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Precipitation changes in the South American Altiplano since 1300 AD reconstructed by tree-rings

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Abstract

During the second half of the 20th century, the Central Andes has experienced significant climatic and environmental changes characterized by a persistent warming trend, an increase in elevation of the 0 °C isotherm, and a sustained shrinkage of glaciers. These changes have occurred in conjunction with a steady growing demand for water resources. Given the short span of instrumental hidroclimatic records in this region, longer records are needed to understand the nature of climate variability and improve the predictability of precipitation, a key factor modulating the socio-economic development in the South American Altiplano and the adjacent arid lowlands. In this study we present the first quasi-millennial, tree-ring based precipitation reconstruction for the South American Altiplano. This annual (November–October) precipitation reconstruction is based on *Polylepis tarapacana* tree-ring series and represents the closest dendroclimatological record to the Equator in South America. This high-resolution reconstruction covers the past 707 yr and provides a unique record to characterize the occurrence of extreme events and consistent oscillations in precipitation, as well as to check the spatial and temporal stabilities of the teleconnections between rainfall in the Altiplano and hemispheric forcings such as El Niño-Southern Oscillation. Since the 1930s up to present a persistent negative trend in precipitation is recorded in the reconstruction, with the three driest years since 1300 AD occurring in the last 70 yr. The reconstruction contains a clear ENSO-like pattern at interannual to multicentennial time scales which determines inter-hemispheric linkages between our reconstruction and other precipitation-sensitive records modulated by ENSO in North America. Our reconstruction points out that century-scale dry periods are a recurrent feature in the Altiplano climate, and that the potential coupling of natural and anthropogenic-induced droughts in the future would have a severe impact on current socio-economical activities in the region. Water resource managers must anticipate these changes to adapt for future climate change, reduce vulnerability and provide water equitably to all users.

1 Introduction

Water availability is the main limitation for the socio-economical development of many regions in the world. In addition, fluctuations in water supply have large impacts on natural ecosystem productivity (Viviroli et al., 2003; Messerli et al., 2004). These affirmations are certainly valid for high-altitude regions in the tropics such as the South American Altiplano (Messerli et al., 1997). This semi-arid plateau with a mean elevation of 4000 m in the Central Andes (15–24° S) has been the physical environment for the settlement of many native communities who have inhabited the region for thousands of years. Historically, human activities in the Altiplano have been strongly modulated by variations in climate, particularly water availability (Tandeter, 1991; Binford et al., 1997; Núñez et al., 2002). Agriculture in the Altiplano region is extremely susceptible to drought conditions with the consequent yields reductions (García et al., 2003, 2007). Episodic summer rainfalls represents the major source of water for human and agriculture, the stream flows, and the recharge of the underground aquifers in the Central and Southern Altiplano as well as in the adjacent arid lowlands of Southern Bolivia, Northern Chile and Northwestern Argentina (Garreaud et al., 2009).

Major droughts across this region are known to convey severe economic and social impacts larger than any other type of natural disaster threatening rural livelihood (Gil Montero and Villalba, 2005). The yields of common crops such as potato and quinoa (*Chenopodium quinoa*) are strongly modulated by precipitation, pointing toward persistent droughts as the main cause of economic stress in the region (Mattos and Crespo, 2000; García et al., 2003, 2007). For instance, the severe drought of 1998 provided a comprehensive view of the adverse impacts of dry events on the socio-economical activities in the rural Altiplano when 60 % of the local camelid livestock "*Llamas*" and other domestic animals died (Gil Montero and Villalba, 2005). Small streams of water disappeared and people competed with animals for water resource (Gil Montero and Villalba, 2005).

Across the Southern Altiplano, summer rainfalls represent more than 90 % of the total annual precipitation (Garreaud et al., 2003; Vuille and Keimig, 2004). Recent studies based on instrumental records have documented important variations in the Altiplano climate together with a positive warming trend since the second half of the 20th century (Vuille and Bradley, 2000; Vuille et al., 2003; Trenberth et al., 2007). This regional increase in temperature has been related to an elevation of the 0 °C isotherm altitude (Vuille et al., 2008; Carrasco et al., 2008), a rapid and likely unprecedented melting of ice caps (Thompson et al., 2003), and sustained shrinking of glaciers (Francou et al., 2003; Coudrian et al., 2005; Jomelli et al., 2011). All these environmental changes have occurred in conjunction with a growing demand of water resources as a result of population increase and the rapidly expansion of the mining industry in the Andean region (Messerli et al., 1997; COCHILCO, 2007). In addition, recent model simulations projected a reduction of precipitation in the Central Andes curtailing water resources availability (Bradley et al., 2006; Urrutia and Vuille, 2009; Minvielle and Garreaud, 2011).

Present knowledge of climate variability in the last 1000 yr in the Altiplano is severely limited by the low number of high-resolution palaeoclimatic records in the tropical Andes, remaining as a research topic of high priority for the paleoclimatology in South America (Jansen et al., 2007; Villalba et al., 2009). The lack of information on past climate variations constrains the possibility of validating climate models used to predict future precipitation trends (Randall et al., 2007; Lohmann, 2008). This is a key issue for developing mitigation and/or adaptation strategies for future climate change scenarios in the region. Instrumental precipitation records for the Altiplano are generally short, fragmentary and non-homogeneous, making them inadequate for the development of a baseline-understanding of long-term trends (Vuille et al., 2003). Therefore we need longer precipitation records to complement the current limited nature of the instrumental registries to properly understand how interannual modes of climate variability have evolved under changes in long-term background conditions.

In contrast to the extratropical Andes where tree-ring studies have yielded more than a hundred chronologies and contributed with over 30 climate reconstructions (Boninsegna et al., 2009), just during the past few years suitable extremely moisture sensitive tree-ring chronologies of *Polylepis tarapacana* (*Queñoa*) have been developed in the South American Altiplano (Morales et al., 2004; Solíz et al., 2009). The current possibility to develop an annually resolved tree-ring reconstruction of precipitation for the Altiplano emerges as a great opportunity to enhance our knowledge about past and present climate variability in the tropical Andes region. This record would contribute to fill a significant geographic gap in the present coverage of dendroclimatological reconstructions within the Andes.

The main goal of our study was to develop an exactly-dated, annually-resolved precipitation reconstruction for the South American Altiplano during the past 700 yr from recently-developed *P. tarapacana* tree-ring chronologies. We analyzed this quasi-millennial paleoclimatic record to describe its temporal evolution, the recurrence of extreme events, the presence of persistent cycles and the relationships with hemispheric climate forcings such as El Niño-Southern Oscillation (ENSO). Our contribution expands the tree-ring based precipitation reconstructions in South America to the tropical Andes and provides the first annual resolution paleoclimatic reconstruction for rainfall in the Altiplano.

2 Setting and climate of the South American Altiplano

The tropical Central Andes represent a formidable obstacle for the atmospheric circulation over South America generating two contrasting regions, the tropical humid lowlands to the east and the Pacific coastal deserts to the west (Garreaud et al., 2003). A particular physiographic features in the Central Andes is the Altiplano, a high-elevation, inter-mountain plateau extending from 15 to 24° S (Fig. 1a). Precipitation across the Altiplano decreases from ~500 mm in the northeast transition to the Amazon basin to <200 mm in the southwest sector adjacent to the Atacama Desert. More than

80 % of total annual rainfalls occur during the austral summer (December–February) (Vuille and Keimig, 2004). The episodic precipitation of convective type is related to the upper-air circulation with an easterly (westerly) zonal flow favoring the occurrence of wet (dry) events (Garreaud et al., 2003). The extreme seasonality of the precipitation is associated with the onset and decay of the Bolivian High, an upper-level high-pressure cell that develops over the Central Andes in response to the latent heat release by the summer deep convection over the Amazon Basin (Lenters and Cook, 1997). Wet intervals are related to a stronger and southward-displaced Bolivian High, which allows the expansion of upper-air easterly flow and the ingression over the Altiplano of the moisture influx from the Amazon basin (Lenters and Cook, 1997; Garreaud et al., 2009).

Year-to-year variability in precipitation is mainly related to changes in the mean zonal wind over the Altiplano largely modulated by sea-surface temperature (SST) across the tropical Pacific Ocean (Vuille et al., 2000; Garreaud and Aceituno, 2001; Bradley et al., 2003). During the warm (cold) phase of the El Niño-Southern Oscillation (ENSO), the Altiplano climate is dry (wet) (Aceituno, 1988; Lenters and Cook, 1999; Vuille, 1999; Vuille et al., 2000). Wet summers are related to a cooling of the central and eastern sectors of the tropical Pacific (La Niña event). Weaker upper-elevation Westerlies during wet episodes facilitates the ingression of the wet easterly flow, transporting humid air masses from the Amazon basin. In contrast, dry summers associated with El Niño events in the tropical Pacific, are characterized by the dominance of westerly flows and the concurrent blocking of the humid air penetration from the east (Vuille, 1999; Garreaud et al., 2003).

3 Data and methods

3.1 Precipitation and tree-ring data

Monthly precipitation records for the Altiplano were obtained from the *Servicio Nacional de Meteorología e Hidrología* in Bolivia (SENAMHI) and the *Dirección General de Aguas* in Chile (DGA). The 17 precipitation stations used in this study are located from 17 to 22° S and range in elevation from 3545 to 4600 m (Fig. 1a; Table 1). We develop a regional monthly precipitation record based on these 17 individual records. Few instrumental records extend prior to 1950 and they are not evenly distributed across the Altiplano. In consequence, a robust and spatially representative record of regional precipitation was built starting in 1961. Total annual precipitation across the Altiplano decreases in a northeast-southwest direction, however the interannual variability in rainfall shows a uniform pattern across the region (Garreaud et al., 2003). To minimize the influences of weather stations with higher rainfall on the regional mean, our regional precipitation record was developed by averaging the precipitation anomalies (expressed as percentages) with respect to the common interval 1982–2000.

The world's highest elevation woodlands of *Polylepis tarapacana* (Rosaceae) in the Altiplano, represents a remarkable resource to develop reliable high-resolution paleoclimate reconstructions in the tropical Andes (Argollo et al., 2004; Morales et al., 2004; Boninsegna et al., 2009; Christie et al., 2009; Solíz et al., 2009). *P. tarapacana* is a unique tree species that reaches over 700 yr old and grows along the Central Andes from 16 to 23° S between 4000 to 5200 m (Fig. 1b; Braun, 1997). Previous studies show that the radial growth of the *P. tarapacana* is strongly related to interannual variations in summer precipitation. At regional scale, tree-growth patterns resemble the spatio-temporal variations of precipitation across the Altiplano, highlighting the great potential of this species to provide precipitation reconstructions with highly significant hindcast skills (Solíz et al., 2009).

In this study, seven regional chronologies from *P. tarapacana* were developed by merging previous single-site records, incorporating new chronologies as well as updating and extending back in time previous records (Argollo et al., 2004; Christie et al., 2009; Solíz et al., 2009). New tree-ring sites were sampled on steep, rocky and xeric environments in the western flank of the Andean Western Cordillera (Fig. 1; Table 2). Due to the twisted stems and the eccentric radial growth patterns of *P. tarapacana*, cross-sections were collected from branches of living trees and sub-fossil wood that have remained on the ground-surface for several centuries due to the dry and cold climate. Wood samples were mounted and sanded following standard dendrochronological techniques (Stokes and Smiley, 1968). For the dating purposes, we followed the Schulman's convention (1956) for the Southern Hemisphere which assigns to each tree ring the date of the year in which radial growth started. Tree-rings were visually cross-dated and measured with a binocular stereoscope with 0.001 mm precision. Precise dating for the floating TER chronology (composed by subfossil woods) was established by cross-dating the individual samples with nearby (GUA, QUE and FSA) chronologies. To assess the quality of the cross-dating and identify measurement errors we utilized the computer program COFECHA (Holmes, 1983).

Interannual variations of *P. tarapacana* growth show consistent spatial similarities across the Altiplano. Previous studies associated the similarity between records with the occurrence of a common pattern of precipitation in the region (Solíz et al., 2009). Based on these observations, a regional, robustly-replicated tree-ring chronology was built by assembling in a single record the 353 tree-ring series from the 7 sites listed in Table 2. An indication of the common signal between the seven site chronologies is the highly significant mean correlation coefficient of all possible pairings among them (21) computed over the well-replicated common period 1668–1776 (>8 samples in all sites) ($r = 0.54 \pm 0.02$ standard error, $n = 109$, $P < 0.001$). A principal component analysis of the seven site chronologies over the period 1668–1776 provide similar loadings (0.72 to 0.84) from the seven records to the first principal component (Table 2).

Ring-width measurements were standardized to remove variability in the time series not related to climate such as tree aging or forest disturbances (Cook et al., 1990). To conserve the low-frequency signal in tree-growth we used a conservative method of standardization by fitting negative exponential or linear curves with zero or negative slope to each individual series. The regional tree-ring chronology was calculated by averaging the detrended *P. tarapacana* tree-ring series with a bi-weight robust mean estimation using the ARSTAN program (Cook, 1985). The quality of the tree-ring chronology was tested by the Expressed Population Signal statistic (EPS), which measures the strength of the common signal in a chronology over time and quantifies the degree to which a particular chronology portrays the hypothetically perfect chronology (Cook et al., 1990). To calculate the EPS, we use a 50-yr window with an overlap of 25-yr between adjacent windows. While there is no level of significance for EPS, values above 0.85 are generally accepted as a good level of common signal fidelity between trees, so we used only the portion of the chronology with EPS > 0.85 as a predictor of the precipitation in the reconstruction (Wigley et al., 1984).

3.2 Reconstruction method

Correlation coefficients between the regional standard *P. tarapacana* chronology and monthly variations in regional precipitation were used to define the seasonal precipitation best related to radial growth (Blasing et al., 1984). Total annual precipitation (November to October) was the period best correlated with annual growth. We developed the annual precipitation reconstruction by regressing the regional standard chronology against total November–October precipitation utilizing a Principal Component regression approach (Cook et al., 2007). Predictors for the reconstruction included the regional chronology in all temporal lags significant correlated ($\alpha = 0.05$) with annual precipitation during the 1961–2009 calibration period. This approach for selecting the predictors allows for a 3 yr response to climate in the tree-ring width chronology (Fritts, 1976). While the chronology is not significant correlated at year t, statistically significant correlation with annual precipitation were recorded at lags $t+1$, $t+2$, and $t+3$ ($r = 0.71$, 0.37 and 0.31, respectively; $n = 45$; $P < 0.05$). These three lags were considered as candidate predictors of annual precipitation and entered in a principal component analysis to reduce the number of predictors and enhance the common precipitation signal. Thus, the intercorrelated set of predictors was converted to orthogonal variables reducing the dimension of the regression problem by eliminating the higher-order eigenvectors that explain a small proportion of the variance (Cooley and Lohnes, 1971). The selection criterion for choosing the best reconstruction model was that basing in maximizing the adjusted R^2 (R^2_{adj}) in a stepwise multiple regression procedure (Weisberg, 1985). Given the relative short precipitation record for calibration, the reconstruction model was developed using the "leave-one-out" cross-validation procedure (Michaelsen, 1987; Meko, 1997). In this approach each observation is successively withheld, a model is estimated on the remaining observations, and a prediction is made for the omitted observation. At the end of this procedure, the time series of predicted values assembled from the deleted observations is compared with the observed predictors to compute the validation statistics of model accuracy and error. The goodness of fit between observed and predicted precipitation values was tested based on the proportion of variance explained by the regression (R^2_{adj}), the F-value of the regression, the linear trend and the normality of the regression residuals, and the autocorrelation in the residuals measured by the significance of the linear trend and the Durbin-Watson test (Draper and Smith, 1981). As an additional measure of regression accuracy we also computed the Reduction of Error (RE) statistic over the verification period (Gordon, 1982), as well as the Root Mean Square Error (RMSE) statistic as a measure of inherent uncertainties in the reconstruction (Weisberg, 1985).

3.3 ENSO, spectral properties and temporal evolution of the reconstructed precipitation

It is widely accepted that ENSO has a strong role modulating precipitation variability in the South American Altiplano (Vuille et al., 2000; Garreaud et al., 2009). Therefore, we expect that our reconstruction will show a strong ENSO signal. To determine the relationships between our reconstruction and ENSO, we estimated the spatial correlation pattern between the reconstructed annual (November–October) precipitation and the mean annual SST (November–October; $2.5^\circ \times 2.5^\circ$ gridded cell) from the NCEP reanalysis global dataset (Kistler et al., 2001). In addition, the relationships in the time frequency space between the reconstructed precipitation and ENSO were assessed using two cross-spectral techniques. Similarities in the temporal evolution of the reconstructed precipitation and the mean November–October Niño 3.4 SST (N3.4) were estimated using cross-singular spectral analysis (SSA; Vautard and Ghil, 1989) and a wavelet coherence analyses (WTC; Grinstead et al., 2004). The SSA detects and extracts the main oscillatory modes of a time series over time, whereas the WTC analysis identifies regions in the time frequency space where the two series co-vary. The WTC detects phase relationships between series and assess the statistical significance against a red noise background using Monte Carlo methods. After assessing the spectral relationships between the precipitation reconstruction and instrumental ENSO, we determined the dominant oscillatory modes of the precipitation reconstruction along the reconstructed 1300–2006 period by performing a continuous Wavelet Transform analysis (WT; Torrence and Compo, 1998). To assess the temporal relationship between the spectral oscillations of our precipitation reconstruction and ENSO across its full length, we used a cross-wavelet transform analysis (XWT) between the Altiplano precipitation and a well-know independent ENSO proxy represented by the first principal component time series of the North American Drought Atlas (NADA) during the 1300–2002 period (Cook et al., 2004; Li et al., 2011). Finally, to examine the relationships between the significant regime shifts and the interannual and low-frequency variability of the precipitation reconstruction, we compared over the entire 1300–2006 period the regime shifts in the mean detected using the Rodionov (2004) method with a window length of 25 yr, the variance in moving windows of 25-yr, and a cubic smoothing spline that reduce 50 % of the variance in a sine wave of 35 yr.

4 Results and discussion

4.1 Tree-ring chronology and calibration of precipitation reconstruction model

Here, we report on the development of an annually-resolved, moisture-sensitive chronology from tree ring widths in the South American Altiplano (Table 2). The record, covering the past 707 yr, starts in 1226 AD but is well replicate for the period 1300–2009 (> than 10 series and EPS > 0.85). The chronology is based on ~ 87 896 annual ring measurements from more than 350 tree-ring series (Table 2). Chronology statistics show high series intercorrelation ($r = 0.54$), a clear indication of the strong internal coherence in the regional record. Additionally, the mean expressed population signal (EPS = 0.95) also indicate a good level of common signal fidelity between trees.

Due to the highest significant correlation between tree growth and November to October precipitation, we used this period as our target instrumental series (1961–2006) to be modeled back in time using the *P. tarapacana* regional chronology. Although at a lag $t = 0$ the correlation coefficient is not significant, correlations with annual precipitation are statistically significant at lags $t+1$, $t+2$, and $t+3$ ($r = 0.71$, 0.37 and 0.31, respectively; $P < 0.05$), corroborating previous studies that showed a persistent influence of previous year precipitation on *P. tarapacana* radial growth (Argollo et al., 2004; Morales et al., 2004; Solíz et al., 2009). The amplitudes from the first and second principal component were included as predictors of the annual precipitation using a multiple regression. Over the 1961–2006 calibration period, tree-ring indices explain 55 % of the total observed variance in the Altiplano annual precipitation. The statistics used to asses the quality of the regression model indicate highly significant hindcast skills of the model. The strength in the relationship between the observed and estimated precipitation (R^2 adj = 0.55) suggests that the tree-ring reconstruction is quite accurate in representing the instrumental precipitation changes, highlighting the predictive ability of the calibration model as indicated by $F = 26.32$ ($P < 0.001$), a positive RE (0.5), and non-significant autocorrelation and trend of the residuals (DW = 2.4; Fig. 2).

4.2 Precipitation variations in the Altiplano during the last 700 yr

4.2.1 Spatial representation and temporal evolution

To evaluate the spatial representation of the reconstructed annual precipitation, we determined the spatial correlation maps across tropical-subtropical South America between the Altiplano precipitation (both observed and reconstructed) with the $0.5^\circ \times 0.5^\circ$ gridded November–October precipitation from the CRU TS 3.1 dataset (Mitchell and Jones, 2005). The two spatial correlation fields (Fig. 3), estimated over the 1961–2006 common period, show significant correlations across the entire Altiplano, a clear indication of the wide spatial representation of both, observed and reconstructed precipitation records. The spatial correlation fields show that the highest correlation coefficients are concentrated in the north-central section of the Altiplano with decreasing values toward the Southern Altiplano. Although in our reconstruction, the correlation coefficient between the estimated and observed values is pretty high ($r = 0.74$; Fig. 2), correlations between the CRU gridded data and the reconstruction are comparatively lower. This observation is consistent with relatively low correlations between our regional instrumental series and the CRU data. Our findings support previous studies indicating the poor representation of climatic variability by gridded products based on few or even non high-altitude stations in remote areas with complex topographies such as the Central Andes (Garreaud et al., 2009; Tencer et al., 2011).

The annual tree-ring based reconstruction covers the past 707 yr and portrays the interannual to multidecadal scales variations in precipitation across the South American Altiplano since 1300 AD (Fig. 4). Several multidecadal-persistent droughts are observed during the 14th, 16th, 17th, 18th and 20th centuries. Almost the entire 14th century was characterized by below-average precipitation with a single subdecadal humid period between 1300 and 1307. This extreme and severe centennial drought persisted until the beginning of the 15th century (around the 1410s). It has been proposed that the negative impact of this persistent-centennial drought on local agricultural-based societies triggered social conflicts and a period of wars in the Altiplano during the 14th and

15th centuries (Nielsen et al., 2002). A Persistence drought have also been recorded during the 14th century in a Palmer Drought Severity Index (PDSI) field reconstruction mainly based on the Quelccaya, Huascarán and Sajama ice-cores for the Altiplano region (Boucher et al., 2011). During the 15th century our precipitation reconstruction presents milder to wet conditions prevail from the 1410s to the 1520s with a particular humid interval at the end of the 15th century. This relative wet interval was interrupted by a remarkable dry event center in the 1450s. Indeed, the year 1451 appears as one of the tenth driest years in the 700-yr reconstruction. Although the 16th century is characterized by persistent dry conditions, extreme dry events were rare. Just the year 1593 recorded precipitation 60 % below the long-term mean. In contrast to our record, wet conditions during the 16th century have been inferred from the Quelccaya ice core (Thompson et al., 1985, 1986). After the persistent dry conditions prevailing during the 16th century, a remarkable pluvial period during the first decade of the 17th century contrast with a pronounced drought during the 1620s. After that, sustained wet conditions prevail up to the mid 18th century. Cold and wet conditions for the region during the first half of the 18th century have been proposed by Liu et al. (2005) and Thompson et al. (2006). Lichenometry dating of glacier moraines at Cerro Charquini in the Cordillera Real, Bolivia (5392 m; Rabatel et al., 2006), suggest that the Little Ice Age maximum occurred during the second half of the 17th century. This observations are consistent with the persistent wet conditions recorded in our reconstruction during the second half of 17th century lasting until the mid of the 18th century. However, it is important to note that within this long-term wet period, two severe decade-long droughts 1615–1637 and 1684–1696 were recorded. The years 1620–1621 and 1694 appears as the extreme dry years associated with these droughts, respectively.

A long-term drought is registered in our reconstruction during the second half of 18th century (1750–1818) characterized, as the long-term drought recorded during the 16th century, by low interannual precipitation variability. Drier conditions from 1780 to 1820 were also recorded in the PDSI reconstruction for the South America subtropical region (Boucher et al., 2011). Based on historical documents, Gioda and Prieto

(1999) recorded during this period severe droughts in Potosí (South Bolivia), with two extreme dry events lasting consecutively 10- (1777–1786) and 5-yr (1801–1805). After the persistent dry conditions from around 1750 to 1818, a steady increase in precipitation occurred. This long-term persistent wet period lasting from around 1818 to 1887 represents the wettest interval during the past 7-centuries showing four extreme wet events centered in 1820–1822, 1837–1839, 1842–1843, and 1876. Based on the dated moraines chronology from Cerro Charquini, Rabatel et al. (2006) assumed the 19th as a dry period with non-advances of glaciers. However, this long-term pluvial event in our reconstruction is coincident (∼1830–1850) with the highest peak in the *Polylepis* pollen concentrations recorded in a 600 yr ice core registry from the Sajama volcano (Liu et al., 2005). Persistence wet conditions may have favored *Polylepis* forest productivity and expansion and consequently the increase in pollen across the Altiplano (see Gosling et al., 2009). Another important peak in *Polylepis* pollen concentrations also occurred during the wet 1700–1720 reconstructed period (Liu et al., 2005).

The wet conditions of the 19th century continued during the beginning of the 20th century (1906–1929). Since the 1930s, a persistent negative trend in precipitation was recorded up to present. Two severe decadal and multidecadal drought events were registered during 1930–1948 and 1956–2006, respectively. Four of the seven most extreme dry years for the past 707 yr in the Altiplano occurred during the 1940–2006 period (1940, 1982, 1994 and 2006, respectively). Our results are consistent with the drier conditions showed by the PDSI record for the region (Boucher et al., 2011), and the rapid retreat of the tropical Andes glaciers during the second half of the 20th century (Ramirez et al., 2001; Francou et al., 2003; Vuille et al., 2008; Jomelli et al., 2009). The two driest years recorded in the past 700 yr (1940 and 1982) are associated with very strong El Niño events.

4.2.2 Spectral properties, ENSO and temporal regimes

The spatial correlation field between SSTs and the precipitation reconstruction for the interval 1948–2006 shows a clear ENSO-like pattern across the Pacific Ocean (Fig. 5).

Wet years in the Altiplano reconstruction are significantly related to negative anomalies in N3.4 SST (La Niña-like), while dry years correspond to positive tropical Pacific temperature (El Niño-like; Vuille et al., 2000; Garreaud et al., 2009).

We compared in Fig. 6a the main dominant oscillatory modes of the precipitation reconstruction and the instrumental N3.4 SST record over the interval 1872–2006. Major oscillatory waveforms at 8.5–13, 5–6.7 and 3–4.7 yr were identified in both the reconstructed precipitation and the N3.4 SST records. These oscillatory modes explain 28 (19), 13 (29) and 10 (26) % of the total variance in past precipitation (N3.4). For these cycles, the SSA-reconstructed precipitation periodicities follow the dominant oscillation modes in the instrumental N3.4 record in an anti-phase relationship (Fig. 6a). However, some non-coherent changes in the amplitudes of the SSA waveforms for the reconstruction and N3.4 occur. For instance the amplitudes of the oscillatory modes at 3.1–4.7 yr were quite similar during the 1872–1925 period, reduced around 1930–1960 and in anti-phase around the interval 1945–1950 and 1975–1985. This observation is consistent with previous studies indicating a low ENSO activity during the period 1930–1960 (Aceituno and Montecinos, 1993; Torrence and Webster, 1999; Sutton and Hodson, 2003).

A particular remarkable feature in the spectral comparison between the precipitation reconstruction and the N3.4 records is the good agreement, both in amplitude and phase relation at decadal (8.5–13 yr) scales. During the common period (1872–2006), the WTC shows a consisting and stable anti-phase relationship between both records (Fig. 6b). A marked shift in the relative importance of the coherence relation from interannual and decadal band to multidecadal cycles is observed around 1930. In the decadal bands of the WTC we identified a significant spectral coherence between both records around the year 1940, suggesting that the 1940–1941 El Niño event was part of the extreme decadal variability in ENSO. This particular feature is clearly observed in the 8.5–13 yr SSA band (Fig. 6a). This particular El Niño event is associated with the second driest year of the past 707 yr in the Altiplano. Shifts in the ENSO strength, together with changes in the ENSO-Altiplano teleconnection pattern may be related to the lack of spectral coherence between records all the time. Therefore, changes in the coherency and phase between the N3.4 SST and Altiplano precipitation records could be related to the non-stationary ENSO behavior and the spatial variability of this ocean-atmospheric phenomenon.

The WT spectrum shows non-stationary periodicities across the precipitation reconstruction with most significant oscillatory modes concentrated in oscillations < 16 and a single period of multidecadal oscillation centered in 1600 and related to the extremely dry events of 1592–1593 and 1621–1622, separated by approximately 30 yr (Fig. 7a). The WT spectrum shows a large percentage of the variance explained by the classical ENSO band (2 to 8 yr; Deser et al., 2010) with significant increases in the precipitation variability during the 14th century, a relatively period of reduced oscillations between 1450 and 1750 and a small increase in the spectral activity since the mid 18th century (Fig. 7a). These spectral characteristics observed since the mid 15th century also have been described for tree-ring hidroclimatic reconstructions from others region of the Southern Andes where ENSO play an active role modulating the local climate (LeQuesne et al., 2009; Christie et al., 2011). Decadal to multidecadal frequencies in our reconstruction are relatively high since the 17th century.

Finally we compared the spectral oscillations from the Altiplano precipitation reconstruction with those from the NADA during the past 700 yr. According to the XWT analysis, both records share a large proportion of common spectral power within the ENSO bandwidth, suggesting inter-hemispheric linkages between paleoclimatic reconstructions from regions influenced by ENSO (Fig. 7b). Vector directions in the XWT analysis revealed anti-phase relationships between both records, consistent with the well-known negative (positive) relationship between warm conditions in the tropical Pacific SST and precipitation in the Altiplano (Southwest North America) (Vuille et al., 2000; Smith et al., 2008). These results are also consistent with previous spatial correlation fields between the precipitation reconstruction and global SSTs shown in Fig. 5, and the spectral analyses included in Fig. 6. This ENSO-precipitation teleconnection across the Western Americas have also been described between precipitation sensitive records from Central Chile (32–35° S) and Southwestern North America during the last 350 yr (Villalba et al., 2011). However, the relationships between ENSO and precipitation in both regions is similar (wet years during the ENSO events) and opposite to the documented relationships between ENSO and precipitation in the Altiplano.

The application of the regimen shift detection to the precipitation reconstruction shows the occurrence of six long-term periods with significant reduced precipitation: a 40-yr interval centered in 1400, almost the entire 16th century connected to a decade-long drought during the first half of the 17th century, the second half of the 18th century, and an unprecedented dry period in the last 20 yr of the reconstruction. Interestingly, the most extended and severe droughts during the 16th and the 18th centuries were concomitantly with a strong reduction in the variance of the reconstruction. In contrast, pluvial periods showed high levels of interannaul precipitation variability (Fig. 8). As droughts in the South American Altiplano are trigged by El Niño-like conditions (Garreaud et al., 2009), it is likely that extended dry periods occurred in conjunction with a reduction of interannual precipitation variability modulated by persistent El Niño like conditions. However, the relationships between relative-high variance and humid conditions break during the last 20 yr of the reconstruction, where interannual variability increased in a long-term interval with reduced precipitation.

5 Concluding remarks

In this study we present the first quasi-millennial, tree-ring based annual precipitation reconstruction (November–October) for the South American Altiplano. This high-resolution precipitation reconstruction covers the past 707 yr in a region devoid of such environmental proxy records. Our reconstruction extends the dendroclimatological studies to the tropical Andes and represents the closest tree-ring based reconstruction to the Equator in South America. Our study provides insight into the Altiplano climate through the identification of long-term wet or dry periods and the temporal evolution of extremes in annual precipitation during the past 7 centuries. In addition, interannual and decadal scales variations in precipitation and ENSO variability are identified, showing common cycles and periodicities between precipitation on the Altiplano and this hemispheric forcing. This reconstruction improves our knowledge on interannual, decadal and multicentury scale precipitation variability in the Altiplano and would serve as a resource for research on the past, present and future climate variability in South America.

Some of persistent drought/wet periods in the past 707 yr are highly consistent with evidence from the few available proxy records in the region. For example, the droughts during 14th century (Boucher et al., 2011) and second half of 20th century (Boucher et al., 2011; Jomelli et al., 2011), and the humid period during 17th century (Rabatel et al., 2006). However, other extreme precipitation anomalies, such as the drought of 1520–1597 or the long pluvial extreme from 1820 to 1880, have not been reported before. A high concentration of extreme dry events occurred during the last 70 yr with 4 of the driest years since 1300 AD. Moreover, the 3 most severe droughts from the past 707 yr occurred since the 1980s (1982, 1994 and 2006). The instrumental analysis of precipitation patterns in the Altiplano region can be addressed only for the last 50 yr, which preclude detecting any robust long-term trend in rainfalls (Vuille et al., 2003). Our 707-yr rainfall perspective allows setting the 20th century and the period of instrumental records in the long-term context. A persistent negative trend in the precipitation reconstruction since the early 20th century suggests that the 50-yr interval of instrumental records is concurrent with the last, long-term dry event in the Altiplano, and consequently is not representative of the precipitation regime in the region.

Results from Regional Climate Models (RCMs) indicate that increased greenhouse gas emissions will exacerbate dry conditions in the Altiplano to the end of the 21th century. Most RCMs predict an increase in the westerly flow over the Altiplano, which will induces a decrease in the transport of humid air masses from the east. RCMs estimate a precipitation reduction in the Altiplano from 10 to 30 % along the 21th century (Urrutia and Vuille, 2009; Minvielle and Garreaud, 2011). As ENSO variability is a key factor affecting precipitation patterns on the Altiplano, high-resolution precipitation reconstructions from the Central Andes can provide valuable information on the functioning of ENSO teleconnections with the Altiplano under different background global climate conditions. On the other hand, our reconstruction together with ENSO-sensitive records around the world will help to understand the spatial dynamics of ENSO teleconnections worldwide and consequently improve ENSO predictability.

Our reconstruction points out that century-scale dry periods are a recurrent feature in the Altiplano. The potential coupling of natural and anthropogenic-induced droughts in the near future would have a severe impact on present socio-economical activities in the region. In the western, drier sector of the Altiplano, water resources are under a severe growing pressure. Human and the fast expanding mining activities obtain water from the scarce streams originated in the Altiplano and from overexploited aquifers that depend on the groundwater recharge from the Central Andes (Messerli et al., 1997; Houston, 2002). Frequency and intensity of future dry and wet episodes need to be anticipated to properly establish the strategies for agriculture, industry and populations water demands. Water resource managers must anticipate these changes to adapt for future climate change, reduce vulnerability and provide water equitably to all users.

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Table 1. Precipitation stations from the Altiplano used to calibrate *Polylepis tarapacana* tree-rings.

Station, code	Lat S, long W	Elevation (m)	Country	Period	Mean mm*
Patacamaya, Pat	17°15′/67°57′	3789	Bolivia	1948–2003	390
Charaña, Cha	17°35′/69°26′	4059	Bolivia	1948–2004	263
Visviri, Vis	17°37′/69°28′	4080	Chile	1968–2007	293
Caquena, Caq	18°03′/69°12′	4400	Chile	1970–2007	411
Putre, Put	18°11′/69°33′	3545	Chile	1970–2007	191
Cotakotani, Cot	18°11′/69°13′	4550	Chile	1963–2007	448
Chucuyo, Chu	18°12′/69°17′	4400	Chile	1961–2006	345
Parinacota, Par	18°12′/69°16′	4420	Chile	1933–2007	324
Chungará, Chn	18°16′/69°06′	4600	Chile	1962–2008	374
Guallatiri, Gua	18°29′/69°09′	4240	Chile	1969–2007	270
Colchane, Cls	19°16′/68°38′	3700	Chile	1978–2007	138
Huaytini, Hua	19°33′/68°37′	3720	Chile	1982–2008	157
Salinas G.M., Sgm	19°38′/67°40′	3737	Bolivia	1948–2001	211
Coyacagua, Coy	20°03′/68°50′	3990	Chile	1961–2008	131
Uyuni, Uyu	20°28′/66°48′	3660	Bolivia	1975–2003	185
Colcha, Col	20°47′/67°47′	3700	Bolivia	1980–2000	207
S. Pablo Lípez, Spl	21°41′/66°37′	4165	Bolivia	1979–2003	289

*Mean annual (Nov–Oct) precipitation (mm) for the common period 1982–2000.

Table 2. Characteristics of *Polylepis tarapacana* tree-ring sites and the regional chronology from the Altiplano, Central Andes.

Site name, code	Lat S, long W	Elev (m a.s.l.)	Country	No. series	Period	*r* PC1*
Volcán Guallatiri, GUA	18°28′, 69°04′	4450	Chile	82	1377–2007	0.77
Salar de Surire, TER	18°56′, 69°00′	4517	Chile	11	1278–1901	0.77
Frente Sabaya, FSA	19°06′, 68°27′	4430	Bolivia	30	1352–2008	0.73
Queñiza, QUE	19°22′, 68°55′	4303	Chile	51	1444–2007	0.78
Volcán Caquella, CAQ	21°30′, 67°34′	4520	Bolivia	63	1226–2009	0.82
Soniqueira, SON	22°00′, 67°17′	4543	Bolivia	35	1431–2003	0.72
Volcán Uturuncu, UTU	22°32′, 66°35′	4457	Bolivia	81	1242–2006	0.84
REGIONAL Chronology statistics: MTR 0.47/MS 0.3/EPS 0.95/				353	1242–2009	0.98

* Loadings (correlation coefficients) of each chronology with the Principal Component 1 (explain 60 % of the variance) of the standard site chronologies for their well replicated common period 1668–1776.
All correlations are siginificant at $P < 0.001$ level. MS: Mean Sensitivity, MTR: Mean Tree-Ring Width (mm), EPS: Expressed Population Signal.

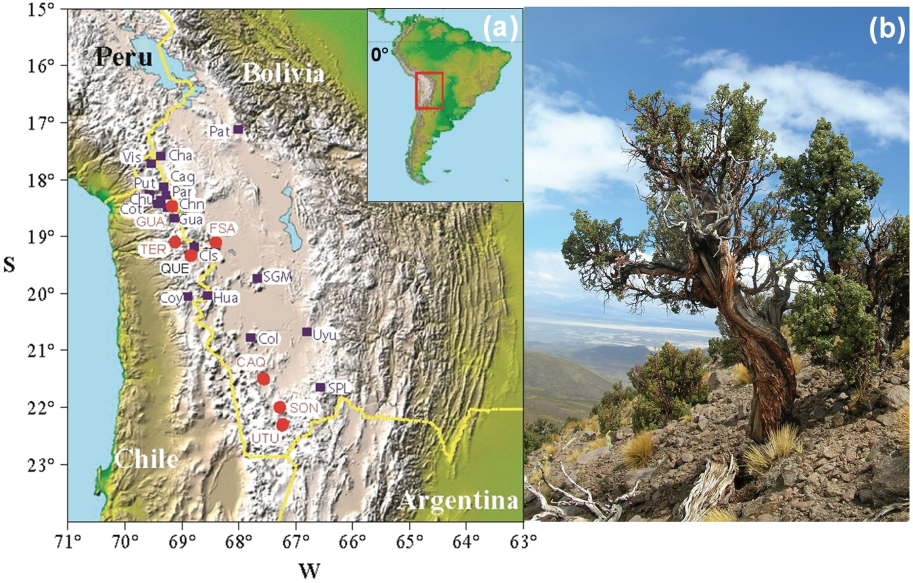

Fig. 1. Location of tree-ring sites (red dots) and precipitation stations (blue squares) in the Altiplano, Central Andes. See Tables 1 and 2 for codes identification **(a)**. A 500 yr old *Polylepis tarapacana* (*Queñoa*) individual growing on the slope of the Tata Sabaya Volcano in Bolivia at 4750 m a.s.l. In the background, the Coipasa salt-lake in the Bolivian-Chilean border **(b)**.

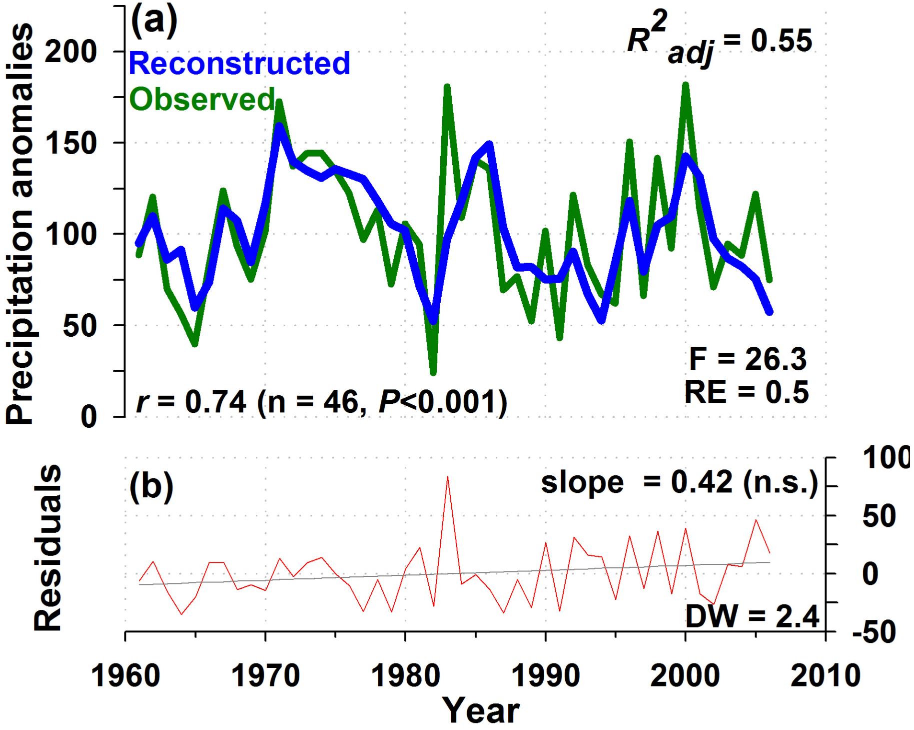

Fig. 2. Observed and tree-ring predicted annual precipitation (November–October) variations across the South American Altiplano (annual precipitation expressed as percentages (%) of the 1982–2000 instrumental precipitation mean). Calibration and verification statistics: explained variance (R^2_{adj}) over the calibration period, the Pearson correlation coefficient (r) between observed and reconstructed values, F-value of the regression, and the reduction of error (RE) **(a)**. Regression residuals (red line) with trend slope (black line). The Durbin-Watson (D-W) statistic and the slope value are indicated **(b)**.

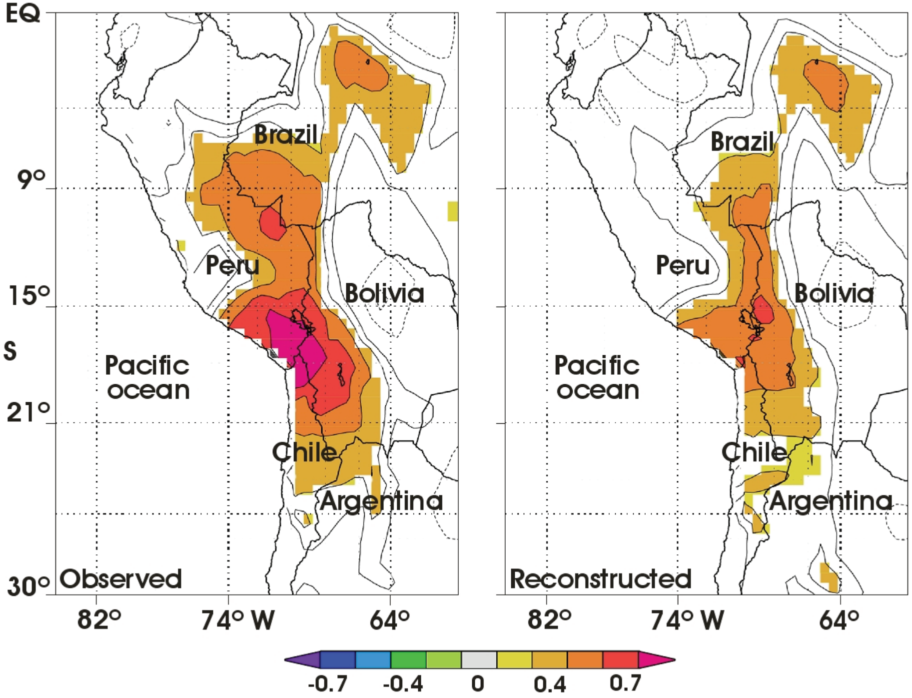

Fig. 3. Spatial correlation field between the CRU 3.1 0.5° × 0.5° gridded November–October precipitation and our regional instrumental precipitation series for the Central Andes (see Table 1) **(a)**, and the Altiplano reconstructed November–October precipitation **(b)** for the 1961–2006 period (only significant correlations are shown).

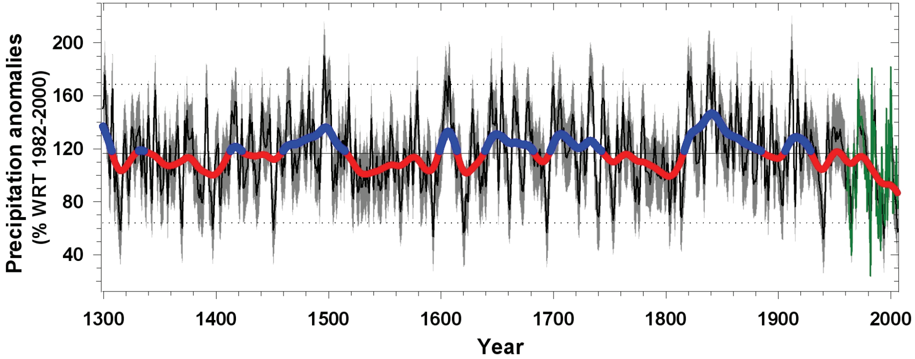

Fig. 4. The tree-ring reconstruction of annual (November–October) precipitation in the Altiplano region, Central Andes, for the period 1300–2006 (annual precipitation expressed as percentages (%) of the 1982–2000 instrumental precipitation mean). The shaded denotes the 1 ± Root Mean Squared Error bars and the green line represents the instrumental record. To emphasize the low-frequency variations a 35-yr smoothing cubic spline designed to reduce 50 % of the variance is shown in blue and red indicating wet and dry periods, respectively with respect to the 1300–2006 mean. The dotted horizontal lines indicate ±2 standard deviations.

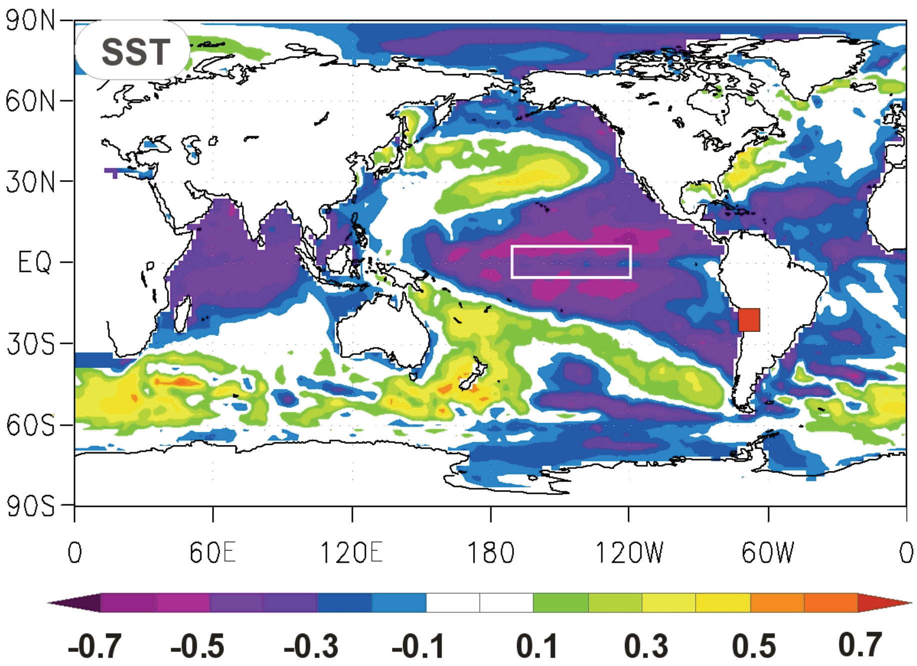

Fig. 5. Spatial correlation field between the annual (November–October) precipitation reconstruction and 2.5° × 2.5° gridded monthly averaged November–October Sea Surface Temperature (SST) for the interval 1948–2006 (NCEP–NCAR reanalysis). The white box indicates the Niño 3.4 region in the tropical Pacific. The reconstructed precipitation region is indicated by the red square.

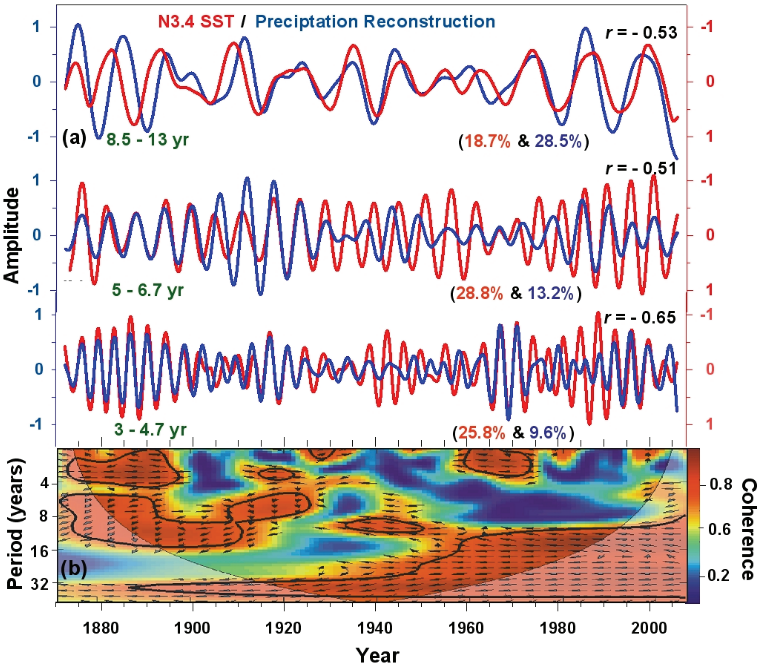

Fig. 6. Comparisons between the spectral properties of the Altiplano precipitation reconstruction and the N3.4 SST record during the common period. Waveforms extracted by Singular Spectrum Analysis (SSA). The frequencies for each SSA are indicated in years with green numbers, the correlation between the two series at the right corner, and the percentage of variance explained by each frequency indicated in parenthesis **(a)**. Wavelet coherency (WTC) and phase spectrum between the Altiplano precipitation reconstruction and the N3.4 SST. The vectors indicate the phase difference between the two records (arrows pointing right and left corresponds to in-phase and anti-phase relationships, respectively). Thick black contours encircled the periods where both series were significant related at a significance level (95 % c.l.). The cone of influence is shown at the bottom lighter shaded **(b)**.

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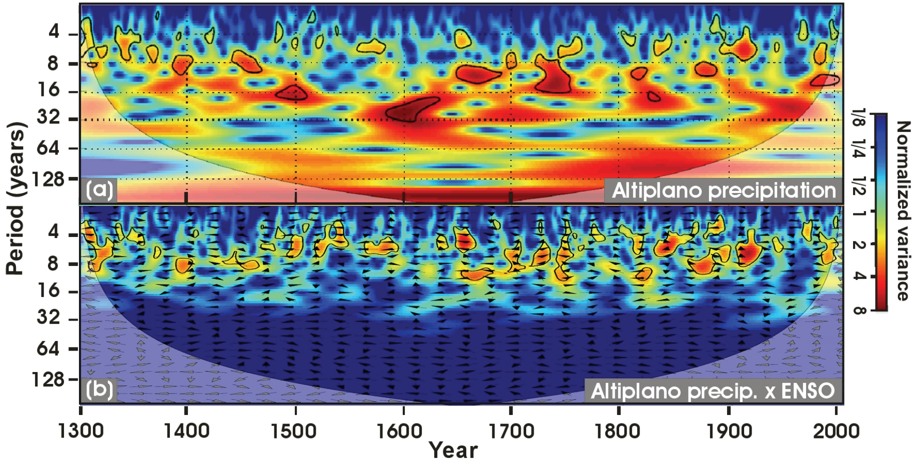

Fig. 7. The wavelet (WT) power spectrum (Morlet) of the annual (November–October) precipitation reconstruction in the Altiplano region **(a)**, and the cross-wavelet transform (XWT) between the precipitation reconstruction in the Altiplano and the first principal component of the North American Drought Atlas (NADA) as an ENSO proxy during the period 1300–2006 (Cook et al., 2004; Li et al., 2011) **(b)**. Thick black contours indicate the 95 % significance based on the red noise model, and the cone of influence is shown as lighter shade at the bottom of both figures. Vectors indicate the relative phase relationship between the Altiplano precipitation and NADA PC1. Horizontal arrows pointing right and left correspond to in-phase and anti-phase relationships between records, respectively.

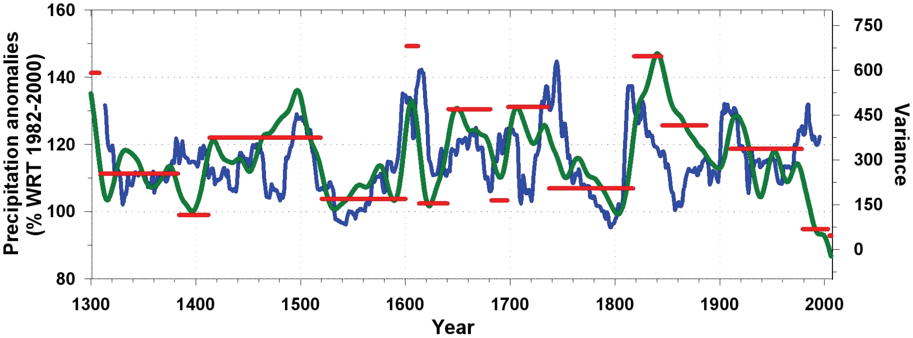

Fig. 8. Comparison between periods of reduced vs. abundant precipitation and interannual variability in the Altiplano precipitation reconstruction. Significant (95 % c.l.) regime shifts (red line) detected by the Rodionov (2004) method (window length = 25 yr), smooth spline (35 yr) of the precipitation reconstruction (green line) shown in Fig. 3, and changes in variance (× 10) calculated for 25-yr intervals plotted on the centroid + 1 for each interval (blue line).