Snowpack variations since AD 1150 in the Andes of Chile and Argentina (30°–37°S) inferred from rainfall, tree-ring and documentary records

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[1] The Andean snowpack is the main source of freshwater and arguably the single most important natural resource for the populated, semi-arid regions of central Chile and central-western Argentina. However, apart from recent analyses of instrumental snowpack data, very little is known about the long term variability of this key natural resource. Here we present two complementary, annually-resolved reconstructions of winter snow accumulation in the southern Andes between 30°–37°S. The reconstructions cover the past 850 years and were developed using simple regression models based on snowpack proxies with different inherent limitations. Rainfall data from central Chile (very strongly correlated with snow accumulation values in the adjacent mountains) were used to extend a regional 1951–2010 snowpack record back to AD 1866. Subsequently, snow accumulation variations since AD 1150 were inferred from precipitation-sensitive tree-ring width series. The reconstructed snowpack values were validated with independent historical and instrumental information. An innovative time series analysis approach allowed the identification of the onset, duration and statistical significance of the main intra- to multi-decadal patterns in the reconstructions and indicates that variations observed in the last 60 years are not particularly anomalous when assessed in a multi-century context. In addition to providing new information on past variations for a highly relevant hydroclimatic variable in the southern Andes, the snowpack reconstructions can also be used to improve the understanding and modeling of related, larger-scale atmospheric features such as ENSO and the PDO.

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1. Introduction

[2] In many respects the history and socio-economic development of the semi-arid Argentinean and Chilean regions flanking the Andes between ca. 30°–37°S have always been linked to the snowpack accumulated each winter in the cordillera. The seasonal melting of this snow is the main control of the river runoff to these regions [*Masiokas et al.*, 2006, 2010] (Figure 1) and provides most of the water needed for human consumption, agriculture, industries, hydroelectric generation and aquifer recharge. Since at least colonial times, snow in the intermontane valleys has also affected transportation and played a central role in the commercial and social relationships between the cities and communities on both sides of the mountains. Even today, heavy snowfalls delay the movement of goods and people on the main road across the Andes that connects the Pacific and the Atlantic sectors of southern South America.

[3] Interestingly however, very few analyses of instrumental snowpack variations are available for this portion of the Andes. Although snow water equivalent data have been diligently collected for over five decades at several high elevation sites, these data have primarily been used locally for predicting spring and summer snowmelt runoff. This short-term forecasting focus has limited our knowledge of the main spatio-temporal patterns of mountain snowpack until recently [*Masiokas et al.*, 2006, 2010]. These studies

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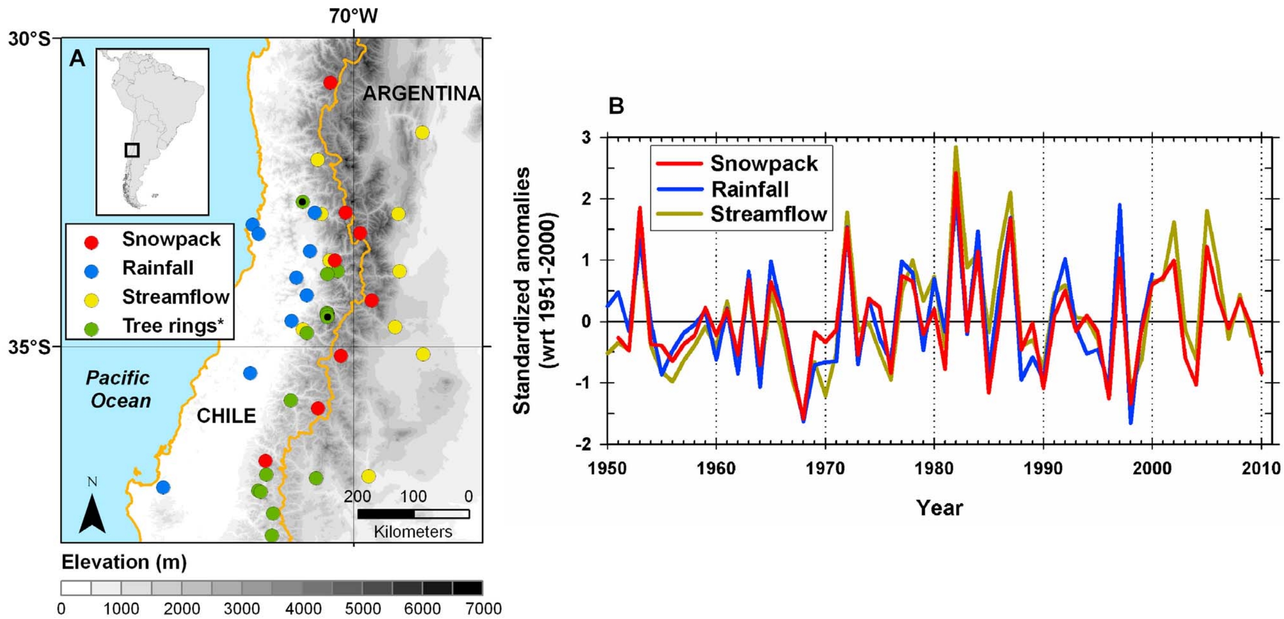

Figure 1. (a) Location of hydro-meteorological stations and tree-ring sites used in this study. Note: The candidate tree-ring series are shown as green dots, and the two tree-ring sites finally selected in the reconstruction models are marked with a black dot. (b) Comparison showing the strong similarities between regional records of winter maximum snow water equivalent data from high elevation sites, winter rainfall totals in central Chile, and mean annual river discharges on both sides of the Andes over the 1951–2010 period covered by the snowpack series.

show that snowpack variations have some local differences but an overall strong common signal on both sides of the Andes. These variations are very closely linked to variations observed in rainfall data in central Chile and river discharges in the Chilean and Argentinean foothills (see Figure 1b). The similarities result from a hydrological system primarily driven by moisture that originates in the Pacific and crosses the Andes in a succession of westerly frontal systems concentrated during the winter months [*Falvey and Garreaud*, 2007]. Melt from the mountain snowpack during the following warm season dominates the regimes of rivers flowing from the Andes to the adjacent semi-arid regions in Chile and Argentina. The strong similarities observed in snowpack, rainfall and streamflow records suggest that it is possible to use a relatively limited number of station records to capture the main hydrologic patterns in this region. Moreover, it is also possible, for a given year, to infer or validate the information provided by one variable (e.g. snowpack) using data from other related variables (e.g. central Chile rainfall and/or Andean streamflows). In this region the contribution of melting glaciers to surface runoff becomes particularly important during extremely dry years, but this contribution is highly variable depending on the size and the glaciated area of the basin under investigation. In the large basins with snowmelt dominated regimes analyzed in this study, the glacial component is of secondary importance.

[4] It is well known that warm El Niño years in the tropical Pacific will likely bring rainy winters and floods in central Chile, above average winter snowpack levels in the mountains, and high river discharges on both sides of the Andes in the following austral spring and summer seasons [*Aceituno*, 1988; *Rutlland and Fuenzalida*, 1991; *Aceituno and Garreaud*, 1995; *Compagnucci and Vargas*, 1998; *Escobar and Aceituno*, 1998; *Montecinos and Aceituno*, 2003; *Masiokas et al.*, 2006; *Garreaud*, 2009; *Cortés et al.*, 2011] (among others). In fact, various analyses of historical ENSO variations have relied on the warm/wet relationship between ENSO and precipitation levels in central Chile to identify and estimate the severity of El Niño years between the 16th and 19th centuries [*Quinn and Neal*, 1983; *Ortlieb*, 1994, 2000]. In contrast, the relationship with La Niña years is not straightforward. Although some extreme dry winters have occurred during La Niña events (most recently, the winter of 2010), several of the driest years on the instrumental record such as 1968, 1990 and 2004 were not associated with cold conditions during winter in the tropical Pacific [see *Masiokas et al.*, 2006]. Recent analyses have also identified a potential influence of the PDO on the low frequency modes of variability in the snowpack and streamflow records of this region [*Masiokas et al.*, 2010; see also *Dettinger et al.*, 2001].

[5] In this paper we develop two new, complementary and annually-resolved quantitative reconstructions of snowpack variations for the high arid Andes of central Chile and central-western Argentina (Figure 1a). These reconstructions cover time intervals of 135 and 852 years and are specifically developed and analyzed separately to illustrate the benefits and limitations of the different proxies available in each case. To our knowledge, this is the first attempt to reconstruct annually-resolved, serially complete snowpack variations spanning most of the past millennium in the Southern Hemisphere. Previous efforts to reconstruct snowpack changes in the study area have relied mainly on

Table 1. Instrumental Data Used in This Paper[a]

Variable	Station	Latitude, Longitude	Elevation (m)	Period	1951–2000 Mean	Data Source
Snowpack[b]	Quebrada Larga	30°43′S, 70°22 W	3500	1956–2010	222 mm	DGA
	Portillo	32°50′S, 70°07′W	3000	1951–2010	637 mm	DGA
	Toscas	33°10′S, 69°53′W	3000	1951–2010	295 mm	DGI
	Laguna Negra	33°40′S, 70°08′W	2768	1965–2010	584 mm	DGA
	Laguna del Diamante	34°15′S, 69°42′W	3310	1956–2010	453 mm	DGI
	Valle Hermoso	35°09′S, 70°12′W	2275	1952–2010	784 mm	DGI
	Lo Aguirre	36°00′S, 70°34′W	2000	1954–2010	943 mm	DGA
	Volcán Chillán	36°50′S, 71°25′W	2400	1966–2010	800 mm	DGA
Rainfall[c]	Los Andes	32°50′S, 70°37′W	815	1907–2000	259 mm	DGA
	Valparaíso	33°01′S, 71°38′W	40	1869–2000	370 mm	DMC
	Peñuelas	33°10′S, 71°32′W	360	1915–2000	614 mm	DGA
	Santiago	33°27′S, 70°42′W	520	1866–2000	314 mm	DMC
	Aculeo	33°53′S, 70°55′W	350	1913–2000	558 mm	DMC
	Rancagua	34°10′S, 70°45′W	500	1910–2000	408 mm	DGA
	San Fernando	34°35′S, 71°00′W	350	1910–2000	687 mm	DGA
	Talca	35°26′S, 71°40′W	100	1907–2000	620 mm	DGA
	Concepción	36°48′S, 73°06′W	10	1876–2000	1093 mm	DMC
Streamflow (river)[d]	Km. 47.3 (San Juan)	31°32′S 68°53′W	945	1909–2007	65.2 $m^3 s^{-1}$	SSRH
	Guido (Mendoza)[e]	32°51′S 69°16′W	1550	1909-2008	48.9 $m^3 s^{-1}$	SSRH
	Valle de Uco (Tunuyán)	33°47′S 69°15′W	1200	1954–2008	28.6 $m^3 s^{-1}$	SSRH
	La Jaula (Diamante)[f]	34°40′S 69°19′W	1500	1938–2008	30.9 $m^3 s^{-1}$	SSRH
	La Angostura (Atuel)	35°06′S 68°52′W	1200	1948–2008	35.2 $m^3 s^{-1}$	SSRH
	Buta Ranquil (Colorado)	37°05′S 69°44′W	850	1940–2008	148.3 $m^3 s^{-1}$	SSRH
	Cuncumén (Choapa)	31°58′S 70°35′W	955	1941–2009	9.6 $m^3 s^{-1}$	DGA
	Chacabuquito (Aconcagua)	32°51′S 70°31′W	1030	1914–2009	33.0 $m^3 s^{-1}$	DGA
	El Manzano (Maipo)	33°36′S 70°23′W	890	1947–2009	107.8 $m^3 s^{-1}$	DGA
	Bajo Los Briones (Tinguiririca)	34°43′S 70°49′W	518	1942–2009	50.7 $m^3 s^{-1}$	DGA

[a]Mean annual values refer to a July–June water year. Data sources: DGA, Dirección General de Aguas, Chile; DGI, Departamento General de Irrigación, Mendoza, Argentina; SSRH, Subsecretaría de Recursos Hídricos, Argentina; DMC, Dirección Meteorológica de Chile.
[b]Snowpack stations used to develop a regionally-averaged, winter maximum snow water equivalent record for the Andes between 30° and 37°S.
[c]Meteorological stations used in the central Chile composite record of cold season (March–November) rainfall variations.
[d]Argentinean and Chilean gauging stations used to create a regionally-averaged record of mean annual river discharges.
[e]July 1909 to June 1956 estimated from Cacheuta (33°01′S, 69°07′W, 1238 m).
[f]July 1938 to June 1977 estimated from Los Reyunos (34°35′S, 68°39′W, 850 m).

historical documents that we incorporate here as validation series; i.e. records of Andean snow conditions for the period AD 1760–1890 and data from a newspaper from the city of Mendoza, in the eastern foothills of the Andes in Argentina, that reported winter snow days and snow height along the main road crossing to Chile between AD 1885 and 1996 [*Prieto et al.*, 1999a, 2000, 2001; see also *Neukom et al.*, 2009]. These proxy records cannot be easily converted to snow water equivalent data, but in the absence of additional quantitative information they provide useful evidence about pre-instrumental snowpack changes. This documentary evidence has also been used by *Gallego et al.* [2008] to assess the influence of large-scale atmospheric variables on snow accumulation in the Andes. We also incorporate information compiled by *Prieto et al.* [1999b; see also *Neukom et al.*, 2009] who used historical records from the city of Mendoza to reconstruct the streamflow variations in Río Mendoza (the city's main water source) between AD 1600 and 1960. Some of the *A. chilensis* tree-ring chronologies available from the western Andean foothills have already been used to reconstruct lowland rainfall variations in central Chile [*Boninsegna*, 1988; *Le Quesne et al.*, 2006, 2009]. However, so far no reconstructions have targeted Andean snowpack variations that have been carefully monitored at several high-elevation sites since 1951. *Neukom et al.* [2010] have utilized some of these historical, instrumental and tree-ring records in their sub-continental assessment of annually resolved, gridded reconstructions of summer and winter precipitation variability in southern South America (i.e. south of 20°S). Their reconstructions extend back to AD 1498 and 1590, respectively.

[6] In this study we specifically compiled and combined an extensive dataset of different precipitation proxies in an attempt to cross-validate the resulting time series and estimate, as reliably as possible, and extend as far back as possible, the information on winter snow accumulation levels in the central Andes of Chile and Argentina. We believe that this complementary approach has not previously been employed in the development and validation of snowpack reconstructions in a mountainous region. Given the crucial socio-economic significance of winter snow accumulation in these semi-arid areas, the development of robust, statistically-validated models specifically designed to reconstruct Andean snowpack changes over most of the past millennium appears as a necessary and important contribution to evaluate the relative severity of recent observed changes in a long-term context and to improve the understanding of this region's hydroclimatic system.

2. Data and Methods

2.1. Instrumental Records

[7] Winter maximum snow water equivalent (MSWE) data from eight high elevation sites (Figure 1 and Table 1) were used to calculate a regionally-averaged record of total winter snow accumulation in the Andes between 30°

Table 2. Tree-Ring Sites Selected as Potential Predictors of Snowpack Variations in the Andes[a]

Site	Latitude, Longitude	Period AD	Cores	Mean Length	Mean r	EPS>0,85	Data Source
El Asiento[b]	32°39′S, 70°49′W	955–2000	114	305	0.675	1425–2000	1, 2
San Gabriel	33°47′S, 70°15′W	1131–1996	70	289	0.613	1170–1996	1, 2
Rio Clarillo	33°49′S, 70°25′W	1204–2002	80	300	0.511	1360–2002	2
Urriola Este	34°27′S, 70°25′W	1205–2001	75	321	0.603	1450–2001	2
Urriola Oeste	34°27′S, 70°26′W	1500–2001	19	323	0.614	1660–2001	2
El Baule	34°29′S, 70°26′W	1146–2001	49	386	0.585	1315–2001	2
Agua de la Muerte[b]	34°31′S, 70°25′W	1072–2001	108	368	0.609	1150–2001	2
Santa Isabel	34°47′S, 70°45′W	1613–1975	64	212	0.614	1650–1975	1
Melado	35°52′S, 71°01′W	1270–2005	108	237	0.573	1490–2005	2
Polcura	37°04′S, 71°25′W	1144–2004	154	242	0.622	1315–2004	2
Huinganco	37°08′S, 70°04′W	190–2000	162	309	0.532	1145–2000	1, 3
Laja	37°20′S, 71°32′W	1418–2005	105	221	0.655	1540–2005	2
El Chacay	37°21′S, 71°03′W	1641–1975	16	234	0.739	1680–1975	1
Nitrao	37°04′S, 71°02′W	1339–2005	119	245	0.557	1475–2005	2
Ralco-Lepoy	38°04′S, 71°19′W	1585–2005	88	223	0.563	1665–2005	2

[a]The number of tree-ring cores used to build each chronology and the mean length (in years) of these cores are indicated. Mean r is the mean correlation between these series. EPS > 0.85 indicates the period for which the tree-ring chronologies have acceptable signal strength [*Wigley et al.*, 1984]. Data sources: 1, International Tree-Ring Data Bank (ITRDB); 2, Laboratorio de Dendrocronología, Universidad Austral de Chile; 3, Laboratorio de Dendrocronología, IANIGLA-CONICET, Mendoza, Argentina.

[b]Record finally included in the regression models.

and 37°S. The dataset has been extended and updated from the one used by *Masiokas et al.* [2006, 2010], and contains the longest and most complete snowpack records for this portion of the Andes. As discussed in these previous studies, the individual series show strong similarities at inter-annual and inter-decadal scales. Prior to computing the regional average, each MSWE series was converted to standardized anomalies (Z scores) using the 1951–2000 reference period. This composite snowpack series covers the interval 1951–2010 and was adjusted to account for temporal changes in variance due to changes in sample size [*Osborn et al.*, 1997]. For any given year, adjustments factors were calculated based on the number of series available and the mean correlation between all possible pairs of records. Regional records of cold season (March–November) rainfall totals in central Chile and mean annual (July–June) streamflow variations on both sides of the Andes were developed using a similar approach. The regional rainfall composite was calculated from the nine longest and most complete records available between 30° and 37°S, and covers the period 1866–2000. The regional streamflow record is based on 10 stations and extends over the 1909–2009 interval (Table 1). In all cases, the variance-adjusted series are extremely similar to the unadjusted composites due to the strong correlations existing between the original records.

2.2. Tree-Ring Records

[8] Tree-ring width chronologies from *Austrocedrus chilensis* (Ciprés de la Cordillera) trees constitute the main natural indicator available for estimating annually-resolved pre-instrumental hydrologic variability in the study area. This long-lived species generally shows a strong, positive response to precipitation levels and excellent dendrochronological characteristics [*Schulman*, 1956; *LaMarche et al.*, 1979; *Boninsegna*, 1988; *Boninsegna et al.*, 2009; *Villalba et al.*, 1998; *Le Quesne et al.*, 2006, 2009; *Lara et al.*, 2008; *Christie et al.*, 2011]. We first compiled all available *A. chilensis* tree-ring chronologies from the Andes between 32° and 38°S (Figure 1 and Table 2). A few, recently collected subfossil samples were processed and added to the El Asiento chronology to improve the sample replication in the earliest portions of the record.

[9] After removing series shorter than 100 years, the quality of the raw tree-ring measurements in each chronology was assessed using the program COFECHA [*Holmes*, 1983; *Grissino-Mayer*, 2001]. Series with an average correlation <0.3 with the site's "master" chronology (an average built from all remaining series filtered using a 32-year spline) were discarded to ensure a strong common signal at each site. The program ARSTAN [*Cook and Holmes*, 1984; *Cook and Krusic*, 2005] was subsequently used to remove the non-climatic, age-related trend in the individual tree-ring width records and obtain dimensionless, stationary tree-ring index series [*Fritts*, 1976; *Cook et al.*, 1990]. We used a fixed 150-year spline with a 50% cutoff [*Cook and Peters*, 1981] to standardize each series and preserve approximately the same amount of low frequency variability in all records. This particular standardization method preserves 50% of the variance at wavelengths of 150 years and 99% of the variance at wavelengths of 47 years or shorter. The so-called "standard" chronologies used in subsequent analyses were developed by compositing the individual detrended tree-ring series using a biweight robust averaging technique [*Cook*, 1985] and the variance-stabilization method of *Osborn et al.* [1997]. The strength of the common signal within these chronologies was evaluated using the Expressed Population Signal (EPS) statistic over moving windows of 30 years with 15 years of overlap [*Wigley et al.*, 1984]. According to these authors, time periods where EPS ≥ 0.85 indicate that the individual series composing the chronology have reasonable signal strength. This information is shown in Table 2.

2.3. Historical Records

[10] Several studies have compiled historical information about meteorological events using Spanish colonial archives and other sources. Most have focused on historical variations of rainfall in central Chile, but there are also detailed accounts of snow conditions in the Andes, and analyses of

streamflow variations in the eastern Argentinean foothills (see references above). We used this rich array of documentary evidence to develop two lists of years with snowy to extremely snowy winters or dry to extremely dry winters in the Andes since the 16th century. The historical sources used in the creation of these lists of extreme events are: (1) documentary accounts of snow conditions over the main Andean pass, commonly known as Paso de Cristo Redentor - Los Libertadores, between 1760 and 1890 [*Prieto et al.*, 1999a]; (2) newspaper-based records of snow depth along the main Andean pass between AD 1885 and 1996 [*Prieto et al.*, 2000, 2001]; (3) documentary accounts of the discharge of the Mendoza river (and other rivers) in the Argentinean piedmont between 1600 and 1960 [*Prieto et al.*, 1999b]; (4) the compilation of evidence for wet years in central Chile used by *Ortlieb* [1994] to infer the severity of El Niño events between AD 1535 and 1900.

[11] The 1866–2000 regional rainfall series developed in section 2.1 was also examined to improve the identification of extreme events during the 19th and 20th centuries and to be able to extend the existing extreme event lists to the year 2000. Superposed epoch analysis [*Holmes and Swetnam*, 1994] was used to test the skill of the tree-ring based regression model in reconstructing snowpack in these extreme years. For each list of extreme years, mean snowpack values were calculated for 7-year windows including three years before and after each event. The mean series for these years were compared to variations in the complete reconstruction by performing Monte Carlo simulations that randomly selected 7-year sequences, calculated their expected means, and provided bootstrap confidence levels to allow testing the statistical significance of the events in relation to the surrounding years [*Mooney and Duval*, 1993]. A record of total monthly hours of rain in the city of Santiago de Chile for the period 1824–1850 [*Taulis*, 1934] was also used for validation purposes. Here we assumed that the duration of the rainfall events during winter is very likely positively correlated with the amount of precipitation received in the lowlands of central Chile and in the adjacent mountains to the east. Following this hypothesis, extended periods of rain in Santiago during the winters of 1824–1850 should correspond with higher than normal snow accumulations in the Andes, and vice-versa.

2.4. Reconstruction Strategies

2.4.1. Reconstruction of Snowpack Using a Regional Rainfall Record as Predictor

[12] The first reconstruction (R1) uses a simple linear regression model in which the predictand is the 1951–2010 regional snowpack record and the predictor is the 1866–2000 regional cold season rainfall record from central Chile. Both records are expressed as standardized anomalies calculated from the means and variances of their overlapping 1951–2000 periods. Given the relative shortness of the common period between series (50 years), the final reconstruction model was developed using a "leave-one-out" cross-validation procedure [*Michaelsen*, 1987]. Linear regression models for each year were successively calibrated on the remaining 49 observations and then used to estimate the predictand's value for the year omitted at each step. This resulted in 50 predicted values which were compared to the actual snowpack observations to compute validation statistics of model accuracy and error. A regression model based on the full calibration dataset (1951–2000) was finally used to reconstruct the snowpack values over the complete period covered by the regional rainfall time series.

[13] The goodness of fit between observed and predicted snowpack values was tested based on the proportion of variance explained by regression and the normality, linear trend, and the first- and higher-order autocorrelation of the regression residuals. As an additional measure of regression accuracy we also computed the Reduction of Error statistic (RE) [*Fritts*, 1976; *Cook et al.*, 1994]. The root mean square error of validation (RMSEv) [*Weisberg*, 1985] was calculated and used as an estimate of the uncertainties of the reconstructions. Independent verification of the regression equation was undertaken by correlating the reconstructed MSWE values to the composite records of mean annual river discharges developed for the region.

2.4.2. Reconstruction of Snowpack Using Tree-Ring Records

[14] The reconstruction of snowpack variations from tree rings (R2) was developed as follows. All available tree-ring chronologies ending in 1999 or later were lagged at t-1, t = 0, t + 1 and t + 2 and initially considered potential predictors of the regional snowpack series. The lagging approach was used to capture any possible tree-growth preconditioning and lagged responses to climate that are often seen in tree-ring series [*Fritts*, 1976]. All these potential predictors were then correlated against the regional snowpack record and only those chronologies that correlated positively at $p < 0.01$ with the predictand data were selected for further analysis. This significantly reduced the number of potential predictors to only two series at t=0 (the El Asiento and Agua de la Muerte chronologies) which are among the longest and best replicated tree-ring chronologies from central Chile (Figure 1 and Table 2). As these two chronologies show strong collinearity, we averaged these series and used the resulting record as a single predictor of snowpack in a linear regression model. We also tested numerous other options for selecting the best possible tree-ring predictors and regression models for reconstructing the instrumental snowpack record (results not shown). These options include the creation of regional composite chronologies and the use of individual series entered in a stepwise approach into regression models, the use of different correlation thresholds ($p < 0.05$ and $p < 0.1$) for selection of the potential predictors for the models, the prewhitening and log-transformation of predictors and predictand series prior to correlation, and the use of Principal Component Analysis (PCA) to reduce the number of well correlated predictors and concentrate their common information further. Interestingly, these trials showed that the best regression results were obtained with one of the most parsimonious models (i.e. the average of the two chronologies that best correlate with the snowpack series). All trials were cross-validated using the leave-one-out approach described above. Independent verification of results was performed by comparing the predicted values with (1) the regional streamflow record, (2) the series of total hours of rain during winter in Santiago, and (3) superposed epoch analyses based on the lists of extreme dry and extreme wet years as described in section 2.3.

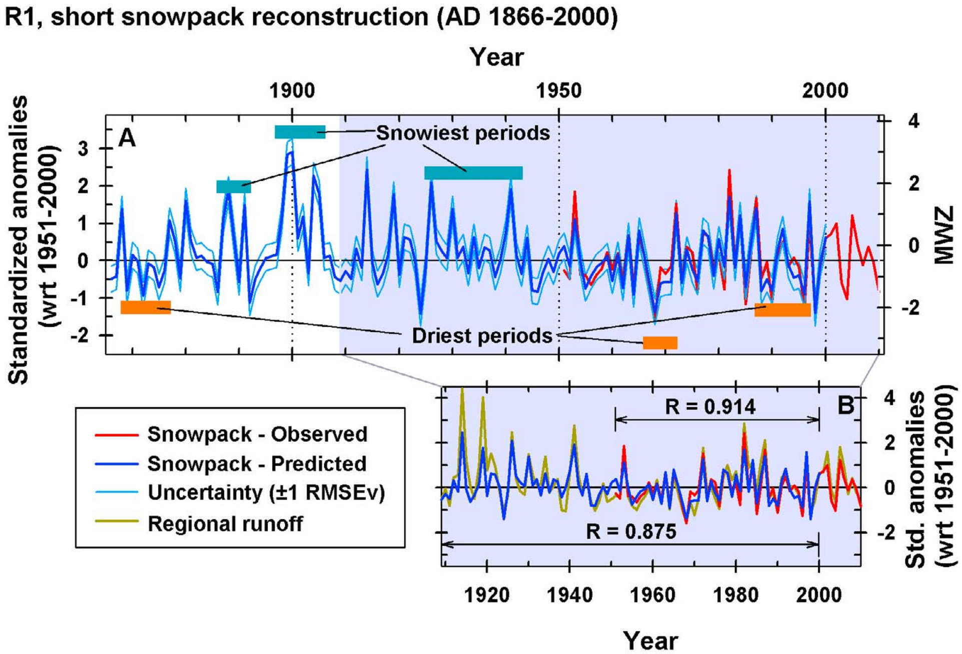

Figure 2. (a) Reconstruction of snowpack changes in the Andes between 30° and 37°S using 1866–2000 winter rainfall records from central Chile. Note the very strong similarities between the rainfall-based predictor series (blue line) and the 1951–2010 regionally-averaged snowpack record (red line). The estimated uncertainty of the reconstructed series (±1 RMSEv) is indicated by light blue lines. Turquoise (orange) bars show the magnitude and location of the three snowiest (driest) periods in the snowpack reconstruction as determined by the running MWZ method (see text for details). (b) Comparison of the observed and predicted snowpack records with a 1909–2009 regional streamflow composite series. This provides further evidence of the regional representativeness and reliability of the reconstructed series.

2.4.3. Assessment of Intra- to Multi-decadal Patterns in the Reconstructions

[15] A recently developed time series analysis technique [*Mauget*, 2003, 2004, 2011] that is novel to paleoclimatic studies was used to identify and test the statistical significance of the most important intra- to multi-decadal (IMD) ranking regimes in the snowpack reconstructions. Here, IMD time scales are considered to include regime periods of *Ns* = 5 to 30 years duration. This relatively simple technique is based on the calculation of Mann Whitney Z (MWZ) statistics over moving time windows and allows the identification of the onset, duration and statistical significance of dry or snowy IMD periods in the snowpack series. As implemented here, this technique (hereafter, the "running MWZ method") first ranks the annual values in a time series, samples those rankings over moving time windows of fixed *Ns* year duration, and then converts each sample to a Mann-Whitney U (MWU) statistic [*Wilcoxon*, 1945; *Mann and Whitney*, 1947]. At a fixed sample size, null distributions of MWU statistics derived from randomly-sampled rankings are normally distributed and proportional to the incidence of high rankings in the sample [*Mendenhall et al.*, 1990; *Wilks*, 1995]. Thus the MWU statistics from sequences of a time series' ranked values can be normalized into MWZ statistics using the mean and standard deviation of an appropriate MWU null distribution. Here, null distributions for an *Ns* year sample size were calculated via Monte Carlo sampling of the rankings of autoregressive red noise time series with the same persistence characteristics of pre-filtered snowpack series. As this pre-filtering removed IMD variance and trends, the null hypothesis associated with the MWZ statistics holds that snowpack time series represent semi-random climate variations with inter-annual persistence, but are essentially stationary and trendless with no significant IMD ranking regimes (see *Mauget* [2011] for more details). As a result, MWZ statistics associated with a significant incidence of high or low rankings in a sample are defined by ±1.960, ±2.576 and ±3.291 MWZ values, at 95%, 99% and 99.9% confidence intervals respectively. Thus significant MWZ statistics from the moving window analysis can identify high or low snowpack periods of *Ns* years' duration in the original series, relative to a null hypothesis that assumes stationary hydroclimatic variability. To identify periods of high and low ranked snowpack values of more arbitrary length, this process is repeated using sampling windows of *Ns* = 5–30 years. The technique then identifies the periods with the largest negative or positive MWZ statistics (i.e. the driest and snowiest intervals in the record) from all possible non-overlapping window lengths.

3. Results

[16] The short snowpack reconstruction (R1) is based on a linear regression model that uses a single, supra-long central Chile regional rainfall record as predictor (Figure 2a). This simple model is able to explain 84% of the regional snowpack

Table 3. Calibration/Verification Statistics and Details of the Two Models Developed for the Reconstruction of Snowpack Changes in the Andes[a]

Model	Predictors	Adj R^2 (F)	Se	RE	DWd	Trend	RMSEv	Reconstructed Period AD
R1	Central Chile rainfall record	0.836 (244.5)	0.335	0.82	1.91	−0.001	0.346	1866–2000
R2	Composite tree-ring chronology	0.454 (40.7)	0.610	0.41	2.53	0.0004	0.624	1150–2001

[a]Both models were cross-validated using a leave-one-out approach (see text). Notes: Adj R^2, coefficient of determination adjusted to account for the number of predictors in each model, used as an indicator of the proportion of variance explained by regression; F, F-value of regression; Se, standard error of the estimate; RE, reduction of error statistic; DWd, the Durbin-Watson *d* statistic used to test for first-order autocorrelation of the regression residuals (Portmanteau Q tests of higher-order autocorrelation of residuals are not shown but, as with the DW tests, indicate non-significant results for the two models); trend, linear trend of regression residuals; RMSEv, root mean squared error of validation.

series' variance over the 1951–2000 period and shows no apparent sign of model misspecification (Table 3). As expected when calibrating using linear regression, the variance of the proxy time series is less than that of the instrumental data. However, the estimated values also show a very good agreement with the regionally-averaged streamflow series calculated from gauging stations on both sides of the Andes (Figure 2b). The R1 series offers the possibility of reliably extending the information on Andean snow changes back to AD 1866. This timeframe is more than twice the length of the 60-year long instrumental record currently available (Figure 2a).

[17] The R1 (rainfall-based) snowpack reconstruction shows important year-to-year changes which very closely follow those observed in the instrumental snowpack record. One of the most striking features in the AD 1866–2000 proxy series is the peaks formed by the very elevated values in the late 19th to early 20th centuries, and particularly the winters of 1899 and 1900 (Figure 2a). These years are reported as extremely rough and snowy winters in the Andes by several historical accounts [e.g., *Taulis*, 1934]. In Argentina the extreme high snowpack in these years resulted in extraordinary snowmelt volumes and the partial destruction of the main dam that diverts the Río Mendoza waters towards the City of Mendoza and into the main irrigation system of the province (for sources, see section 2.3).

[18] Embedded in the strong inter-annual variability of the reconstructed series there are other extremely severe winters that are between 1.5 and 3 standard deviations above the long term mean (1880, 1888, 1904–05, 1914, 1919, 1926, 1941, 1982, and 1997). The driest winters on record are 1924, 1998, and 1968 (all ca. 1.4 standard deviations below the mean) followed by 1892 and 1872 (Figure 2). These snowiest reconstructed winters and most of these extreme dry events have generally been recognized and discussed in the historical literature and are prominent features in river discharge records on both sides of the Andes (see section 2.3). Based on the running MWZ method, the first, second and third non-

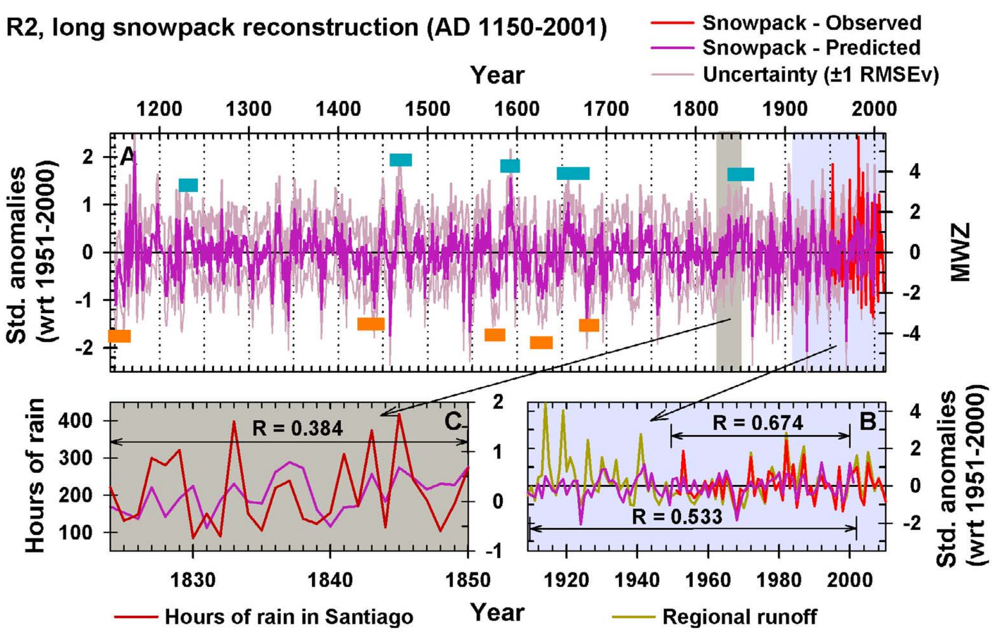

Figure 3. (a) Tree-ring based snowpack reconstruction between AD 1150 and 2001. The estimated snowpack values show statistically significant ($p < 0.05$) positive correlations with (b) the regional streamflow composite series and with (c) historical accounts of total duration of Santiago winter rainfall events. The estimated uncertainty of the reconstructed series (±1 RMSEv) is indicated by light purple lines. As in Figure 2a, turquoise (orange) bars show the magnitude and location of the five snowiest (driest) periods in the snowpack reconstruction as determined by the running MWZ method (see text for details).

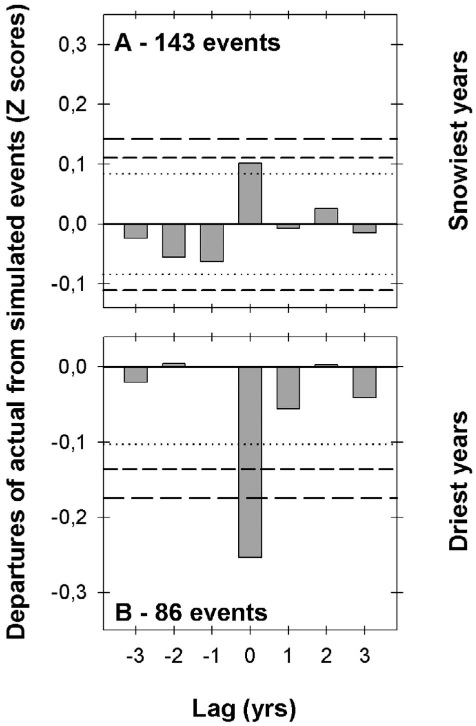

Figure 4. (a) Superposed epoch analysis of the snowiest years identified from historical and instrumental sources between AD 1535 and 2000 and the estimated values derived from the tree-ring based reconstruction (R2). The values shown represent differences between the actual and simulated values three years prior and three years following the extreme winters. Statistical significance is estimated from a bootstrap simulation of 1,000 trials of randomly selected sequences of seven years in the reconstruction. Dotted, short-dashed and long-dashed lines represent the 95%, 99% and 99.9% significance levels, respectively. (b) Same as Figure 4a, but for the series of extreme dry winters identified in section 2.3. Note that dry winters at t = 0 are better captured than snowy winters by this reconstruction model.

overlapping periods with the highest concentration of extreme snowy winters during the past 135 years are AD 1898–1905, 1926–1942, and 1887–1891 respectively. The three driest intervals are 1967–1971, 1988–1996, and 1869–1876 (Figure 2a).

[19] The linear regression model that uses the single composite tree-ring series as predictor (R2) explains over 45% of the variance of the regional snowpack record over the 1951–2001 calibration period and allows the reconstruction of snow accumulation values back to AD 1150 (Figure 3a). The model performs reasonably well in the calibration and verification tests, with regression residuals showing neither trend nor first and higher order autocorrelation that would denote serious model misspecification (Table 3). Comparison of the reconstructed values with the regionally-averaged streamflow series and the record of hours of rain in Santiago indicate statistically significant ($p < 0.05$) positive correlations of 0.533 and 0.384, respectively (Figures 3b and 3c).

[20] The lower skill of the R2 model compared to the R1 model appears to be partially related to the limited performance of this model in predicting extreme high snowpack values in the instrumental series (Figures 3b and 3c). This is very likely due to the differential tree-growth response to extreme dry versus extreme wet or snowy conditions: As with other precipitation-sensitive tree-ring species, the chronologies usually reflect the dry to extreme dry conditions better than extreme wet years for which the soil moisture content exceeds the physiological needs of the trees [*Fritts*, 1976]. The results from the superposed epoch analyses (Figure 4) support this hypothesis: Although both extreme wet and extreme dry years at t = 0 are properly captured by the tree-ring based reconstruction, the values estimated for the 86 driest years are highly significantly different than expected (well outside the 99% significance level, Figure 4b) whereas the estimates for the snowiest 143 years are only barely significantly different than simulated mean values (Figure 4a). The lower skill of the R2 model is also reflected in the uncertainties associated with the estimated values, which are almost two times larger than those obtained for the R1 model (see SE and RMSEv; Table 3).

[21] Despite these limitations, the very simple reconstruction approach derived from tree-ring data offers not only the opportunity of assessing the main intra- to multi-decadal hydrological patterns changes for the past 850 years, but it also allows testing the relative severity of recent "extreme" conditions in a substantially longer context. The MWZ method indicates that the five driest non-overlapping periods on record occurred in AD 1622–1633, 1150–1161, 1571–1580, 1677–1685 and 1429–1445. Extremely snowy conditions were inferred during AD 1465–1475, 1589–1596, 1652–1674, 1843–1858 and 1229–1236 (Figure 3a). The results from the running MWZ method were also used to evaluate the comparative significance of "moving decades" or running windows of 10-year in length in the two snowpack reconstructions (Figure 5). The running decadal MWZ values in both series show strong similarities that suggest the regression models are relatively reliable in capturing the slowly evolving pattern of fluctuations in the regional snowpack series. Examination of Figure 5 also suggests that, at decadal timescales, the conditions observed during the past 60 years (both for extreme high and low snowpack values) have not been unusual when assessed in the context of the past eight centuries. Longer and more severe droughts than those observed in the instrumental record probably occurred several times during the past millennium. The most extreme dry decades are concentrated between the late 16th century and the mid 18th century; statistically they would be considered highly significantly lower than "normal" conditions (i.e. below the 99.9% significance level, Figure 5). The reconstructions also provide evidence for decade-long periods of high snowpack levels that equaled or probably surpassed those recorded during the past six decades in the Andes. The extreme snowy interval concentrated at the beginning of the 20th century is the most recent period with values that were very likely higher than those observed in the 1980s in the instrumental data. However other periods ending in the late 1100s, during the early 1300s and late 1400s, during the early and late 1600s, and during the mid 1700s and mid 1800s appear to have been at least as extreme as the early 20th century period and well above the highest

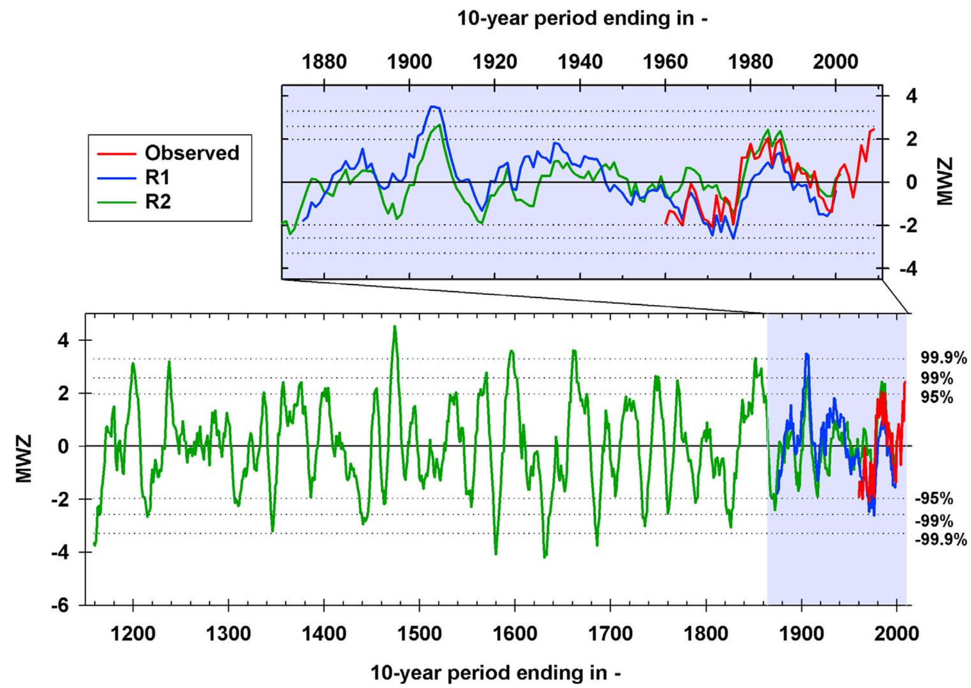

Figure 5. Mann-Whitney Z statistics for running 10-year samples of the regionally averaged snowpack record (red line) and the two reconstructions developed in this study (blue and green lines). Horizontal dashed lines indicate positive and negative significance at the 95%, 99%, and 99.9% confidence levels. Note the overall similarities of decadal variations between the instrumental and reconstructed series.

levels seen in recent decades of the instrumental record (Figure 5).

4. Conclusions and Discussion

[22] This study presents the first reconstruction and quantitative analysis of variations in snow accumulation over the past two to eight centuries in the Andes between 30° and 37°S (Figure 1). The evidence used to develop and validate these complementary reconstructions discussed here includes instrumental rainfall and streamflow data from adjacent lowlands, a variety of documentary records, and century-long tree-ring series of precipitation-sensitive species from the western side of the Andes. To a large extent, all these records share a common regional hydroclimatic signal (see Figure 1), and were specifically collected and combined in an attempt to maximise the length and reliability of information on snowpack variations in the Andes. The verification tests performed on the reconstructions offer some confidence in the reliability of these time series, but due to the varying nature of the proxy data available, the resulting reconstructions contain certain characteristics that can be considered either strengths or limitations depending on the needs of the end-users of these data. For example, the first reconstruction approach based on central Chile rainfall data (R1) shows exceptional skill for modeling Andean snowpack variations (84% of variance accounted for by the model, see Table 3), but is restricted to the period after AD 1866 by the length of the instrumental records (Figure 2). The high explanatory power of this reconstruction model results from the time-stable, strong positive correlation between measurements of winter rainfall in central Chile and Andean snow water equivalent data [*Masiokas et al.*, 2006] (Figure 1b). R1 could be very useful for local water resource managers, decision makers and stakeholders interested in high quality mountain snowpack data as a basis for infrastructure planning and for developing a better understanding of Andean hydroclimatic variability over the past 145 years.

[23] The tree-ring based reconstruction (R2) is less accurate in modeling the instrumental snowpack record over the calibration period (45% explained variance and higher associated uncertainties than the other model; Figures 2 and 3 and Table 3). It is common to the vast majority of tree-ring based climate reconstructions to have a varying but usually large percentage of variance not properly accounted for by the regression models. In this particular study, the regression trials involving tree rings showed that the model captured the normal and below normal snowpack years relatively accurately but failed to capture the full magnitude of the extreme snowy years in the regional snowpack series. Despite the inherent limitations of this regression exercise, the R2 reconstruction offers the unique possibility of assessing in a quantitative and objective manner the history of snowpack fluctuations over the past 850 years in this Andean region (Figure 4) thereby providing the first annually-resolved mountain snowpack reconstruction covering most of the past millennium in the Southern Hemisphere.

[24] Reconstructions of central Chile rainfall patterns based on tree-ring records [*Boninsegna*, 1988; *Le Quesne et al.*, 2006, 2009] may provide complementary information to the snowpack time series presented here. However, as the pre-processing of tree-ring records and the predictor and predictand series used in these previous studies differ from

the ones used here, the comparison of the main temporal patterns of variability observed in each case needs to be exercised with caution. *Le Quesne et al.* [2009], for example, standardized the tree-ring series with the specific purpose of retaining as much low frequency variability as possible in the resulting rainfall reconstructions. They identified extended periods of wet conditions in Santiago de Chile around AD 1500, 1650 and 1850 and a secular drying trend over the past 150 years. An overall negative trend can also be observed in the R1 series presented here (Figure 2) [see also *Vuille and Milana*, 2007] but in this case it starts after the extreme high snowpack values of 1899–1900. The R2 reconstruction shows an extended period of elevated snowpack values in the mid 19th century (AD 1843–1858 is one of the five snowiest on record, Figure 3), but the long term negative trend in the past 100–150 years is not really discernible. However, standardization of the tree-ring series for the R2 model used a 150-year spline to target the IMD variations observed in the regional snowpack record and would not capture the longer, centennial-scale variations. Such differences highlight the need to examine the different hydroclimatic reconstructions available with care taking into account the purposes of each study together with the processing and proxy series used in each case.

[25] The approach used for assessing the main IMD modes of variability in the snowpack reconstructions is novel in paleoclimatic studies and offers very promising perspectives for identifying the most extreme intervals in a given time series while testing their statistical significance in a simple and objective manner [*Mauget*, 2011]. The tests performed using a 10-year period as a reference revealed patterns that were very similar in the instrumental and reconstructed series (Figure 5). This was somewhat expected in the rainfall-based reconstruction, but the similarities with the tree-ring based reconstruction are interesting and suggest that despite the inherent limitation of this reconstruction in capturing extreme high snowpack values (see discussion above), the overall decadal-scale regional patterns of snowpack variations are still represented relatively accurately in this time series. This information could be useful for both local water resource managers and ongoing modeling analyses intended to predict decadal climate patterns [*Meehl*, 2009; *Keenlyside and Ba*, 2010] which are of more direct relevance to society than the more distant, centennial-scale climate projections generally discussed in the literature.

[26] These main temporal patterns identified in the reconstructions can also provide relevant, quantitative information for larger-scale studies involving the main factors affecting hydroclimate in this portion of the Andes. The longer time frames discussed here allow an assessment of the strength and time stability of the relationships with ENSO and the PDO over most of the past millennium and may complement the great number of studies [e.g., *Stahle et al.*, 1998; *Cobb et al.*, 2003; *MacDonald and Case*, 2005; *D'Arrigo et al.*, 2005; *D'Arrigo and Wilson*, 2006; *Gergis and Fowler*, 2009] that discuss the regional manifestations and long term variability of these large-scale ocean-atmosphere features. Interesting comparisons are also possible with similar tree-ring based snowpack reconstructions recently developed for western North America [*Pederson et al.*, 2011]. As snowpack variations in this region are also strongly influenced by ENSO and the PDO, establishing long-term linkages between the North American records and the Andean reconstructions developed here may help understand other related processes (e.g. changes in streamflow, glacier recession) that could be sharing a similar forcing north and south of the Equator.

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