years low-pass filter is applied to the PC1.

**Fig. 6.** 19 years sliding correlation (x-axis shows the first year of the 19 years interval) between SAM and LPB precipitation during OND for observations based on SLP (magenta line) and for the AMIP-type ensemble (black lines). Solid and dashed black lines represent the correlation applied to the ensemble mean and to the average of the correlation applied to each member of the ensemble, respectively. The horizontal solid lines correspond to the threshold values statistically significant at 95%.

**Fig. 7.** Composite anomalies of SST (°C, shaded) and 200 mb eddy streamfunction (10^6 m^2/s, contours) composite anomalies for wet and dry LPB years in (a,b) HadISST/CRU datasets and (c,d) AMIP-type ensemble.

**Fig. 8.** Composite anomalies of SST (°C, shaded) and 200 mb eddy streamfunction (10^6 m^2/s, contours) for wet (positive) and dry (negative) LPB years in AMIP-type ensemble members grouped as (a,b) "In Phase", (c,d) "Out of Phase" and (e,f) "Partial In Phase" PC1 values (the classification is described in the text).
Fig. 9. Same as fig. 8 but for vertically integrated moisture (kg/m², shaded) and vertically integrated moisture flux (kg/m s, vectors).

Fig. 10. 1982 OND composite of (a) SST (°C, shaded) and 200 mb eddy streamfunction (10⁶ m²/s, contours), (b) precipitation (mm/d, shaded) and 850 mb wind (m/s, vectors) and (c) 1982 OND standard deviation among members of 200 mb streamfunction (10⁶ m²/s) for the AMIP-type ensemble.

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