Extreme events in the La Plata basin: a retrospective analysis of what we have learned during CLARIS-LPB project

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ABSTRACT: Extreme climate events over the La Plata basin (LPB) can produce significant impacts due to the importance of the regional agriculture and hydroelectric power production for the local economy. Progress on describing, projecting and understanding extremes in LPB, in the framework of the CLARIS-LPB Europe-South America Network for Climate Change Assessment and Impact Studies in La Plata Basin Project, are reviewed. The paper is based on recent studies and publications, as well as some new diagnostics as indicators of works in progress, and can be considered as an update for the LPB region of previous reviews by Cavalcanti et al. (2015; *J Hydrol 523:211–230*) and Rusticucci (2012; *Atmos Res 106:1–17*). Despite the significant advances on regional extremes, some gaps have been identified, and many challenges remain. Much of the recent progress considers temperature and precipitation extremes on timescales varying from synoptic to long-term variability and climate change, essential for impact and vulnerability assessments. Research lines on extremes requiring further efforts include the relative roles of local versus remote forcings, the impact of land use and land management changes, the specific role of soil moisture and land-atmosphere feedbacks as catalysts for heat waves, the impact of the local inhomogeneities in soil moisture, feedback and uncertainties in projections of extremes, as well as seasonal forecast and climate change attribution studies. We suggest combining intensive monitoring and modelling to address these issues.

KEY WORDS: Temperature extremes · Precipitation extremes · La Plata Basin · Climate models · Regional climate change

1. INTRODUCTION

The La Plata Basin (LPB) is a region in southeastern South America (SESA) of ~4 144 000 km^2 that covers parts of Argentina, Brazil, Bolivia, Paraguay and Uruguay. The basin, which is one of the largest on earth, is the most populated region of South America (SA), and relies economically on agriculture and hydroelectric power. Extreme events occurring in the LPB can greatly affect the natural systems of the basin as well as the social and economic sectors.

The region is vulnerable to several types of precipitation extremes, spanning a wide range of scales. On hourly to daily time scales, flash flooding due to intense precipitation events associated with deep convective cells (e.g. Cavalcanti 2012, Cavalcanti et al. 2013, Fernández-Long et al. 2013, Boers et al. 2014a) can destroy buildings and threaten lives. On monthly to seasonal time scales, deficit or surplus precipitation, usually influenced by El Niño–southern oscillation (ENSO) episodes (e.g. Grimm et al. 2000, Berbery & Barros 2002, Camilloni & Barros 2003, Carvalho et al. 2004, Cavalcanti 2012, Müller & Berri 2012, Tedeschi et al. 2014), affect the agriculture, water resources, fluvial transportation, energy supply and the population living in the affected areas. Lastly, drought-like conditions accompanied by warm spells (mostly occurring during the summertime in the LPB; Barrucand et al. 2014) cause unfavorable effects on vegetation and crops and eventually lead to wildfires. Cold and heat outbreaks have a huge impact on agriculture, human health and energy demand. Wintertime cold air outbreaks (e.g. Garreaud 1999, 2000, Vera & Vigliarolo 2000, Müller & Ambrizzi 2007, Müller & Berri 2012) sometimes produce frosts, causing serious damage to crops and fruits. Summertime cold air incursions result in less dramatic variations of temperature, but produce deep convection close to the connection between the northwestern Argentinean Low and the trough to the south of it, forming a low pressure tongue around the saddle point (e.g. Arraut & Barbosa 2009, Boers et al. 2014a, 2015). The intrusions of tropical air into higher latitudes have been studied in less detail (e.g. Rusticucci & Vargas 1995b, Rusticucci et al. 2003, Cerne et al. 2007). These intrusions occur close to the Andes between 20°S and 30°S. This region is connected to the climatological position of the South Atlantic Convergence Zone (SACZ) (Seluchi & Marengo 2000), and SACZ activity on intraseasonal scales plays an important role in the development of the most persistent heat waves (Cerne & Vera 2011) and extreme rainfall events (Carvalho et al. 2004).

A changing climate leads to changes in the frequency, intensity, duration, timing and spatial extent of extremes, which could result in unprecedented extremes (e.g. Marengo et al. 2009, Seneviratne et al. 2012). In this context, the impacts of climate change on a community or region depend on the severity of the extreme events as well as on the vulnerability of this community or region. Christensen et al. (2007) noted the lack of regional studies related to extreme events and their projections over SA. With this in mind, and under the premise that coordinated efforts are likely to reap greater rewards than individual actions, one of the CLARIS-LPB goals was to advance the understanding of the key processes associated with the occurrence of extremes. Particular attention was paid to the study of trends and projections, as well as to the understanding of the role of the large-scale forcing versus the local feedbacks that may lead to the occurrence of extremes. As discussed by Rusticucci (2012), a high spatial and temporal resolution database is a necessary input to the analysis of extreme events. CLARIS-LPB has played an important role in the creation of databases, and the working group on extremes has benefitted from that. Some activities have used the quality controlled daily station data collected by the database working group (Penalba et al. 2014) and the gridded datasets delivered by the project (Tencer et al. 2011, Jones et al. 2013). In other cases, different sets of reanalysis as well as model integrations from international programs (e.g. CMIP3 and CMIP5), from ad hoc experiments (e.g. see Cavalcanti et al. 2015 and references therein) and from the CLARIS (Menéndez et al. 2010b) and CLARIS-LPB regional climate models (RCMs) coordinated experiments (Solman et al. 2013, Sánchez et al. 2015) have been used. At this point, it is worth noting that although regional climate modeling has undergone continuous development during the last decades, in the IPCC AR4, all the regional-scale information for South America was taken from general circulation models (GCMs). Following Roads et al. (2003) and Marengo et al. (2010a), a new set of coordinated RCMs over South America was available (e.g. Menéndez et al. 2010a,b) in the framework of CLARIS (Boulanger et al. 2010). Nevertheless, most of the information related to extreme events was derived from the CMIP3 global models (e.g. Menéndez & Carril 2010). During the CLARIS-LPB project, some studies on extremes were carried out by analyzing simulations of RCMs integrated in the framework of both the CLARIS (e.g. Carril et al. 2012) and CLARIS-LPB (e.g. Cavalcanti et al. 2015, Lopez-Franca et al. 2015) projects. Moreover, because the

RCMs used to simulate South American climate in the framework of both CLARIS and CLARIS-LPB projects were originally developed for Europe and North America, the RCMs were evaluated over South America (e.g. Carril et al. 2012, Solman et al. 2013), and sensitivity experiments were carried out to attain an optimal model setup for optimum performance (Solman & Pessacg 2012).

This paper focuses on what we have learned during the CLARIS-LPB EU FP7 project regarding the occurrence of extremes over LPB (without excluding the works performed outside the project).

2. TEMPERATURE EXTREMES

In a previous review paper about the extremes in temperature, Rusticucci (2012) highlighted the geographical imbalance with respect to the number of published studies on temperature extremes in South America, limiting our understanding of this over the continent, mainly due to the lack of data. In this section, we highlight the advances on temperature-related extremes reached in the last years, during the CLARIS-LPB project. Definitions of indices cited in the paper are summarized in Table 1. Most of these indices were defined by the Expert Team on Climate Change Detection and Indices (ETCCDI; Zhang et al. 2011).

2.1. Variability and trends in temperature extremes

When the project began, we knew that the strongest changes in daily maximum (TX) and minimum temperature (TN) values in a vast region of LPB had been registered in summer: TN was increasing (but its interdiurnal standard deviation was decreasing), and TX was mostly decreasing (e.g. Sansigolo & Kayano 2010). Overall, negative trends had been documented for the number of cold nights and warm days per summer, while the number of warm nights and cold days had increased at certain locations. A positive trend in the summertime mean temperature in central-northern Argentina had been mostly explained by the increasing frequency of warm events (see e.g. Barrucand & Rusticucci 2001, Rusticucci & Barrucand 2004, Vincent et al. 2005). Focusing on long-term variabilities, Vargas & Naumann (2008) showed that the occurrence of wet days is one of the main factors driving secular variations of TN and TX and variations in precipitation.

CALRIS and CLARIS-LPB allowed us to advance knowledge about temperature extremes over the Argentinean sector of LPB, and to start analyzing series of temperature extremes over Uruguay for which there was no previous literature on the subject.

Over the Argentinean sector of the LPB, it was found that return values of the highest (lowest) maximum (minimum) temperature of the year denote changes in the return values related to decadal variability (Rusticucci & Tencer 2008, Tencer & Rusticucci 2012). In particular, over Buenos Aires Central Observatory after 1977, the frequency of occurrence of high values of TN (TX) has been greater (lower) than before 1977, leading to a decrease in the annual temperature range (Rusticucci & Tencer 2008). The frequency of occurrence of extremes in the south of the Argentinean LPB has changed throughout the last century, and changes go beyond the climatic jump, or discontinuity, of 1976 (Tencer & Rusticucci 2012). The intensity of warm extremes in both TX and TN has decreased over the period 1941–2000, while the frequency of occurrence of TX (TN) has decreased (increased) from 1941–1960 to 1961–1980 (1961–1980 to 1981–2000). In contrast, the intensity of cold extremes has mostly decreased. Rusticucci & Tencer (2008) and Tencer & Rusticucci (2012) illustrate that changes in the intensity and frequency of occurrence of annual and daily extremes are not always consistent, implying that the behavior of the most extreme value in the year is not always equivalent to the behavior of the upper or lower 10 % tail.

Over Uruguay, Renom (2009) reported on the existence of negative trends in the number of warm (cold) days and cold (cold) nights for summer (winter) and on warming in autumn night temperatures. Trends in temperature indices were not affected by any significant trend in the occurrence of summer heat waves. Moreover, preliminary results by Renom et al. (2014) show a negative trend in the occurrence of frost events during the cold months (May to September), with May being the month with largest decrease in the occurrence of frosts, in agreement with the warming trends in night temperatures in autumn.

2.2. Remote forcings

Regional climate is modulated by the interaction of local and remote forcings. Regarding the large-scale patterns that lead to the occurrence of extremes, pre-project knowledge related the interannual variability of temperature extremes with the phases of ENSO

Table 1. A summary of the indices cited throughout the text, for both temperature and precipitation (Pp). The percentiles of reference are calendar day percentiles, centered on a 5 d window and for the base period 1961 to 1990

No.	Index	Name	Description	Units
1	TN10	Cold nights	Percentage of days when TN < 10th percentile	%
2	TX10	Cold days	Percentage of days when TX < 10th percentile	%
3	TN90	Warm nights	Percentage of days when TN > 90th percentile	%
4	TX90	Warm days	Percentage of days when TX > 90th percentile	%
5	WSDI	Warm spell duration index	Annual count of days with ≥6 consecutive days when TX > 90th percentile	(d yr^{-1})
6	WSDI_NUM	Frequency of warm spells	Annual frequency of WSDI	(events yr^{-1})
7	CSDI	Cold spell duration index	Annual count of days with ≥6 consecutive days when TN< 10th percentile	(d yr^{-1})
8	CDD	Consecutive dry days	Maximum number of consecutive days with Pp < 1 mm d^{-1}	(no. d)
9	DIER75	Intensity of daily extreme precipitation	The ratio between the monthly-accumulated extreme precipitation and the number of days with extreme precipitation, where extreme precipitation is when Pp > 75th percentile	(mm d^{-1})
10	Rx5day	5 d precipitation	Amount of monthly maximum of consecutive 5 d precipitation	(mm)
11	SPI	Standardized precipitation index	The number of standard deviations that the observed value would deviate from the long-term mean for a normally distributed random variable. Because precipitation is not normally distributed, a transformation is first applied so that the transformed precipitation values follow a normal distribution. A separate SPI value is calculated for a selection of time scales	(dimensionless)
12	DHI	Drought hazard index	Index based on drought frequencies weighted by severity, with 3 categories identified through the tertiles of its spatial distribution	(dimensionless)

(e.g. Müller et al. 2000, Rusticucci & Vargas 2001, Barros et al. 2002), although large inter-monthly variability had been noticed (e.g. Grimm et al. 2000, Rusticucci & Vargas 2002, Rusticucci et al. 2003). In particular, on the intraseasonal scale, the phases of the Madden-Julian Oscillation (MJO) display a consistent signal with the wintertime temperature variability in the LPB (Naumann & Vargas 2010). Some other authors have discussed processes associated with cold surges (e.g. Müller 2007, Müller & Berri 2007, Müller & Ambrizzi 2010). In particular, the occurrence of wintertime cold outbreaks leading to generalized frosts is usually associated with wave trains propagating inside the subtropical and polar waveguides that merge just before entering the continent, favoring anticyclonic anomalies and intense cold advection (Müller & Ambrizzi 2007). The mechanisms related to the tropical air intrusions have been less studied. On one hand, the physical processes related to the occurrence of heat waves in central-north Argentina are related to the activity of synoptic-scale waves (Rusticucci & Vargas 1995a), but on the other hand, the intraseasonal variability has a strong impact on the evolution of the summer-

time anomalies (Cerne et al. 2007). A large number of persistent summertime heat waves occur in association with the strengthening of anticyclonic anomalies (embedded in a wave train excited by anomalous convection at the equatorial western and central Pacific) that are in turn associated with the active phase of the SACZ (Cerne & Vera 2011).

During the project, the large-scale atmospheric anomalies in which temperature extremes are embedded have been further characterized. A principal finding is that the results are non-stationary: the relationships have changed over time (Renom et al. 2011). Before 1976, the interannual variability of the summertime cold nights (TN10; see Table 1) coincided with the negative phase of the southern annular mode (SAM), related to cyclonic anomalies centered off Uruguay and favoring the entrance of cold air from the south. The interannual variability of the wintertime warm nights (TN90; see Table 1) was dominated by ENSO teleconnections, driving anomalous northerly warm winds into Uruguay. After 1976, the summertime cold-nights have been associated with an isolated vortex at upper levels over southeastern South America. Moreover, as ENSO

teleconnections are weakened, the variability of the wintertime TN90 is dominated by a barotropic pattern of anticyclonic anomalies located in the South Atlantic Ocean and cyclonic anomalies over South America. This configuration also strengthens the northward flow of warm air over Uruguay. Renom et al. (2011) suggest that changes in El Niño evolution after 1976 may have played a role in altering the relationship between occurrence of extreme temperatures in Uruguay and the atmospheric circulation.

2.3. Projections and uncertainties

When CLARIS-LPB started, we knew that reanalysis had to be used with caution for studies of the magnitude of day-to-day temperature changes. Rusticucci & Kousky (2002) had documented that the NCEP reanalysis is able to reproduce the timing of mid-latitude extreme events in temperature, but fails to capture the amplitude and frequency of extremes in both subtropical regions and areas of high relief. During the CLARIS-LPB project, Zaninelli et al. (2015a,b) explored the uncertainty of TX and TN from reanalysis in SESA. Three different reanalyses (NCEP/NCAR, Kalnay et al. 1996; ERA40, Uppala et al. 2005; and 20CR/NOAA, Compo et al. 2011) have been compared using a novel daily gridded dataset of minimum and maximum surface temperature (Tencer et al. 2011; hereafter TNCR dataset) and with integrations of 4 regional climate models driven by boundary conditions from ERA40 (details are in Carril et al. 2012). Fig. 1 summarizes how well RCMs (upper panels) and reanalyses (lower panels) capture the geographical patterns of anomalies of the 75th percentile of TX for DJF (warm extremes, left panels) and of the 25th percentile of TN for JJA (cold extremes, right panels) in SESA using Taylor diagrams (Taylor 2001). Those diagrams are a graphical representation of the statistical relationships between 2 random variables: (1) the series of the P75 of TX from TNCR (our reference) and (2) the series of the P75 of TX from other datasets. To estimate the 3 statistics (standard deviation, correlation and centered root mean square difference), input series must have the same dimension. Therefore, in the case of the ensemble, the input series is the mean among the P75 of TX series of every climatology. In addition, all the statistics are calculated after removing the spatial mean value of each dataset. In general, the ensemble mean (either for regional models or reanalyses) has a good performance relative to each individual ensemble members for both variables and both seasons, which corroborates previous outcomes

(e.g. Menéndez et al. 2010a). The spread between ensemble members for TX in DJF is larger than for TN in JJA for both regional models and reanalyses. Note that the inter-member spread is similar for the regional models and reanalyses in both seasons, and that errors are similar for both the models ensemble and the reanalyses ensemble. Based on that similarity, Zaninelli et al. (2015a,b) concluded that maximum/ minimum surface temperature from reanalysis cannot be used to validate regional models, highlighting the existing uncertainties for studying extremes in temperature using both state-of-the-art RCMs and gridded databases.

Assuming that TNCR is an appropriate climatology to validate models in the LPB region, the capability of the CLARIS-LPB RCMs to capture the interannual variability of extremes in TX and TN is displayed in Fig. 2. The figure depicts the indices for warm days and cold nights (TX90 and TN10, respectively); (see Table 1), representative of the northern-LPB (15°–25°S; 55°–45°W) and southern-LPB (25°–35°S; 60°–50°W) regions (NLPB and SLPB, respectively) for the 11 yr period in which the TNCR climatology and the integrations overlap (the mean spatial structure of the fields can be found in Lopez-Franca et al. 2015). As a first criterion, we consider the ensemble as 'satisfactory' for simulating interannual variability of the index if (1) the observed index correlates well with the ensemble mean index and if (2) its observed annual value is contained within the simulated min–max range of the ensemble mean. In general, this criterion is fulfilled in both regions. Nevertheless, the spread among the members of the ensemble is somewhat large: in many cases, the ensemble uncertainty (i.e. the min–max range) is larger than the interannual variability of the ensemble mean, limiting the assessments of the models. The correlation between the ensemble mean and TNCR is best for the NLPB for both indices. Similarly, the inter-model spread also tends to be lower in the NLPB. This pattern suggests that models have more difficulties simulating the extremes of TX and TN in the SLPB. This result is consistent with the fact that the average temperature bias tends to be higher in the SLPB than in the NLPB (e.g. Fig. S11.25 in Christensen et al. 2007). The reasons behind this pattern are not entirely clear, and are an active research topic. Menéndez et al. (2016, this Special) showed that in the SLPB, the soil influences the variability of temperature through evapotranspiration, while in the NLPB, this process has a relatively minor significance. Larger errors in the SLPB may thus suggest problems simulating local feedbacks between precipitation, soil moisture and temperature in that
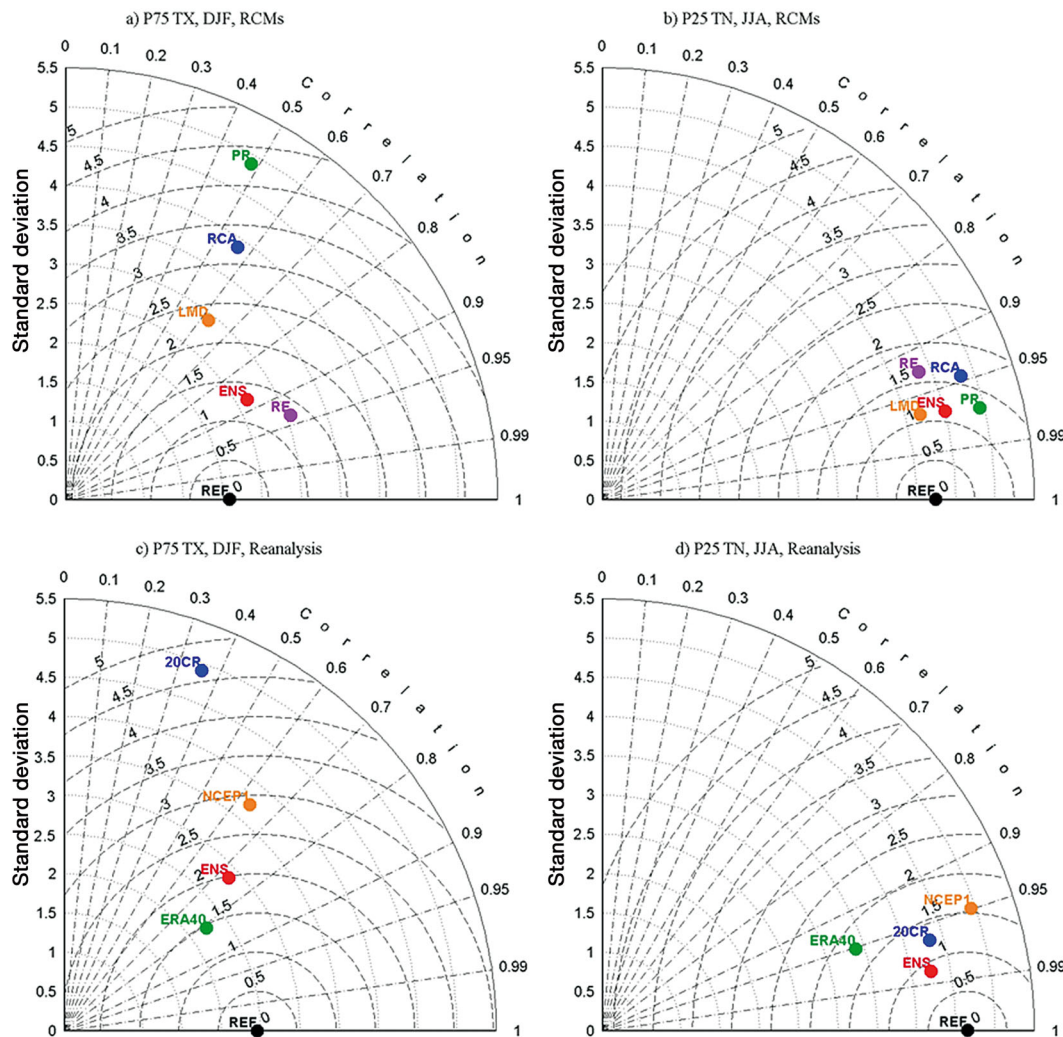
Fig. 1. Taylor diagrams illustrating the agreement or disagreement between the spatial anomalies in temperature from different data sets. The upper panels show regional models (PROMES in green, RCA in blue, LMDZ in orange, REMO in violet and the multi-model ensemble [ENS] in red). The bottom panels show reanalysis (ERA 40 in green, 20CR in blue, NCEP in orange and the multi-reanalysis ensemble [ENS] in red). In all the panels, TNCR is the climatology of reference (REF is the black point), and the domain is SESA. Each panel illustrates 3 statistics: the standard deviation (vertical axis), the correlation coefficient (axial) and the mean root square error (concentric). Left column is for threshold of warm extremes in maximum temperature in DJF, and right column is for cold extremes in minimum temperature in JJA

area. This process is more important in summer, and thus influences TX90. In winter, models typically underestimate rainfall in SLPB (Solman et al. 2013), suggesting that related variables, such as cloud cover or wind, may also be wrong. Lopez-Franca et al. (2015) showed the sensitivity of TN to variations in cloud cover and meridional wind in LPB.

Other CLARIS-LPB participants have explored the ability of the IPCC AR4 global models to simulate mean values, interannual variability and trends of annual indices of extremes based on TN. Results show that in the lowlands of the LPB and for the last decades of the 20th century, the models represent the number of warm nights better than the number of

frost days (Rusticucci et al. 2010). The models are also able to capture the positive trend observed in the warm nights during the last decades (Marengo et al. 2010b).

Regarding climate change projections of warm (cold) days (nights) over South America (indices 1 to 4 in Table 1), using a subset of CLARIS-LPB RCMs, Lopez-Franca et al. (2015) found that the areas where the warm nights are projected to increase more than warm days are also those areas where TN is projected to increase more than TX in summer, as is the case for the LPB. The lower increase in warm day than warm night occurrence over the LPB under future conditions could reflect the effects of increases
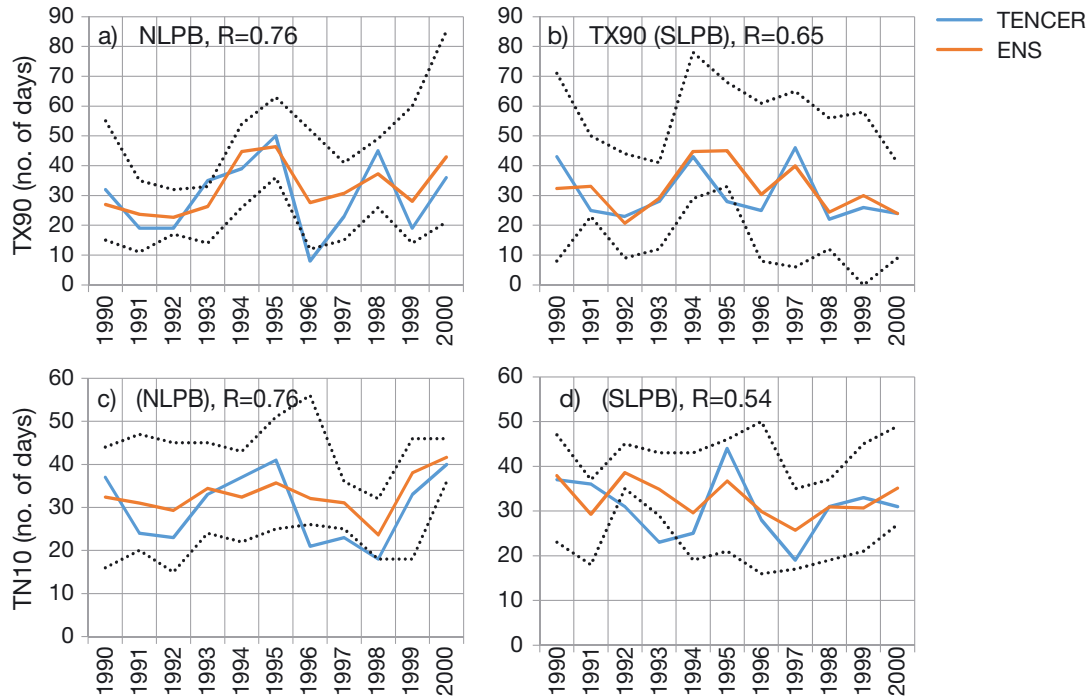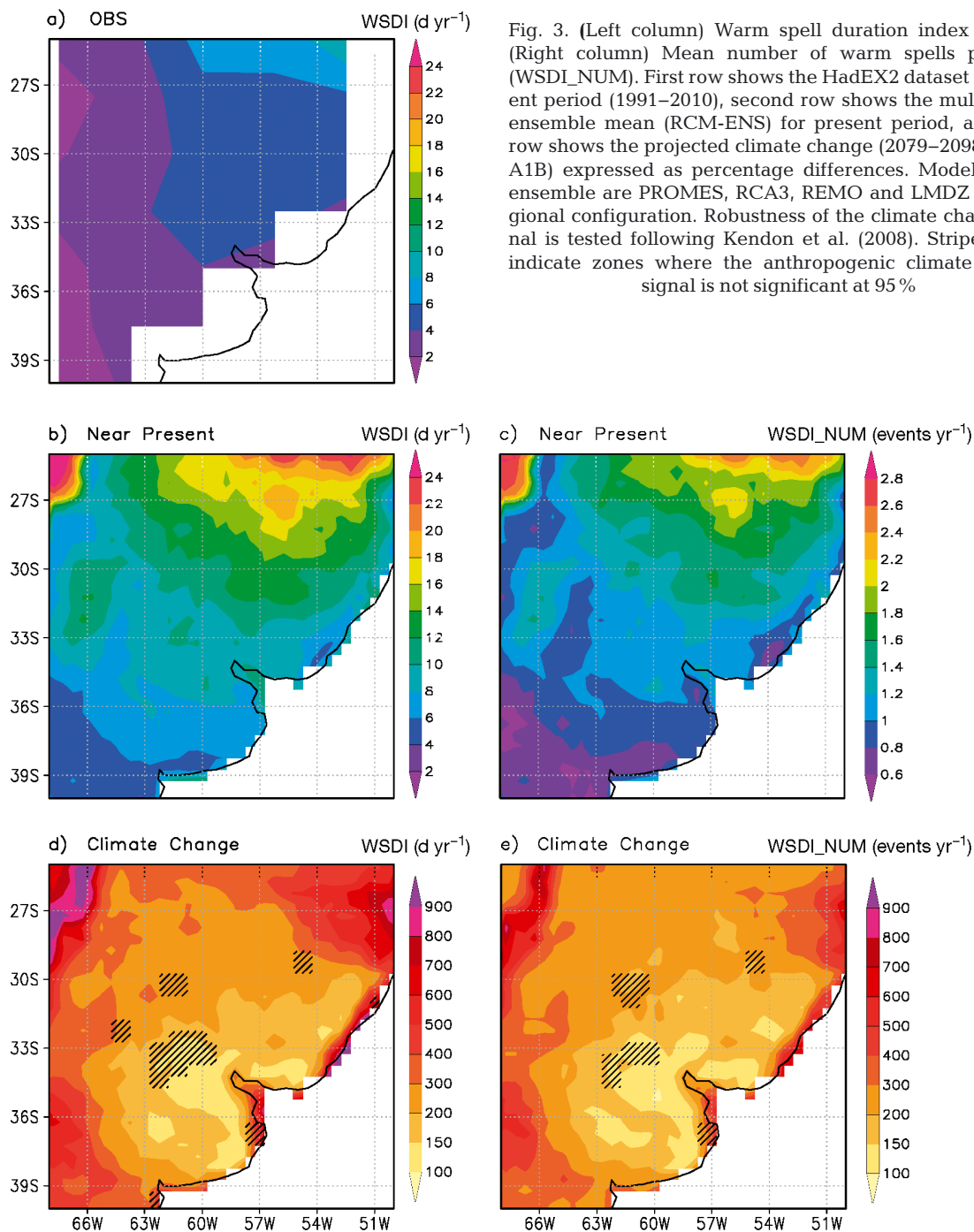
Fig. 2. Interannual variability of extremes in 2 boxes representing the north-LPB region (left column) and south-LPB region (right column). First row is for TX90 (warm day-times) and second row for TN10 (cold nights). Each panel displays time series according to TNCR climatology (blue), the ensemble mean (red), the correlation coefficient between them (R) and a measure of the inter-model spread (max–min range; dotted lines). Models in the ensemble are PROMES, ETA, REGCM3, REMO, LMDZ, MM5 and RCA

in cloudiness over the LPB domain under future conditions (Lopez-Franca et al. 2015). Two main factors are involved related to the cloud effect. First, the cloud effect tends to decrease temperature during the daytime (TX) because clouds reflect the solar radiation. Second, clouds enhance the greenhouse warming during nighttime, increasing TN. Both factors tend to damp the diurnal temperature range. This result is consistent with Zhou et al. (2009), who found that large decreases in diurnal temperature range reflect the effects of increases in cloudiness over the LPB under future conditions (see their Fig. 2). A factor that could contribute to a dampening of the projected increase of TX is the response of summer evapotranspiration in the LPB (Menéndez et al. 2016).

A complementary analysis, using the same subset of CLARIS-LPB RCMs as Lopez-Franca et al. (2015), considers annual warm spell indices (indices 5 and 6 in Table 1) and is displayed in Fig. 3. TheHadEX2 dataset (Donat et al. 2013), available on a 2.5° × 3.75° longitude-latitude grid from the CLIMDEX project (Alexander et al. 2006; www.climdex.org) is used to assess the simulated ETCCDI temperature indices. It is known that the models' surface temperature has a warm bias in the LPB (Sánchez et al. 2015). That bias is probably related to a deficit of

precipitation in the LPB (Menéndez et al. 2016). Consecutive extremes in TX (WSDI, see Table 1) also show a warm bias: models overestimate the duration of warm spells by a rate of about 60 to 150% in LPB. Nevertheless, the geographical structure of the field is well captured (WSDI increases from SW to NE; Fig. 3a,b), and the pattern of mean number of warm spells registered per year (Fig. 3c) is similar to that of WSDI (northern LPB is affected by particularly large and frequent warm spells). Projections to the end of the 21st century (forced by SRES-A1B) indicate that changes in the duration and frequency of warm spells have a similar geographical pattern, with a maximum north of 30°S and in coastal regions (Fig. 3d,e). The northern portion of the LPB appears as a particularly vulnerable region to the occurrence of warm spells in the future, with spells up to 600% larger and 400% more frequent than in present climate (in the southern portion of the LPB, the rate of change is about 200% larger and more frequent). This result could be interpreted in the context of previous studies. Results by Junquas et al. (2012) confirm that future summertime rainfall variability in SESA is strongly related to changes in the SACZ dipole pattern, associated with an increase in the frequency of the

Fig. 3. (Left column) Warm spell duration index (WSDI). (Right column) Mean number of warm spells per year (WSDI_NUM). First row shows the HadEX2 dataset for present period (1991–2010), second row shows the multi-model ensemble mean (RCM-ENS) for present period, and third row shows the projected climate change (2079–2098, SRES-A1B) expressed as percentage differences. Models in the ensemble are PROMES, RCA3, REMO and LMDZ with regional configuration. Robustness of the climate change signal is tested following Kendon et al. (2008). Striped areas indicate zones where the anthropogenic climate change signal is not significant at 95%

dipole events (i.e. above-normal rainfall in SESA and below-normal rainfall in the SACZ region). This projection could play a role in modulating the increasing trend of heat waves in the southern portion of the LPB; Cerne & Vera (2011) have shown that 73% of the heat waves in the southern portion of the LPB are associated with intensified SACZ activity. Regarding cold spells, when calculating the cold spell duration index (CSDI, see Table 1) fields

from both HadEX2 and RCMs, the CSDI is almost constant and close to zero in the LPB, suggesting that the commonly accepted definition for CSDI is not appropriate for the LPB region. Previous works on cold spells in Argentina note that using the 25th percentile as the threshold for cold days (instead of the 10th percentile imposed by ETCCDI), the typical duration for summertime (wintertime) cold waves are 4 (6) d (Rusticucci & Vargas 2001).

3. PRECIPITATION EXTREMES

Here, we highlight the advances during CLARIS-LPB related to the occurrence and intensity of wet/dry episodes and spells, as well as recent observed trends, projections and uncertainties.

3.1. Variability and trends in wet and dry extremes

Research on interannual variability and trends in rainfall extremes has been increasing during recent years. In particular, when CLARIS-LPB started, we knew that (1) the annual cycle of daily events is strongly related to the annual cycle of monthly precipitation (Penalba & Robledo 2010); (2) trends in extremes in the period 1961–2000 are similar to the trends in the total annual rainfall (Haylock et al. 2006); and (3) trends in frequencies of extreme daily precipitation have been positive in central and eastern Argentina (period 1959–2002, Re et al. 2006) and in the subtropical sector of Brazil (Groisman et al. 2005) since the 1940s.

Outcomes from CLARIS-LPB show that the frequency of monthly wet extremes is greatest in the southern portion of the LPB (Cavalcanti et al. 2015). Moreover, positive trends in the frequency of occurrence of daily extreme precipitation events in the LPB have been observed (period 1961–2000; Penalba & Robledo 2010). The largest increase occurs in

southern Brazil, while negative trends are observed in winter in the southern LPB. The changes in the frequency of daily extremes have had an impact on the annual cycle of rainfall. In particular, Sugahara et al. (2009) documented positive trends in the frequency of days with extreme rainfall in Sao Paulo and showed that the threshold of the 90th percentile has increased by 40 mm in the period 1933–2005.

Trends in the intensity of daily extreme rainfall are illustrated by changes of the DIER75 index (see Table 1) and displayed in Fig. 4. DIER75 has increased (decreased) over vast areas of the LPB in fall and spring (winter). In summer, changes are spatially less uniform: DIER75 has decreased (increased) over central-east Paraguay, Parana and Mato Grosso (Minas Gerais, Rio Grande do Sul and the western Argentinean sector of the LPB).

From the synoptic climatology perspective, the study of the occurrence of daily rainfall extreme events and the associated circulation could be addressed through either a circulation-to-environment approach (in which the classification of weather patterns into categories is first performed and then related to an extreme event) or an environment-to-circulation approach (in which a particular extreme event determines which days are classified) (Yarnal & Draves 1993). The literature on South American synoptic climatology has mostly dealt with precipitation extreme events from the latter approach (e.g. Alessandro & Lichtenstein 1996, Alessandro 1998, Carvalho et al.
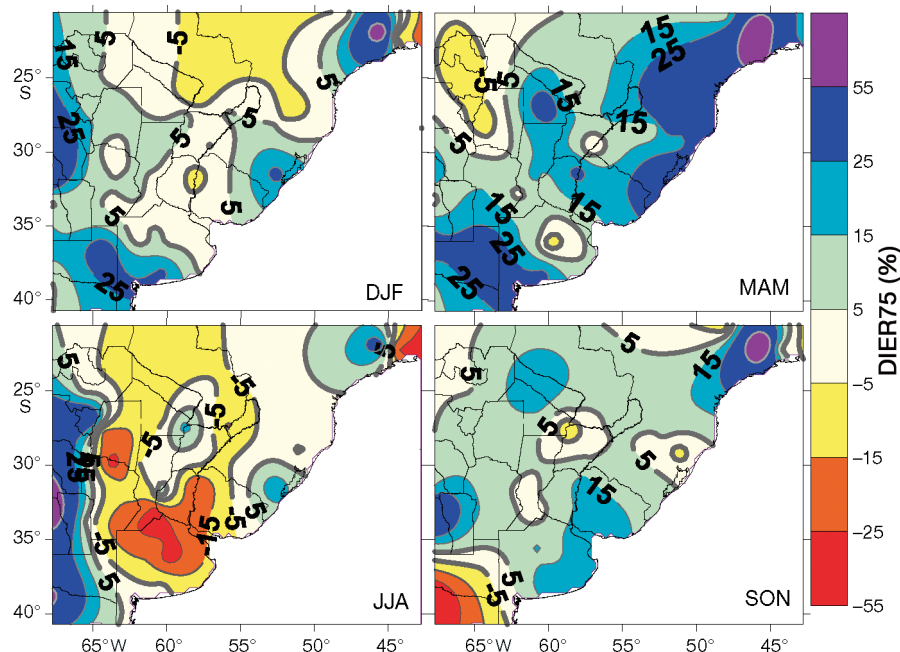

Fig. 4. Observed changes of DIER75 between 1961–1975 and 1980–1996, for the austral summer (DJF), autumn (MAM), winter (JJA) and spring (SON). Seasonal mean values of the percentile 75th are in Fig. 15 in Cavalcanti et al. (2015)

2002, Labraga et al. 2002, Bischoff & Vargas 2006, Cavalcanti 2012). The former approach has received less attention (e.g. Chaves & Cavalcanti 2001, Solman & Menéndez 2003, Bettolli & Penalba 2014). During CLARIS-LPB, Penalba & Bettolli (2011) employed a circulation-to-environment scheme to identify sea level pressure types and to associate these types with precipitation events of the central Pampas region during summer and winter. They studied frequency, distribution, temporal variability and the probability of daily rainfall conditioned to each circulation type. The probability of summer dry days is enhanced by the most persistent circulation type, corresponding to an intensification of the southern Atlantic anticyclone, which diverts perturbations to the south. Heavy rainy days are benefited by the occurrence of a cyclonic anomaly at the center of the continent associated with a cold front passage. Winter dry days are favored by a high-pressure system that extends from the Atlantic Ocean to the center of the continent. Heavy rainy days are benefited by a high-pressure system at the south of the continent, enhancing an anomalous flow from the east-southeast to the central region of Argentina and a corresponding moisture advection at low levels. The authors also assessed the ability of 17 atmospheric global circulation models to simulate the daily rainfall events in the Pampas (characterized by indices defined in terms of counts of days crossing thresholds) and the synoptic structures associated with each group. Most of the models are able to reproduce the observed seasonal differences between the statistics of daily rainfall fairly well. In particular, the relation between circulation patterns and rainfall in winter is better represented than in summer, due to the models' ability (deficiency) to reproduce synoptic scale patterns (small-scale processes) associated with the wintertime (summertime) precipitation (see Table 4 in Bettolli & Penalba 2014). The projections for the 21st century suggest an increase in the variability of daily rainfall, with a trend of reduction of the summer circulation patterns associated with dry days and an increase of the frequencies of the patterns associated with rainfall over the Pampas region.

Extratropical cyclones are particularly relevant atmospheric systems affecting the weather conditions in the LPB. The western South Atlantic Ocean is a region with a high frequency of these systems (e.g. Sinclair 1994, 1995, 1997, Simmonds & Keay 2000a,b, Hoskins & Hodges 2005). During CLARIS-LPB, Pereira (2013) investigated the relationship between cyclogeneses and extreme rainfall over the LPB. Reboita et al. (2010) confirmed the existence of 3

main regions with high frequencies of cyclogenesis over the southwestern South Atlantic Ocean: the south/southeast coasts of Brazil, the Uruguayan and the extreme southern Brazilian coasts (UsB) and the southern coast of Argentina. For UsB, Pereira (2010) showed that cyclones are related to extreme rainfall events over the LPB (Fig. 5). When all cyclones are considered, a large density of event occurs eastward of Uruguay (Fig. 5a), and in the composite of rainfall extremes (85th percentile; Fig. 5b), the most intense values are located over southern Brazil, Uruguay and northeast of Argentina. The spatial pattern of rainfall extremes in Fig. 5b is similar to that found by Arraut & Barbosa (2009) associated with high frequency of frontogenesis over central-northern Argentina. This similarity occurs because most of the cyclogeneses in UsB are of extratropical type with associated cold and warm fronts. For the initially most intense cyclogeneses, the density is larger over the ocean (Fig. 5c), and the associated composite of rainfall extremes (85th percentile) shows most intense values covering southern Brazil, northeastern Argentina and Uruguay and extending southeastward over the ocean (Fig. 5d). The cyclogeneses density is sensitive to the choice of the initial cyclonic vorticity threshold (Reboita et al. 2010), and furthermore, when intense cyclones occur, the area impacted by extreme rainfall increases. Seasonal analysis of the relationship between the density of the cyclones and extreme precipitation (not shown) indicates that the area affected by extremes is larger during winter and spring than during the rest of the year.

Comparing wet and dry extremes, Cavalcanti et al. (2015) showed that the frequency of observed monthly wet extremes is higher than the frequency of dry extremes in the whole LPB region. Before CLARIS-LPB, the literature about droughts mainly referred to the rainfall conditions over southern South America during La Niña conditions. Dry conditions are related to positive pressure anomalies (Barrucand et al. 2007) and to negative anomalies in the moisture transport both from the Amazonia and from the Atlantic Ocean (Labraga et al. 2002). Although several indices have been used to characterize drought events in the Pampas (e.g. Scian & Donnari 1997, Minetti et al. 2007, Serio et al. 2010), intercomparison of these studies is not an easy task. During CLARIS-LPB, we used the standardized precipitation index (SPI, McKee et al. 1993; see Table 1), an optimum index for studying droughts in the LPB (Penalba & Rivera 2015). Results for 1961–2008 indicate that, on average, the LPB is affected by moderate long-term (12 mo) drought conditions (SPI12
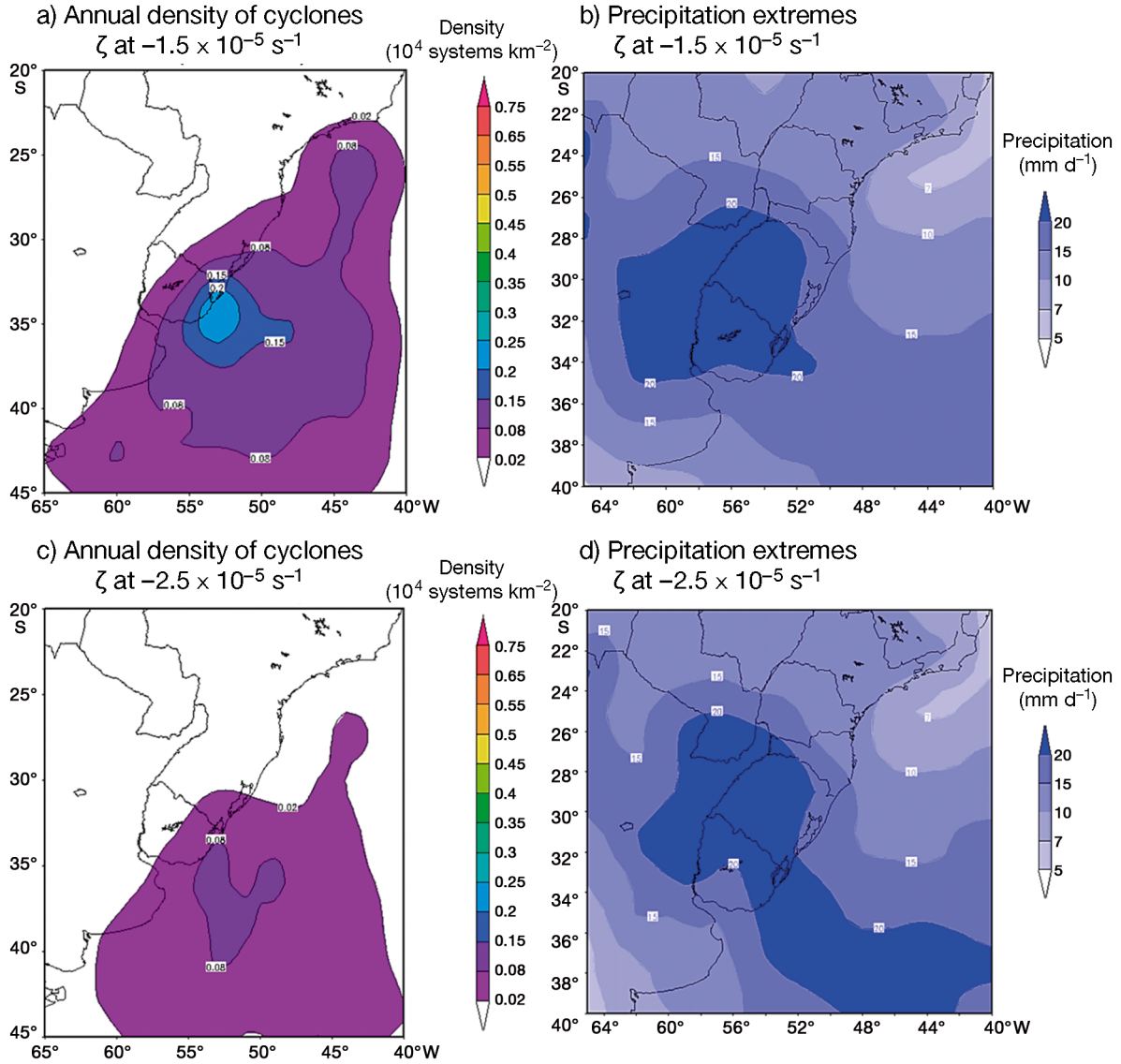
Fig. 5. (Left column) Annual density of cyclones related to the extreme precipitation events over LPB for 2 initial relative vorticity ζ thresholds: (a,b) -1.5×10^{-5} s^{-1} and (c,d) -2.5×10^{-5} s^{-1}, using ERA-Interim reanalysis for the period 1989 to 2007. (Right column) Composites of precipitation extremes (percentile 85) related to the density of cyclones for both vorticity thresholds. Figure adapted from Pereira (2013)

values lower than −1.0, which means that the precipitation departures from average conditions exceed 1 standard deviation). Moreover, drought occurrence in the LPB is high, with >20 drought events recorded during that period. These events have a mean duration of 3 to 6 mo, although extreme events can last for >20 mo. In terms of a drought hazard index (DHI; Shahid & Behrawan 2008), most of the productive area of the LPB presents low to moderate drought hazard, while the western portion of the LPB shows a higher risk (Cavalcanti et al. 2015). Regarding trends, regionalized indices are shown in Fig. 6: dry

periods (negative values of SPI) occurred during the 1960s in most of the regions, followed by decades with prevalence of humid periods. This result is in agreement with the trends in the annual number of dry days (Rivera et al. 2013) and with the negative trend in the percentage of stations with drought conditions (about −3% every 10 yr) (Cavalcanti et al. 2015). An unprecedented increase in climate variability was observed during the last century in northeastern Argentina, shifting from dryness to wetness in decadal time steps (e.g. Piovano et al. 2002, Lovino et al. 2014).
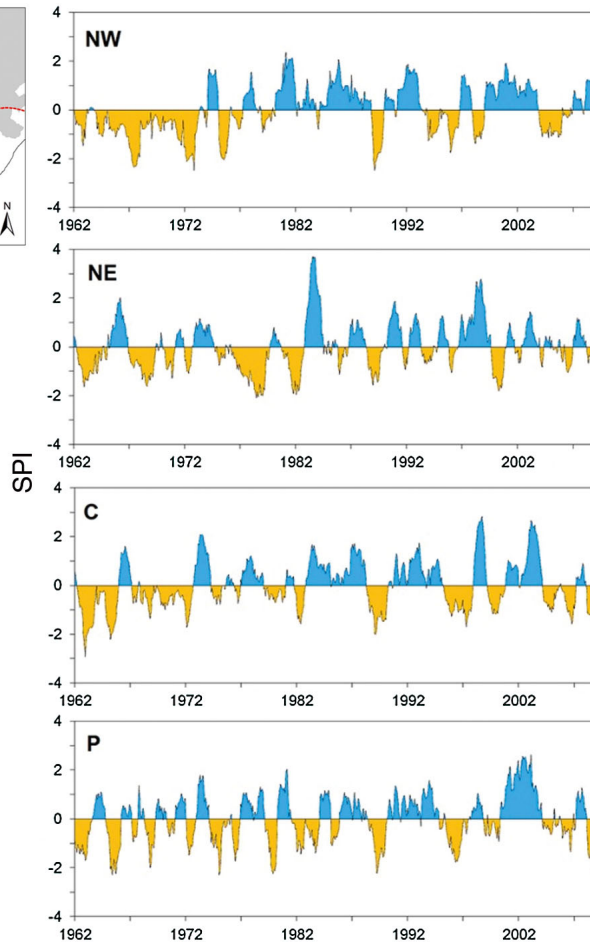
Fig. 6. Spatial regionalization of time series and the temporal evolution of the regional standardized precipitation index (SPI) patterns. Dashed red lines in the left panel separate the four climatic regions of the LPB (shaded): north-west (NW), north-east (NE), central (C) and Pampas (P). The LPB region is shaded

3.2. Remote forcings

Precipitation anomalies in the LPB are influenced by teleconnections triggered by atmospheric circulation disturbances, which produce Rossby wave trains, connecting anomalous convection occurring in distant regions to anomalous conditions over the LPB (e.g. Berbery et al. 1992, Vera et al. 2004). The source regions for teleconnections are the anomalous warm pools in the tropical Pacific and Indian oceans (e.g. Grimm et al. 2000, Barros & Silvestri 2002, Báez Benitez & Monte Domecq 2014, Cherchi et al. 2014). In addition, precipitation anomalies in the LPB are influenced by mechanisms associated with SST anomalies in the tropical Atlantic (e.g. Diaz et al. 1998, Doyle & Barros 2002, Berri & Bertossa 2004, among others). Interdecadal precipitation variability is influenced by SST variations over all the oceans (Grimm & Saboia 2015).

In particular, precipitation extremes in the LPB are also related to these phenomena, some of which were studied during the project (see e.g. Grimm & Tedeschi 2009, Barreiro 2010, Barreiro et al. 2014, Cavalcanti et al. 2015, Grimm et al. 2016, this Special). Typically, a low-pressure anomaly located on the southwest Atlantic off the coast of Patagonia, originating from Rossby-wave activity, propagates northward, led by a cold front causing rainfall extremes in SESA through the uplifting of warmer air masses (Boers et al. 2014a). When the frontal system propagates northeastward through the LPB, the low-pressure anomaly merges with the Northwestern Argentinean Low, affecting the warm, moist low-level flow from the north reaching the LPB (Seluchi & Saulo 2003, Arraut & Barbosa 2009).

The most prominent summertime climate pattern in the LPB is the South American Monsoon System (SAMS; Zhou & Lau 1998), in which climate variability at various timescales is associated with a dipole-like pattern with active nodes in the northern and southern LPB (e.g. Marengo et al. 2012 and references therein). The relationship between this dipole with rainfall extremes is discussed by Cavalcanti et al. (2015). As suggested in previous works (e.g. Silva & Berbery 2006), the precipitation extremes dipole is related to propagations of planetary wavetrains (Fig. 3 in Cavalcanti et al. 2015), and it is the most prominent mode of an oscillatory pattern that extends over the entire continent (Boers et al. 2014b).

The major source for the planetary wave propagation is located in the tropical Pacific Ocean, from where ENSO events influence the rainfall and extremes in the LPB, especially from October (0) through April (+1), with a small interruption in January (+1) of El Niño episodes (Grimm & Tedeschi 2009). The indexes (0) and (+) refer to the starting year of an episode and the following year, respectively. In November (0), the ENSO impact on the La Plata Basin is stronger than in other months, in terms of relative precipitation intensity, while in austral autumn (+), some of the strongest ENSO-related floods occurred in the basin. Nevertheless, each type of ENSO event (e.g. Ashok et al. 2007, Kao & Yu 2009)

is associated with anomalous tropospheric circulation in different regions (e.g. Larkin & Harrison 2005, Hill et al. 2011). Tedeschi et al. (2013) showed that during the Canonical El Niño (La Niña) events, precipitation increases (decreases) are expected in the LPB during all seasons. Results on the frequency of extreme precipitation events are generally consistent with the behavior of total seasonal rainfall (Tedeschi et al. 2014). However, the areas affected by significant sensitivity in the frequency of extreme events are more extensive than those affected by significant sensitivity in the mean precipitation (i.e. there is more sensitivity to ENSO events in the extremes of the precipitation distribution; Grimm & Tedeschi 2009). During ENSO Modoki events and during the austral spring, however, a dipole between the northern and southern LPB regions changes sign from El Niño to La Niña, indicating an increased (decreased) number of wet extremes in the southern (northern) LPB region during El Niño, and the opposite during La Niña (Fig. 4 in Cavalcanti et al. 2015).

Lastly, Robledo et al. (2013) focused on the role of the tropical oceans in modulating the intensity of extreme daily precipitation (DIER75 index; see Table 1) over the Argentinean sector of the LPB. Their results are in agreement with findings in the abovementioned studies. Niño3.4-like anomalies (warm tropical Atlantic anomalies) are related to enhanced (decreased) DIER75 index values year-round except in winter. In spring, the Indian Ocean Dipole Index (Saji et al. 1999) also plays a role favoring positive anomalies of DIER75. However, it should be highlighted that the relationship between the SST anomalies in tropical oceans and the DIER75 index in NE Argentina is modulated by decadal variations.

3.3. Projections and uncertainties

It is generally accepted that the range of validity of the physical parameterizations for an RCM depends on the region and the season (e.g. Seth & Rojas 2003 for South America). Solman & Pessacg (2012) evaluated the sensitivity of the MM5 RCM to several combinations of cumulus and planetary boundary layer parameterizations to simulate the frequency and intensity of extreme precipitation events during November–December 1986. This period was characterized by a clear see-saw pattern, associated with one of the main modes of intraseasonal variability over the region (Díaz & Aceituno 2003). November 1986 was characterized by wet conditions and higher frequencies of moderate and heavy precipitation

events compared to December 1986, when higher frequencies of occurrence of light precipitation events occurred. Although experiments were able to capture the main features of the intraseasonal variability, they overestimate the frequency of light precipitation events and underestimate the frequency of moderate rainy events. The heavy precipitation events were fairly well simulated. The authors found that the underestimation of rainfall is generally related to the underestimation of the amount of rainfall of individual rainy events, and concluded that improvements in the cumulus schemes are necessary to represent the intensity of rainfall more accurately.

In the climate change context, results from CLARIS-LPB RCMs nested into different GCMs under the SRES A1B emission scenario were documented by Sánchez et al. (2015), but focusing on the mean climate conditions (extremes were not considered). Based on daily defined indices, Sörensson et al. (2010a) explored the potential effects of climate change on the extreme rainfall events, using a single RCM (RCA3-E) nested into ECHAM5/MPI-OM GCM, forced by the A1B emission scenario. The extreme precipitation risks (defined as the probability of exceeding the present-day 95th percentile of precipitation) increased in the LPB during all seasons but spring. The increase in intense precipitation for the LPB is also reflected in the positive response of the maximum 5 d precipitation (Rx5day index; see Table 1). This index indicates the risk of flooding (Christensen & Christensen 2003); therefore, its increase over the LPB of up to 50 % probably implies a large impact for the late 21st century. Results presented by Sörensson et al. (2010a) suggest that seasonal precipitation and heavy rainfall events tend to increase by the end of this century, especially in the northern part of the LPB, although with no clear pattern of change for the dry spell duration (CDD index; see Table 1). In contrast, results by Zaninelli (2015), based on a subset of CLARIS-LPB RCMs, and applying a gridpoint- and model-dependent definition to determine dry spells, showed that the climate in the LPB is projected toward a wetter regime together with consistent changes in other measures of dryness (more frequent and shorter dry spells with respect to the present climate). Although this result is in agreement with findings by Menéndez et al. (2016), it must be taken with caution because RCMs still have inaccuracies simulating extremes in precipitation on both daily (Carril et al. 2012) and monthly scales (Menéndez et al. 2010a).

Projections made with coupled GCMs for the last period of the 21st century indicate a trend toward

more frequent extreme (severe and moderate) wet months in the northern (southern) sector of the LPB and less frequent extreme and severe dry months. In particular, the impact of ENSO on extreme precipitation events in the LPB is projected to increase in the future (Cavalcanti et al. 2015).

Other authors have explored the future evolution of droughts. Results based on a CMIP5 multi-model ensemble forced by the RCP4.5 emission scenario
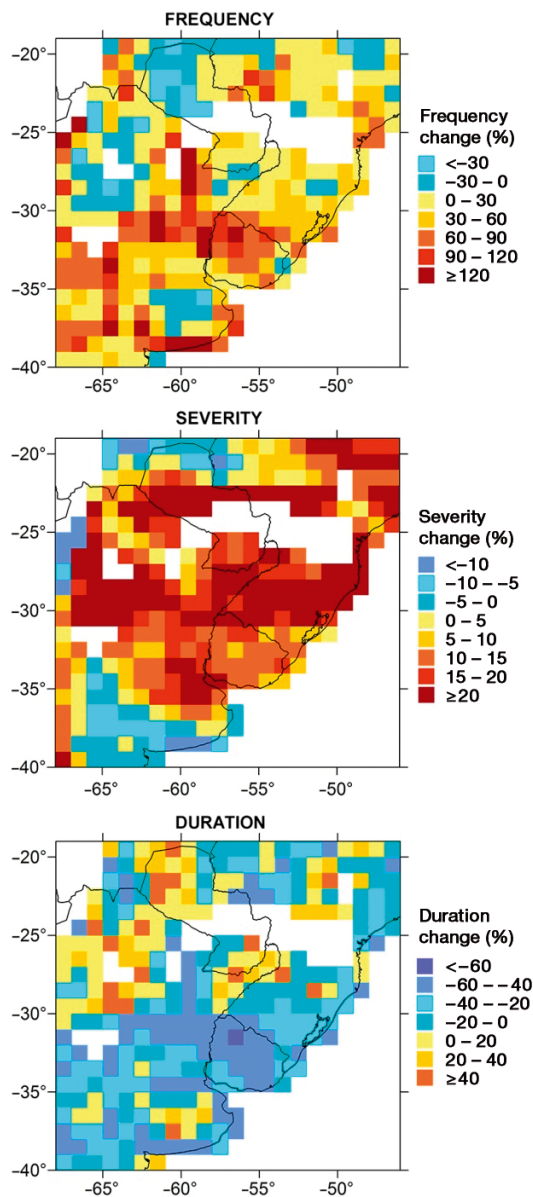

Fig. 7. Mean changes in long-term (12 mo) SPI drought characteristics projected by the multi-model ensemble for the period 2011–2040 relative to the 1979–2008 baseline along the RCP4.5 scenario. Severity is defined as the average SPI value during the drought events. The white areas over the continent indicate non-significant fits to the gamma probability distribution. Figure adapted from Penalba & Rivera (2013)

(Fig. 7) indicate that in general, long-term droughts in the LPB will be more frequent in the medium-term future (2011–2040 with respect to the 1979–2008 period) but also shorter and more severe. Nevertheless, there are some specific areas where lower drought frequencies but longer durations are projected, such as north-western Argentina, western Paraguay and parts of eastern Argentina. Changes in the same sense are also documented for the distant future (2071–2100) and for a more extreme emission scenario (RCP8.5) (Penalba & Rivera 2013).

4. THE INFLUENCE OF LOCAL PROCESSES

In addition to remote influences, local features can also contribute to the occurrence of extremes in the LPB. In that sense, the study of the soil moisture–atmosphere interactions over the LPB is a relatively new issue. During the last decade, some studies have focused on the importance of soil moisture–atmosphere interactions during the wet season of the SAMS (Fu & Li 2004, Li & Fu 2004, Grimm et al. 2007, Collini et al. 2008, Sörensson et al. 2010b, Barreiro & Díaz 2011). These publications showed that the soil moisture state at the onset phase of the SAMS is important for the development of the monsoon. Koster et al. (2009b) suggested that the regions where soil moisture could influence precipitation are transition zones between dry and humid climates. The southern sector of the LPB (a region with high interannual variability of water content in the soil, located between the arid diagonal of South America [Garleff et al. 1991] and the humid region of southern Brazil) is a transition zone and, therefore, a relevant region for studying these phenomena. The coupling of soil moisture with precipitation (i.e. the quantitative effect of one variable on the other in a 1-way interaction) could be quantified as described by Sörensson & Menéndez (2011). The coupling strength of soil moisture with precipitation was relatively high in the LPB (with values of memory larger than 20 d within part of the hot spot) during the onset and mature phase of the SAMS (see e.g. Sörensson et al. 2010b, Ruscica et al. 2014a). Humid areas, such as the Amazon, are not regions of high coupling between soil and rainfall because evapotranspiration is limited by the energy required to evaporate water (i.e. soil moisture is not a limiting factor).

These results motivated further studies on coupling strength and its relation to extreme events, realized within CLARIS-LPB. Sörensson & Menéndez (2011) calculated the coupling strength index between soil

moisture and both precipitation and evapotranspiration in austral summer (details of the index are in their section 3). The LPB was the most important hot spot of soil moisture–evapotranspiration (SM → ET) coupling in the continent, both in extension and magnitude (their Fig. 5). The soil moisture–precipitation coupling (SM → PP) was high in a smaller region within this hotspot, in the central part of the southern LPB (their Fig. 4). As a measure of extreme precipitation, the fraction of the 95th percentile rainfall contribution to total rainfall (Extreme Precipitation Index) was evaluated. The authors suggested that the interaction between soil moisture and atmospheric processes has some impact on this index over the LPB.

Regarding the memory of soil moisture over LPB in summer, model results by Ruscica et al. (2014a) indicated that regions with high SM → ET coupling are characterized by low root-zone soil moisture memory (a few days). However, there are also smaller areas contained in the main hot spot in which memory and SM → ET and SM → PP couplings have relatively high values of coupling strength index (Fig. 8b in

Ruscica et al. 2014a). These results suggest that the soil–atmosphere interaction is a factor that can affect rainfall and contribute to their predictability in areas with high memory. However, the relationship between soil moisture and extreme precipitation events is not *a priori* evident. This issue clearly requires further investigation.

Moreover, Koster et al. (2009a) have shown that regions in which evaporation is controlled by soil moisture availability could experience drought-induced warming. Therefore, the LPB is one of the regions in the world where soil moisture–evaporation feedbacks could have an impact on heat waves. Preliminary results give evidence that soil moisture variability could be playing a role in modulating the temperature field. We highlight differences in daily TX probability distribution function (Fig. 8) between 2 samples: TX for all DJF and TX for dry DJF. The period considered is 1961–2000, and statistics of TX are estimated from TNCR. Seasonal precipitation is used as a proxy for the summertime soil moisture (see discussion in Section 3a in Koster et al. 2009b). Dry summers are
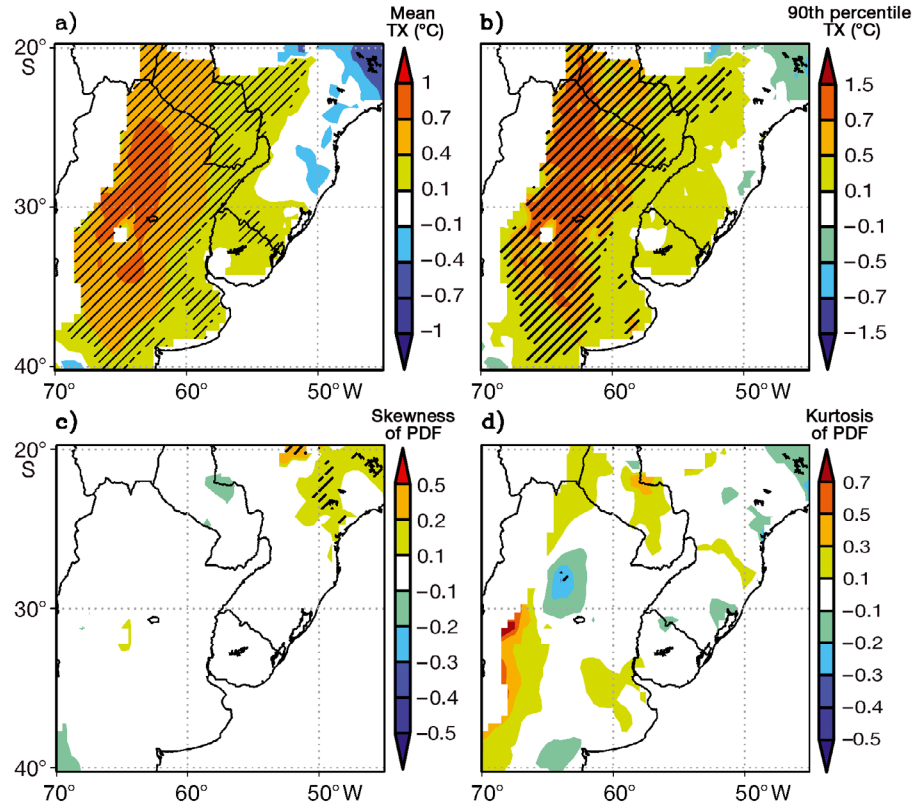

Fig. 8. Statistics estimated from 2 samples: TX DFJ and TX of 'dry' DJF. Top panels are for (a) the mean value and (b) the 90th percentile of TX. Bottom panels are for the (c) skewness and (d) kurtosis of the probability density function (PDF, dimensionless). Period is 1961–2000, and panels illustrate differences (statistics of TX minus statistics of TX dry summers). Striped areas indicate significant difference, according to the following test. For each grid point, the 2 distributions (TX from all years and TX from dry years) were concatenated, shuffled randomly, and redrawn 500 times; differences in those statistics were estimated to be significant when they were greater (lower) than the 95 % (5 %) quantile of the corresponding distribution of differences

defined as those during which the SPI (precipitation accumulated during 3 mo) is <–0.5 (precipitation datasets from Jones et al. 2013). According to this definition, clusterization results in 18 dry summers. The top panels in Fig. 8 illustrate the leading-order impact of the summertime soil moisture variability in TX, in the sense that soil moisture–atmosphere interactions contribute to the variability in summertime surface temperature. Mean values of TX significantly increase during dry summers (Fig. 8a), with a hotspot in central-north Argentina and Paraguay, where TX exhibits large interannual variability (Zaninelli et al. 2015b). In contrast, mean values of TX appear to change only modestly over Uruguay (and over the south of Brazil where small cooling is also noted). Fig. 8b shows that the shift in mean values of TX over a vast region of LPB is associated with increases in the 90th percentile of TX. High-order moments further reveal that soil moisture dynamics have an impact on shifting the statistical distribution of TX toward higher values during dry summers but not on reshaping the probability distribution function (i.e. distributions do not manifest differences in either skewness or kurtosis; Fig. 8c,d). Although further research is still needed, preliminary results suggest that soil moisture–atmosphere interactions in the LPB do not particularly impact the tails of daily TX distribution, but influence both the mean values and the threshold of extremes in TX.

We know that land-use changes modulate the effects of large-scale variability locally (Pielke et al. 2011). Moreover, warming trends have sometimes been attributed to changes in land use, including the development of large cities, such as São Paulo and Rio de Janeiro (e.g. Marengo 2001, 2003). During the last decades, large areas in both the LPB and in the Argentinean Pampas have been extensively deforested; however, there are very few studies that address this topic (and none of them are related to extremes). During the project, the potential impacts of land-use changes over the LPB and the Argentinean Pampas were examined using model simulations forced by different idealized land-use scenarios representing agricultural expansion, reforestation and desertification. The results of Pessacg & Solman (2012) showed significant warming and drying when forests are replaced by bare soils due to an increase in the net radiation budget and a reduction in the latent heat flux. However, the replacement of forests by crops resulted in a decrease in both the net radiation budget at surface and the latent and sensible heat fluxes, leading to significant cooling over central and eastern Argentina. Lastly, the modification of

the actual land cover to crops produced a cooling and wetting over northern Argentina, Paraguay and parts of Bolivia due to a decrease in both the net radiation budget and the sensible heat flux and an increase in the latent heat flux.

5. FUTURE CHALLENGES

This paper reviews what we have learned during the CLARIS-LPB project regarding extremes in the LPB. The conceptual framework in which extremes are embedded is basically given by ocean–soil–atmosphere interactions and their variabilities at different timescales. That is, extremes are the result of multiple interactions, including regional processes and remote forcings. Regarding the regional processes, we highlighted soil moisture–atmosphere interactions, the impact of the land use and the extratropical systems. Tropical and extratropical oceans force regional anomalies and atmospheric large-scale circulation patterns, and could leave imprints in regional variability (see e.g. Barreiro 2010). Moreover, forcing sources are affected by variability on several time scales (including low-frequency modulation), which is then transferred to regional scales, affecting extremes.

Despite the significant progress that was made on describing and understanding extremes, many challenges remain, and some gaps have been identified. At regional scales, additional efforts to better understand the link between soil characteristics and the atmosphere are needed (e.g. Ruscica et al. 2014b, Menéndez et al. 2016), and a gap was found in studies that relate extremes with changes in land use. Further research on the relationship between the development of cut-off lows and the occurrence of extreme precipitation events is also needed (Godoy et al. 2011). In addition, trends and variability in the diurnal cycle of precipitation, with implications for the frequency of occurrence and intensity of extremes in the LPB, need to be studied. This research will be challenging because both models and reanalysis had deficiencies with regard to appropriately representing the diurnal cycle of precipitation. At the hemispheric scale, focus should be on improved description and understanding of the role that the boundary forcing parameters play (ocean and land) and on the large-scale circulation patterns. The relative importance of the regional feedbacks (including hotspots of coupling strength) versus the large-scale forcings (including seasonal predictability sources) is still largely unknown. Efforts in that

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sense could potentially improve seasonal forecasts of extremes and the implementation of operational early warning systems for agricultural applications.

Some specific research questions that need further investigation include the following: (1) What are the relative roles of local versus remote forcings in the occurrence of extremes (SST, land surface and topography)? Several studies have shown the role of teleconnections and local forcings with regard to extreme precipitation over the LPB, as mentioned in the 'Introduction'. However, the relative role of large-scale versus local forcings has not yet been analyzed. (2) What is the effect of land use on amplifying/decreasing extremes? This is a very important problem to be analyzed in face of Amazon deforestation. There are several studies of land-use effects on precipitation of the region, but there are not studies on such effects over the LPB. (3) What is the role of the interaction processes between the land surface and the atmosphere with regard to precipitation and temperature extremes? The feedbacks related to soil moisture, the memory in the system introduced by its anomalies, and related features such as surface inhomogeneities can influence different types of extremes on different scales (e.g. heat waves, daily precipitation events, etc.). (4) How will extreme events evolve in the future? Several studies have been performed regarding extremes in future projections. However, with global and regional model developments, this projection is an issue that needs continuous analysis. (5) What are the uncertainties in extreme climate change projections? This uncertainty is another issue that deserves continuous analysis, because the models are improving, and it is expected that the uncertainties will be reduced in the new-generation models. Obstacles that may affect progress toward answering questions mainly include (1) the lack of high-resolution and high-frequency data, and (2) the current state of the art of models and parameterizations. Combining an intensive monitoring and modeling strategy should be considered a challenge for the regional community. Improvements in monitoring and modeling will also result in reduced uncertainties and increased confidence in regional climate change projections.

Finally, we have noticed that recent research efforts are mainly focused on a better description and understanding of extremes in temperature and precipitation. On one hand, we know that a vast area on the northern coast of Buenos Aires province (including Buenos Aires City and surrounding areas) is periodically affected by storm surges in the Río de la Plata estuary (Escobar et al. 2004), flooding the pop-

ulated coast. There is a need for further research on winds and wind-wave extremes (Dragani et al. 2013a,b) (which were not addressed at all in CLARIS-LPB) to move toward an integrated forecast system of the interface Rio de la Plata-atmosphere. On the other hand, we have noticed that attribution of extreme events to the emissions of greenhouse gases from anthropogenic sources (Stott et al. 2013) has been little explored in the LPB (Hannart et al. 2015). Among the limitations to develop this research line in the LPB, we highlight (1) the state-of-knowledge of physical causes of extremes and how extremes may link to anthropogenic emissions and (2) the reliability of the simulations to capture both the extreme events and the real-world predictability (e.g. Seneviratne et al. 2012, Peterson et al. 2012).

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