Long-term glacier variations in the Central Andes of Argentina and Chile, inferred from historical records and tree-ring reconstructed precipitation

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A R T I C L E   I N F O
Article history:
Received 11 July 2007
Accepted 18 January 2008
Available online 7 September 2008

Keywords:
Austrocedrus chilensis
Tree-rings
Historical documents
Glacier retreat
Dendroclimatology

A B S T R A C T
Snow and ice in the Central Andes of Chile and Argentina (33–36 °S) are the major source of water for the highly populated regions near the cities of Santiago and Mendoza. However, our knowledge of the forces driving the general glacier retreat in the region is limited. In order to obtain a long-term perspective of glacier fluctuations and their relationships with climate in the Central Andes, historical glacier variations were documented and compared with a tree-ring precipitation reconstruction based upon Austrocedrus chilensis trees. A multi-proxy approach (historical documents, old aerial photographs and satellite imagery) was used to map the fluctuations of eight glaciers, including the Cipreses Glacier, which provides the oldest record of glacier variations in the region starting in AD 1842. All the studied glaciers exhibited a negative trend during the 20th century with mean frontal retreats between −50 and −9 m y−1, thinning rates between 0.76 and 0.56 m y−1 and a mean ice area reduction of 3% since 1955. More than 350 tree-ring cores were combined into three tree-ring chronologies, which strongly correlate with the instrumental precipitation in Santiago de Chile (33°26′ S; 70°41′ W, 520 m asl). Based on these records, a 712-year precipitation reconstruction was developed. This reconstruction is characterised by a centennial oscillation indicating marked dry conditions around the years 1440 and 1600 AD. Wet conditions were prevalent in the years 1500, 1650 and particularly around 1850 AD. Since this precipitation maximum, the reconstruction shows a clear, secular, decreasing trend. The reduction in precipitation indicated by this reconstruction for the last 150 yr, in combination with a significant warming recorded in Central Chile, are the main causes of the observed current glacier retreats.

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1. Introduction

Due to its transitional location between subtropical and temperate latitudes, the climate in Central Chile is characterised by a high interannual variability in precipitation (Aceituno et al., 1993; Montecinos and Aceituno, 2003; Quintana, 2004). At the end of 1960s and during the 1990s, this region experienced extreme drought events (Quintana, 2000), which seriously affected water consumption, irrigation, and generation of hydroelectric power (Norero and Bonilla, 1999). These natural restrictions in water supply occurred concurrently with higher human pressure on water resources, generating strong competition for water allocation as a result of the regional economic growth (Rosegrant et al., 2000). The drought events in this region are partially correlated to cold phases of ENSO, the so-called La Niña events (Masiokas et al., 2006), which are typically associated with negative glacier mass balances, as opposed to positive glacier mass balances during El Niño events (Escobar and Aceituno, 1998).

Water consumption in this region of Argentina and Chile is highly dependent on snow accumulation and glacial melt, particularly during summers preceded by episodic winter droughts when up to 67% of the water flow of Río Maipo (Peña and Nazara, 1987) and 70% of the discharge of Río San Juan (Leiva, 1999) has been generated by glacial meltwater. In this regard, and additionally due to recent glacier retreat and wasting ice, runoff from some glaciated Chilean basins has increased (Carrasco et al., 2005). In the long-term, runoff is expected to drop significantly during the dry seasons due to the disappearance of glaciers, in similar fashion to what is currently occurring in the Tropical Andes (Juen et al., 2007).

In spite of the general trend of glacial retreat on both sides of the Andes during the 20th century (Llorens and Leiva, 1995; Leiva, 1999; Rivera et al., 2002), some glaciers experienced sudden advances. This occurred both in Chile during the 1950s (Lliboutry, 1956) and in Argentina during the 1930s, and 1980s, when Glaciar Grande del Nevado del Plomo advanced rapidly, damming a valley and generating two floods (Bruce et al., 1987). Similarly, Glaciar de la Laguna in Río Atuel advanced 1400 m between 1970 and 1982 (Cobos and Boninsegna, 1983). Some of these advances have been related to
surge events (Leiva, 1999), which are not necessarily responses to climatic changes. In this situation they may be generated by changes in the normal subglacial drainage system, particularly when temperate glaciers are involved, such as in Central Chile and Argentina (Paterson, 1994; Harrison and Post, 2003).

Recent glacial responses in this part of the Andes have previously been related (Rivera et al., 2002) to both atmospheric warming (Rosenblüth et al., 1997) and decreases in precipitation (Quintana, 2004). As yet, little research has been done regarding Holocene glacial responses (e.g., Espizúa, 1993). These long-term variations are useful for understanding how water resources may change in the future as a consequence of predicted temperature and precipitation changes (e.g., IPCC, 2001).

In order to obtain a better insight into the long-term climate variability of this part of the Andes, two independent proxies were used: historical documents on glacier variations and precipitation variability reconstructed by moisture-sensitive tree-ring chronologies. These two proxies allow extension of analysis into the past, for periods where instrumental records are limited or non existent (Bradley et al., 2004).

For this purpose, eight glaciers located in the Cachapoal and Tinguiririca basins in Chile, and the Atuel basin in Argentina were selected (Table 1, Figs. 1 and 2). Most of these glaciers have previously been analysed (e.g., Caviedes, 1979; Cobos and Boninsegna, 1983; Rivera et al., 2006), allowing the reconstruction of long-term variations of frontal glacier tongues. One of the glaciers with the longest historical record of frontal variations is Glaciar Cipreses (34°33′ S; 70°22′ W), whose frontal position was first noted by Domeyko in 1842 (Domeyko, 1978). More recent aerial photographs and satellite images (Table 2) have enabled updating of glacier positions to the year 2007.

Glaciar Cipreses was also chosen for more detailed analysis, due to disjunctive stands of Austrocedrus chilensis at lower altitudes in its valley (Ciprés de la Cordillera, Fig. 2). This species has previously been used to provide a proxy record for precipitation in central Chile (Boninsegna, 1988, 1992; Luckman and Villalba, 2001). A robust moisture-sensitive tree-ring network was developed to analyse the long-term precipitation variability during the last seven centuries. This annually-resolved proxy allowed the determination of the leading modes of precipitation variability at different time scales.

Table 1
Geographical description of studied glaciers.

<table>
<thead>
<tr>
<th>River basin</th>
<th>Chile</th>
<th>Argentina</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacier name</td>
<td>Cipreses</td>
<td>Cortaderal</td>
</tr>
<tr>
<td>Latitude S</td>
<td>34°33′21″</td>
<td>34°38′32″</td>
</tr>
<tr>
<td>Longitude W</td>
<td>70°22′10″</td>
<td>70°15′35″</td>
</tr>
<tr>
<td>Length (km)</td>
<td>10.6</td>
<td>9</td>
</tr>
<tr>
<td>Total area (km²)*</td>
<td>35.7</td>
<td>15</td>
</tr>
<tr>
<td>General aspect</td>
<td>W</td>
<td>SE</td>
</tr>
<tr>
<td>Min. elevation (m)**</td>
<td>2616</td>
<td>2738</td>
</tr>
<tr>
<td>Max. elevation (m)**</td>
<td>4365</td>
<td>4845</td>
</tr>
</tbody>
</table>

*Area and length based upon a Landsat ETM+ acquired on January 20, 2000.
**Altitude extracted from SRTM data.
(e.g., inter-annual, decadal/inter-decadal and secular oscillation modes) and enabled inferences on the effect of possible climatic forcing on glacial behaviour (e.g., ENSO).

2. Data and methods

2.1. Glacier variations

Variations in glacial fronts and area were obtained through comparison of geo-located and ortho-rectified satellite images, acquired from 1976 to present (Table 2), where glacier boundaries were digitised on screen based upon colour composite images produced from Landsat MultiSpectral Scanner MSS, Thematic Mapper TM and Enhanced Thematic Mapper Plus ETM+ bands with histogram-equalised stretches. The horizontal errors of these satellite images, after geo-location and ortho-rectification, are within one pixel size (Table 2).

Several aerial photographs (Table 2) were manually analysed with mirror stereoscopes to map the approximate position of the glaciers. The resulting interpretation was then transferred to a 1:50,000 scale regular cartography of the area using a Zoom Transfer Scope analogue instrument, yielding horizontal errors of a couple of tens of metres. The final interpretations were then digitised and incorporated into a Geographical Information System (GIS), where all available data sets were compared, taking into account any temporal changes in glacial characteristics.

The positioning of glacial fronts from historical accounts (prior to 1955) is typically less accurate than using remotely sensed imagery, however, the reports available in the literature provide precious information from a time where very little is known about glacier behaviour. Thus, all historical records were carefully interpreted and compared between available sources, in order to estimate the position of the glacial fronts at specific times. Some of these reports provided a photograph or a sketch map of the glacier; others describe retreats or changes with respect to previous positions. In some accounts the glaciers were only described in terms of minimum altitude or relative position in comparison to a geographically prominent feature (e.g., river confluence), making interpretation more difficult. The positions of the glacier for these historical accounts are considered a first order approximation, within several tens or even a couple of hundred of meters.

The Equilibrium Line Altitudes (ELAs) of the glaciers were determined by the analyses of Landsat ETM+ images acquired in the summer. At the end of the dry season in the Central Andes of Argentina and Chile, the transient snow line is coincident with the ELA due to the temperate condition of the glaciers (Rivera et al., 2000).

2.2. Ice elevation changes

A Digital Elevation Model (DEM) was generated by interpolation of photogrammetrically-derived contour lines (published by Instituto Geográfico Militar (IGM)) of Chile, based upon 1955 Hykon aerial photographs (Table 2). No error estimations for the regular cartography of the IGM are available, however, the regular cartography at 1:50,000 scale is considered to be of class 1 accuracy in terms of the American Society for Photogrammetry and Remote Sensing classification (Falkner, 1995). This cartography is therefore, most precise, with vertical inaccuracies related to the contour interval (25 m) possessing a vertical error of 17 m. After applying several interpolation methods, a jack-knifing procedure (Lythe et al., 2001) allowed determination of the DEM of best fit, yielding 9 m of vertical random error when an Inverse Distance Weighted (IDW) method was applied. Therefore, the total RMS error of the DEM was 19 m. The coverage of this DEM was not successful in some sectors where the aerial photographs used by the IGM in the restitution process exhibited stereo matching problems due to snow covered surfaces, steep flanks or shadows slopes.

Shuttle Radar Topography Mission (SRTM) data obtained from JPL/NASA in 2000 were also collected covering the majority of the study area, with the exception of small sections that lacked topographic information due to their rough topography and steep slopes. At the margins of these non-data areas some spurious pixels were identified.
with altitude values more than 50 m lower than the surrounding pixels. These incorrect elevations are a consequence of the conversion applied by JPL/NASA when reducing the original 30 m pixel size to the 90 m resolution available in the public domain. These spurious pixels were manually detected and eliminated, following a similar procedure to Rignot et al. (2003).

In order to estimate SRTM vertical errors, rock areas were compared between both DEMs yielding a small bias (1.8 m) and a vertical random error of 20.9 m (one standard deviation). This vertical random error is composed of the inaccuracies of the IGM DEM (19 m) and the error of SRTM (9 m), including a possible bias between SRTM and altitude (Berthier et al., 2006). These errors are similar to the

Fig. 3. The upper left panel shows the location of the main glacier areas (year 2000) and the Austrocedrus chilensis tree-ring chronology sites (circles). See Table 3 for details of each site. Individual panels show the fluctuations in frontal positions of the main glaciers in the study area. Arrows indicate the main ice flow directions.
results obtained in Patagonia by Rignot et al. (2003), after comparing SRTM and DEMs derived from the IGM regular cartography.

2.3. Dendrochronological records

Austrocedrus chilensis tree-rings are well defined (Roig, 1992) and highly sensitive to moisture variability. During the last decades, this conifer has been extensively used as a precipitation proxy in the south-central Andes of Argentina and Chile (Boninsegna, 1992; Villalba et al., 1998; Le Quesne et al., 2006).

The trees sampled for this study were found near the northern limit of the natural distribution area of the species (Fig. 1), at 32°39′ S in El Asiento (some 80 km north of Santiago) and at 34°40′ Si n Rí o Cipreses Valley (100 km south of Santiago). The trees grow in very steep rocky environments between 1600 and 2200 m altitude, where they are subject to moisture stress during the dry summer season (Fig. 2). The arid and relatively cold habitat favours good preservation of subfossil wood, which was used to extend the chronologies back in time. Some of the A. chilensis sites that yielded chronologies in early collections (LaMarche et al., 1979) were revisited and updated with new collections. Other sites were added to develop a chronology network on Río Cipreses Valley (Figs. 1 and 3), with some specimens older than nine hundred years (Le Quesne et al., 2006). The network used for this study includes: El Asiento (ASI) chronology from the northern limit of the species (LaMarche et al., 1979); a composite chronology (ELU, Table 3) of three sub-sampled sites (East Urriola-URE, West Urriola-URO and El Baule-ELB; Fig. 3) located in the lower portion of the Cipreses valley; and a chronology in Agua de la Muerte (AMU) from the upper portion of the valley, only 5 km from the present frontal terminus of Glaciar Cipreses (Table 3; Figs. 1 and 3).

Core samples were prepared following standard dendrochronological techniques (Schweingruber et al., 1990; Stokes and Smiley, 1996). Ring widths were measured using an incremental measuring machine at 0.01 mm precision level. The ring width series were visually cross-dated and then, the accuracy of the dating process tested using the COFECHA programme (Holmes, 1983).

The cross-dated samples were standardised using the regional curve method (Briffa et al., 1992) in order to remove the biological growth trend, while retaining the low frequencies related to climate. The standard chronologies were generated using ARSTAN program (Cook and Holmes, 1984).

2.4. Precipitation reconstruction

Santiago de Chile has a continuous record of monthly mean precipitation beginning in 1866. The station is located in Quinta Normal (33°26′ S; 70°41′ W, 520 m asl) where it is controlled by the Dirección Meteorológica de Chile (DMC), and provides the longest and most reliable record in the area.

The growth of A. chilensis shows significant positive correlation with the monthly mean Santiago de Chile precipitation from April to November (Boninsegna, 1988; Le Quesne et al., 2006). Although the May–November period accounts for ca. 80% of the total annual
precipitation, the total annual precipitation was used, in order to better capture the inter-annual precipitation variability. As the relationship between tree-ring indices and annual precipitation is not linear; a lineal model was fitted to the instrumental precipitation through natural logarithmic transformation.

A tree-ring based multivariate model was developed to reconstruct Santiago precipitation between 1866 and 1999, using the three standard tree-ring chronologies as predictors (Table 3). In the multi-regression model the three chronologies were incorporated at year $t$ and lagged at year $t-1$ and $t+1$, because the radial growth of trees is influenced by climatic conditions during the current and in the following year through complex physiological processes affecting buds, leaf growth, and reserve accumulation. Due to this relationship, the ring width at the year $t-1$ is statistically related to the ring width of the year $t$ and is included in the model to account for autocorrelation (Fritts, 1976). Therefore, the chronologies in year $t-1$, $t$ and $t+1$ were used as candidate predictors of year $t$ precipitation (e.g., D’Arrigo and Jacoby, 1991; Villalba et al., 1997).

In order to avoid co-linearity, principal components were extracted from the nine candidate tree-ring predictor variables and the factor scores were used to replace the original variables of the time series (Villalba et al., 1998; Bradley, 1999). To select the best subset regression models, the maximum adjusted $r^2$ criterion (Adj $R^2$) was used (Draper and Smith, 1981). The reconstruction models were tested by dividing instrumental precipitation (1866–1999) into early and late sub-periods to perform the calibration-validation exercises (e.g., Fritts, 1976; Cook and Kairiukstis, 1990).

2.5. Spectral methods

Two spectral methods were used to detect non-stationary oscillatory modes of variability in the 712-yr series of reconstructed precipitation. Firstly, the distribution of the variance and the signals contained in the reconstruction were examined using Continuous Wavelet Transform (CWT) analysis (Torrence and Compo, 1998). CWT analysis allows the detection of periodic significant signals (waveforms) by expanding one-dimensional time series into two-dimensional time-frequency space (Grinsted et al., 2004). This method allowed the simultaneous representation of the dominant modes of variability (periodic signals) and the variation of those modes in time (Torrence and Compo, 1998). The CWT analysis was conducted using the Matlab routine WTC-R15 provided by Grinsted et al. (2004). Secondly, the periodic behaviour of the reconstructed time series was evaluated by extracting the dominant (i.e., eigenvalue $>1$) oscillatory components (wave forms) using Singular Spectral Analysis (SSA; Vautard and Ghil, 1989). The dominant wave forms of the reconstruction were combined into secular, decadal/inter-decadal, and inter-annual timescale periods, in order to facilitate the comparison with glacier variations. The Singular Spectral Analysis was performed by using the SSA programme (Boninsegna and Holmes, unpublished operating manual [Laboratory of Tree-Ring Research, University of Arizona]).

3. Results

3.1. Glacial variations

Glaciar Cipreses is one of the few glaciers in the study area with a century-long historical record. This glacier has been frequently visited by naturalists since 1831 when the first record was obtained by Claudio Gay who documented moraines and described erratic blocks at the front of the glacier. Since then, several accounts (summarized in Fig. 4 and Table 4) were obtained from Pissis (1860), Espejo (1897), Barros Arana (1911), Risopatrón (1924) and Domeyko (1978). The only interruption to the retreat trend of Glaciar Cipreses during the 19th century took place in 1858, when a re-advance of the glacier was reported by Pissis (1875) and also documented by Röthlisberger (1986).

Concurrently, an advance was recorded for Glaciar Los Piquenes (34°22'S 70°04'W), located 30 km to the northeast of Glaciar Cipreses (Plagemann, 1887). In 1882, Glaciar Cipreses was again under a severe retreat trend, as illustrated by a large mass of ice located 1.5 km downstream from the glacier tongue (Güssfeldt, 1884). Unfortunately, no historical records were collected between 1882 and 1955, with the exception of the visit of Nogues (1894). Since 1955, the glacier retreat has been documented thanks to several aerial photographs and satellite images (Table 3).
Other glaciers in the study area have followed a similar trend, with generalized retreats since the beginning of the 20th century and in some cases since the first remotely sensed imagery were collected in the 1950s (Table 4, Figs. 3 and 5). In spite of the overall receding trend, several glaciers experienced advances during the 1970s and 1980s, for example, Glaciar de la Laguna, which advanced 1400 m between 1970 and 1982 (Figs. 3 and 5) in what is considered a surging event (Cobos and Boninsegna, 1983). Since then however, all glaciers have been retreating, with significant area losses up to the year 2007 (Table 5).

After comparing DEMs from 1955 and 2000 for the main glaciers on the Chilean side of the Andes, significant changes were documented in Glaciar Cipreses, Palomo and Universidad, which have experienced ice elevation changes of $-0.76$, $-0.6$ and $-0.56 \pm 0.47$ m y$^{-1}$ (SRTM-IGM DEM) respectively.

### 3.2. Tree-ring precipitation reconstruction

For this study, 367 tree-ring samples from dead and living trees of long-lived coniferous $A. \text{chilensis}$ were combined to develop a robust network of moisture-sensitive tree-ring data (Table 3). The network comprises tree-ring series ranging from 1200 to 150 yr, with a mean segment-length of ca. 335 yr. The chronologies ASI, ELU and AMU, include 42, 51 and 58% of the samples >300 yr, respectively. This feature, along with the appropriate sample replication and the conservative standardization method (RCS) used to maximize the climatic signal, assumes that the dendrochronological record captures an appreciable amount of both low and high frequencies in precipitation variability.

The results of the calibration–verification experiments using PC factor scores of tree-ring chronologies as predictors of annual precipitation are summarized in Table 6. The number of orthogonal tree-ring variables entered into the regression models was 5. Cross-validation tests were performed from the 1866 to 1999 interval (full period). The observed precipitation time series was split into two sub-periods, using the earlier part of the record (1866–1933) for calibration and the later part (1934–1999) as independent verification of tree-rings estimates of precipitation. Subsequently, the same calibration–verification procedure was applied with reversed sub-periods. Pearson coefficient results (0.71, 0.76) are indicators of the reconstruction validity. The RE statistic ($>0.5$) is a measure of shared variance between the actual and modelled series, and any positive value denotes that the regression model provides valuable palaeoclimatic information (for further details in calibration and verification see Cook and Kairiukstis, 1990).

The regression model for annual precipitation accounts for 52% of the total variance over the full calibration interval (134 yr). Although the relationship for the entire calibration interval is not as good as for the earliest sub-period, the full record was used in order to enhance the ability of the model to capture the long-term precipitation variability (Table 6). The observed precipitation with the tree-ring reconstructed precipitation is shown in Fig. 6. As previously noted by Boninsegna (1988), $A. \text{chilensis}$ tree-rings are better at estimating extreme dry- than extreme wet-years such as 1877, 1880, 1888, 1899–

![Fig. 5.](image-url)
1900, 1914, 1926, 1941, 1982, 1987 and 1997, the ten wettest years in the instrumental precipitation series of Santiago since 1866 (Fig. 6). Tree-rings typically underestimate wet extremes, because soil properties and tree physiology limit the response of tree growth to abundant precipitation (Meko et al., 1995; Villalba et al., 1998).

Using the model derived in the calibration, the precipitation was reconstructed back to AD 1280. The entire reconstruction is shown in Fig. 7a (top panel), along with a 25-yr smoothing spline (Cook and Peters, 1981), to highlight long periods of above and below-mean precipitation.

The second panel of Fig. 7 shows the wavelet power spectrum (CWT) of the precipitation, displayed as a function of the period and time. At the beginning of the end of the time series, the edge effect (defined by the cone of influence) becomes relevant, and consequently the results should be interpreted cautiously (Torrence and Compo, 1998).

In general, the results of the long-term reconstructed precipitation variability show that the amplitudes and relative power of the detected signals changes substantially over time (Fig. 7b). The wavelet analysis shows a significant contribution of the variance in the interannual and decadal/inter-decadal (<23 yr periods) frequency band. This is shown by the oscillatory modes extracted with SSA and grouped in 2.1–3.4 (Fig. 7e) and 13–23 yr (Fig. 7d), which explain 37% of the total variance. At lower frequencies a secular trend (Fig. 7b) delineated at the bottom of the Cone of influence (>100 yr) contribute with ca. 11%. The long-term multidecadal waves around 35 and 85–93 yr are not detected by SSA, because they account for only 0.5% and 3% of the total variance.

The secular trend (208-yr waveform) extracted by SSA (Fig. 7c), explains 11.6% of the variance and captures the long-lasting dry and wet periods over the last seven centuries. Note the decline since the late 19th century (ca. 1890) to the present, and the prolonged dry spells during early 18th (1700–1735), late 16th and early 17th (1570–1635), and the majority of the 14th to the first half of 15th (1330–1450) centuries. Wet conditions occur with maxima centred on the years AD 1500, 1650 and 1850.

Decadal/inter-decadal modes (13.0–23.0 yr; Fig. 7d), explain 18.5% of the variance, and are characterised by several multiyear periods of reduced amplitudes over the 700-yr long reconstruction. To mention the most noticeable episodes: early to mid 20th (1915–1970), mid 17th to mid 18th (1660–1740), 16th and first half of 17th (1500–1570 and 1595–1640) and mid 14th to mid 15th (1350–1440) centuries. Enhanced decadal variability took place during the following periods: AD1890–1910, 1740–1770, 1640–1660, 1570–1595 and 1445–1500.

The inter-annual oscillatory mode (2.1–3.4 yr; Fig. 7e), which explains 14.8% of the variance, is characterised by several periods of both reduced and increased variability. Within the episodes of lesser variability, the ca. 150-yr long period (1700–1850) is interrupted by a peak of variance that occurs in the late 19th century. Other periods of increased and decreased inter-annual variability include the late 16th to early 17th centuries (1590–1605) and the majority of the 15th century (1410–1490), respectively.

The comparison of inter-annual and decadal modes (Fig. 7d–e) with the El Niño 3.4 SST (Hurrell et al., 2008) and PDO index (not shown here), reveals a strong relation with tropical Pacific conditions contrasting with a weak relationship with the PDO (Mantua et al., 1997). The N3.4 inter-annual signal (2.0–3.6 yr) extracted by SSA correlates at r = 0.36 (for the common period 1871–1993), whilst the PDO index reaches only an r = 0.22 (for the common period 1900–1993). Interestingly, the decadal signal (13.0–22.0 yr) extracted from N3.4 series correlates with r = 0.44 for the common period 1871–1993, and with r = 0.69 from 1886 to 1993 (Fig. 7d).

4. Discussion

Several sources have documented the widespread retreat of glaciers in the Central Andes during the 19th and 20th century (Cobos and Boninsegna, 1833; Rivera et al., 2000; Luckman and Villalba, 2001). The historical record from Glaciar Cipreses, shows the same trend observed in both slopes of the Andes during the 19th century (Philippi 1867; Güßfeldt 1883; Hauthal 1895, Martin 1909). In spite of the longer record, no information is available for Glaciar Cipreses between the end of the 19th century to 1955 (Table 4; Fig. 4) thus exhibiting a hiatus in the historical record. The reconstructed period of glacier fluctuations is particularly important as it coincides with decades of relatively high (1880 to 1910) and low precipitation (1950 to 1970; Figs. 5 and 6), when several glaciers in the Mendoza (Luckman and Villalba, 2001) and Chilean (Lliboutry, 1956) Andes experienced advances and retreats, respectively.

The tree-ring chronologies obtained here allowed reconstruction of 700 yr of precipitation variability in Santiago of Chile. This is the longest and most reliable record ever obtained for Santiago, where the interannual and decadal/inter-decadal variability explains up to 37% of the variance. The centennial trend contributes to 11% of the variance. In the
700-yr series, a non-stationary mode of variability was found, with periods of prevalent decadal oscillations (e.g., 1450 to 1500 AD). Conversely in other periods the inter-annual variability dominates (e.g., 1670–1700 AD). At the beginning and end of the time series (i.e., before 1430 AD and after 1900 AD), the centennial wave is indicating noticeable drought conditions. In between, three peaks of wetter conditions are clearly visible around years 1500, 1650 and 1850 AD. Precipitation is reconstructed to have decreased from this last maximum in precipitation to the present. This trend is coincident with the instrumental (Quintana, 2004) and previously reconstructed precipitation (Boninsegna, 1988) records, with increasing risk of drought conditions in Central Chile (Le Quesne et al., 2006), upward migration of the snowline (Carrasco et al., 2005), and generalised glacier retreats (Rivera et al., 2002). Therefore, the Austrocedrus tree-ring chronologies utilised here must be considered a valid proxy for precipitation reconstructions, confirming previous studies by Boninsegna (1988) and Luckman and Villalba (2001).

Unfortunately, temperature response of these trees is relatively weak and confined to the spring season (Le Quesne et al., 2000). Therefore, only instrumental temperature records have been considered for the analysis. For this purpose, we used the annual mean temperature data (°C 1873–2003) for Central Chile (30–35°S, 70–
75°W) acquired from land and marine temperature grid cells (Jones and Moberg, 2003) (Fig. 5b). This series shows a long-term increase in temperature, with two clear jumps in 1947 and the 1970s, the latter coincident with the well-known PDO shifts (Giese et al., 2002). The first part of the series (before the 1940s) is characterised by a semi permanent cold anomaly followed by warmer conditions during the 1950s. During the period of frequent La Niña years in the 1960s and 1970s, temperature decreased, reaching a minimum just before 1976. After the PDO shift in 1976, temperature rapidly increased, and has remained relatively stable in the last quarter of 20th century.

The Central Chilean and Argentinean glaciers are part of the complex, mountain systems, directly affected by changes in temperature and precipitation, and subsequently, snow variability. The response of the glaciers and trees to these climatic variables is not immediate, with lag times varying from years to several decades in the case of glaciers (Paterson, 1994), and from seasons to years in the case of Austrocedrus (Fritts, 1976). In both cases, years with above (below) mean precipitation result in wider (thinner) Austrtocedrus ring widths and positive (negative) glacier mass balances. Additionally, lower (higher) than normal spring temperature values cause wider (thinner) tree-ring widths and positive (negative) glacial mass balances. Although modern glaciers and trees from this region exhibit similar responses to variations in precipitation and temperature during historical times, the interactions between the climate, trees and glaciers in central Chile and Argentina are not always straightforward.

Little is known about glacial behaviour during pre-historical times; with the exception of some moraine dating conducted in the region (Röthlisberger, 1986; Villalba, 1994; Espizúa, 2003). Thus, we speculate that glaciers were following climatic trends as illustrated in our reconstruction. Based on above-mean precipitation, ca. 1450, 1650 and 1850 AD, glaciers were most likely advancing during these intervals. However, the only glaciological data we have to support this suggestion is the reported advance of Glaciar Cipreses in the mid 1850s, presumably in response to the extended last wet period in the reconstruction.

The high variance of the reconstructed precipitations (ca. 18%), as explained by the ENSO related frequencies (2.1–3.4 yr), is a result of the documented sensitivity of central Chile climate to the equatorial Pacific conditions (Escobar and Aceituno, 1998; Montecinos and Aceituno, 2003; Garreaud et al., 2009–this issue). However, the results obtained here suggest that this climatic forcing can also modulate the precipitation variability at lower frequencies (13.0–23 yr; i.e., decadal/inter-decadal scale). This means that decadal/inter-decadal warm (cold) sea surface conditions are associated with warm (cold) and wet (dry) climate conditions in the study area, similar to the inter-annual phenomenon El Niño and La Niña (see also Barichivich et al., 2009–this issue).

5. Conclusions

All glaciers have a net retreat between 1955 and 2007, with mean frontal rates of −22 m yr^{−1} and −15 m yr^{−1} in the Chilean and Argentinean sides of the Andes, respectively. If we consider the historical data obtained for Glaciar Cipreses between 1842 and 1955, the frontal retreat of ca. 4.5 km resulted in a total area loss of approximately 3.5 km². Ice elevation changes measured in several glaciers also confirm the shrinking trend within the ablation areas, with thinning rates between −0.76 and −0.56 ± 0.47 m yr^{−1} between 1955 and 2000. On average, the glaciers of the study area in year 2000 were approximately 3% smaller than in 1955. This surface loss occurred in both sides of the Andes and was observed mainly at the ablation areas.

All these changes in glacier area are predominantly driven by the combination of the atmospheric warming, as illustrated by the increase in altitude of the 0 °C degrees isotherm, and the secular reduction in precipitation, namely during the last 150 yr as inferred from the reconstructed series. Atmospheric temperatures in Central Chile and Argentina have experienced two distinctive positive shifts since the beginning of instrumental records in mid 19th century (ca. 1940s and 1970s), the latter in agreement with the Pacific Decadal Oscillation shift widely described in the climatological literature.

The glacial retreat has been interrupted by two minor periods of advance (mid 1850s and mid 1980s), however, the lack of glaciological data from the beginning of the 20th century is most likely obscuring a third advancing cycle, which has been registered in glaciers located further north (Lliboutry, 1956).

From the wet period observed in the last two decades of the 19th century, the reconstructed precipitation of Santiago shows a negative trend during most of the 20th century (Aceituno et al., 1993; Luckman and Villalba, 2001; Quintana, 2004). Both, the instrumental record and the reconstructed precipitation indicate that the 19th century was wetter than the 20th. This is coincident with Central Chile rainfall reconstructions derived from different methods (Le Quesne et al., 2006).

The trees used in this study are located at relative long distances from the glacier. Due to the extreme scarcity of organic matter in the upper mountain regions in Central Chile, it is not possible to use 14C to date the moraines, as is common in many sites in Patagonia (Villalba et al., 1990). In order to overcome this problem, it is recommended that future studies of glacier variations in Central Chile utilize a multi-proxy approach, including several proxies such as historical documents, lake sediments, and lichenometry.

Acknowledgements

This research was funded by a CONICYT-Chile project FONDECYT 3020011: The Inter American Institute for Global Change Research grant CRN03 (Americas TreeLine Environment) and CECS. CLQ thanks the Millennium Nucleus P01-051-F, Universidad Austral de Chile, for a postdoctoral fellowship. We are grateful to D. Shea from the NCAR Climate and Global Dynamics Division for the N3.4 SST record compiled from Hurrell et al. (2008), CONAF (in particular Julio Cesar Vergara and Joaquín Bruna) provided valuable field-work support. María Eugenia Solari collaborated with the searching of historical documents. CECS is funded by the Base Financing Program of CONICYT-Chile and the Millennium Science Initiative. Comments from Mathias Vuille, Neil Glasser and Ricardo Villalba are highly appreciated. We thank Louise Newman, PAGES Office, for improving the English style.

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