1. Introduction

Southern South America is a key area to reconstruct past Southern Westerlies (SW) dynamics through the last glacial-interglacial cycle since it intersects the entire SW domain and offers appropriate terrestrial paleoclimate records. Indeed, the strong positive correlation between zonal wind speeds and local precipitation that exists today throughout the Pacific coast and inland areas (Garreaud, 2007) make this area ideal to track the changes of SW. During the last two decades significant improvement has been achieved in relation to the understanding of the SW dynamics through the last glacial-interglacial cycle all over southwestern South America (e.g. Villagrán and Armesto, 1993; Bennett et al., 2000; McCulloch and Davies, 2001; Heusser, 2003; Markgraf et al., 2003; Moreno, 2004; Abarzúa and Moreno, 2008; Tonello et al., 2009; Lamy et al., 2010; Villa-Martínez et al., 2012). Nevertheless, controversies particularly about the behavior of SW during the LGM and their latitudinal shifts/strength changes...
through the Lateglacial and Holocene, are still a matter of discussion (Lamy et al., 2010; Moreno et al., 2010a; Tonello et al., 2009; Villa-Martínez et al., 2012).

Central Chilean Patagonia (44°–49°S) is a key area for studying past SW dynamics since its position at their northern margin (particularly in summer) and within the maximal zonal flow (Garreaud, 2007). In fact, pollen and charcoal paleoecological records from Central Chilean Patagonia have been proven to be useful for reconstructing the position and/or strength changes of the SW in the past (Lumley and Switsur, 1993; Bennett et al., 2000; Haberle et al., 2000; Haberle and Bennett, 2004; Markgraf et al., 2007; Villa-Martínez et al., 2012). However, the paleoclimatic inferences and timing of changes from Central Chilean Patagonia records do not completely agree so it is at present difficult to picture a common regional climatic dynamics since deglaciation.

The Taitao Peninsula and Chonos Archipelago records (Fig. 1a) suggest the presence of Ericaceous heath and grasslands associated to cold and dry conditions from 17 to 15 cal ka BP. Then, the Nothofagus dombeyi type, Pilgerodendron and Podocarpus rainforest developed indicating an increase of temperature and precipitation up to modern conditions around 11 cal ka BP. The rainforest persisted without variation until today except for a sudden increase of Weinmannia trichosperma and Tepualia stipularis between 11 and 6 cal ka BP which was interpreted as warmer than present conditions. Since 6 cal ka BP similar to present rainforest and climate conditions established (Bennett et al., 2000; Haberle and Bennett, 2004).

The Mallín Pollux record (Fig. 1a) documents a sparse scrub steppe associated to cold and dry conditions between 18.6 and 14 cal ka BP which shifts to a species-rich steppe up to 11 cal ka indicating an increase in effective precipitation. Around 11 cal ka BP, N. dombeyi type forest begin to develop associated to high fire activity until 7.5 cal ka BP suggesting increased precipitation and marked seasonality. Similar-to-present N. dombeyi type forests and climatic conditions established by 7.5 cal ka BP (Markgraf et al., 2007).

The Lago Augusta record (Fig. 1a) point out vegetation integrated by herbs, shrubs and evergreen rainforest taxa (Fitzroya, Pilgerodendron and Podocarpus) between 16 and 15.6 cal ka BP associated to cold and wet conditions. Since 15.6 cal ka BP a gradual development of the N. dombeyi type forest associated to evergreen rainforest taxa suggest cool and wet conditions. Around 11.8 cal ka BP the presence of a dense N. dombeyi type forest associated with the disappearance of rainforest taxa, high fire activity and laminated carbonate deposition suggest a warm pulse and decline in precipitation. Since 9.8 cal ka BP N.
dombeyi type forest remained with little variation around Lago Augusta (Villa-Martínez et al., 2012).

According to the latter, Lateglacial and Early Holocene climatic dynamics at Central Chilean Patagonia is well reflected but records failed to show changes for mid and Late Holocene. It is possible that all these records might be insensitive at some degree to subtle precipitation changes given their location in the rainforest close to the Pacific Ocean coast or within the Nothofagus forest at the eastern flank of the Andes. Both vegetation units experiment highly enough annual rainfall amounts (3400–700 mm) so water is not a critical factor and hence slight changes in precipitation directly related to the SW past dynamics would be under-represented in these records.

The present paper consists of a 19 cal ka BP (calendar thousand years before present) pollen and charcoal record from Lake Shaman located in the modern forest-steppe ecotone at the eastern flank of the Andes at Central Chilean Patagonia (44° S; 71° W, upper Río Cisnes valley). Shaman record provides the chance to test the past forest-steppe ecotone dynamics which is mainly modulated by the amount of precipitation in which is associated with the SW dynamics and, secondarily by ecological and human factors. Therefore, this paper aims to reconstruct vegetation, fire and climate dynamics at Central Chilean Patagonia forest-steppe ecotone during the last 19 cal ka. The results will be integrated to other records of Central Chilean Patagonia in order to determine local/regional climatic signatures. The results will be also compared with pertinent extra-regional South American records (North and South Patagonia) in order to discuss the changes of dynamics in the SW during the last glacial-interglacial cycle.

2. Environmental setting

Río Cisnes valley (44°40’S; Fig. 1) is a hydrographic basin product of glacial erosion which presents several paleo-lake terraces and a LGM morainic complex in the upper part of the valley that constitutes the present Chile-Argentina international border (Fig. 1b) (Steffen, 1909; Quensel, 1910; Caldenius, 1932). Lake Shaman (44°26’S; 71°11’W, 919 m asl) is a closed small lake (150 ha) situated in an intermorainic depression in the proximal bank of the Lago de Shaman using a modified microscope to determine glass

- Evergreen Nothofagus betuloides forest dominated (Nordpata-gonic forest; Heusser, 2003): by N. betuloides, D. winteri and P. nubigena associated with shrubs such as Desfontainia spinosa accompanied by Blechnum magellanicum, Fuchsia magellanica and Pseudopanax lartevidens.

- Deciduous Nothofagus pumilio forest: dominated by Nothofagus pumilio associated with Ribes cuccullatum to the west and Berberis illicifolia to the east. Associated shrubs include Chlo- trichum rosmarinifolium, Escallonia alpina, Berberis serrato-dentata, Myoschilos oblonga, Maytenus disticha and Gaultheria myrtilloides var. nana and herbs such as Rubus geoides, Adenocaulon chilense and Valeriana laxiflora.

- Deciduous shrubland is dominated by Nothofagus antarctica associated with Berberis microphylla, Ovidia andina and Ribes magellanicum and herbs such as Fragaria chiloensis, Geranium berterianum and Osmorhiza chilensis.

- Festuca Pallescens grass steppe with Mulrum spinosum and Acana splendens accompanied by Bromus setifolius, Festuca argentina, Hordeum comosum, Vicia magellanica, Polygonula dar-winiana, Acanae magellanica, Oreoplos sp., Cerastium arvense and Calceolaria biflora.

Lake Shaman is located at the forest-steppe ecotone between the N. antarctica and Berberis sp. shrubland and the F. pallescens grass steppe.

Apart from the climate, vegetation distribution, structure and composition at the Río Cisnes valley have also been determined by fire dynamics and human disturbance to some degree.

The historic frequency of fire and its dynamics in the forest-steppe ecotone at the Río Cisnes valley is unknown. Nevertheless modern studies suggested that low-severity surface fires are likely to have played some role in shaping Nothofagus shrubland and steppe ecosystems in Central Patagonia (Veblen et al., 2008). For instance, N. antarctica shrublands (40–43°S) are quite prone to be affected by fire due to its structure and rapid fuel recovery (Mermoz et al., 2005).

On the other hand, the forest-steppe ecotone of the Río Cisnes valley has been intensely affected by human disturbance since the beginning of the 20th century (around 1904) in order to allow sheep breeding (~50,000 animals) (Martinc, 2005). Thus, massive intentional fires and loggings were conducted to clear the deciduous forests close to Río Cisnes Ranch (Fig. 1b), having as consequence the sprouting of N. antarctica accompanied by a strong enrichment of foreign species, such as Muehlenbeckia hastulata, immediately after fire and Rumex acetosella and Verbascum thapsus later (Donoso et al., 2006). Sheep overgrazing triggered a progressive succession process in the grass steppe in which shrubs become dominant at expense of grasses (Aguiar et al., 1996).

3. Materials and methods

In October 2006, a 613-cm-long sediment core (LS0604A) was recovered from the wetland area located at the current shore of Lake Shaman using a modified Livingstone piston corer.

The stratigraphy of the core was characterized by naked-eye lithological description and X-radiographs. Subsamples were separated for loss-on-ignition (LOI), tephra, pollen and macroscopic charcoal analysis. LOI was performed at 1-cm intervals along the sediment core in 1 cm³ of material to determine organic and inorganic (carbonates and clastic fraction) contents (Begtsso and Enell, 1986; Heiri et al., 2001). Preliminary tephra analysis was performed on sediment subsamples from six white bands identified in the X-radiographs. The sediment subsamples for tephra analysis were washed in water to remove organics and clay, and examined under a petrographic microscope to determine glass
color and mineral content. Bulk tephra were crushed to a fine powder and dissolved in a dilute solution of hydrofluoric acid for chemical analysis by Inductively Coupled Plasma Mass Spectroscopy (ICP-MS) techniques in the Laboratory for Environmental Geochemistry at the University of Colorado. Repetitive analysis of internal laboratory standards indicates that the trace-element concentrations are precise to ±10% at the determined concentration levels.

LS0604A chronology was based on eight radiocarbon dates, calibrated using CALIB 5.01 program (Stuiver et al., 2005). With this purpose, dates younger than 9560 years 14C BP were calibrated using CALIB 5.01 program (Stuiver et al., 2005) and for older dates, the Northern Hemisphere curve (SHCal04) (McCormac et al., 2004) and for older dates, the Northern Hemisphere curve was applied (IntCal04) (Reimer et al., 2004). Assuming instantaneous deposition of clastic and tephras levels, adjusted depths were used to construct an age model by subtracting them. The age–depth model was performed using MC2Age program applying a cubic smoothing spline and a bootstrap approach (Higuera et al., 2009). The confidence intervals for each age–depth model were derived from 5000 bootstrapped chronologies, reflecting the combined uncertainty of all age estimates in the model. The final chronology represents the median age at each depth from the 5000 bootstrapped chronologies.

Pollen analysis was performed from 1 cm^3 of sediment samples which were taken at 5-cm intervals. Standard laboratory techniques were applied for pollen extraction from the sediments (Faegri and Iversen, 1989) including KOH 10%, sieving (120 μm mesh), hot HF 40% (80 °C), and acetylation, followed by ultrasonic treatment. Tablets of the exotic spore Lycopodium clavatum were added to each sample to calculate pollen concentration (grains cm^-3) (Stockmarr, 1971) and pollen accumulation rates (grains cm^-2 yr^-1). Pollen grains were identified at 400 and 1000× magnification of the basic pollen sum for each level includes at least 300 terrestrial pollen grains per sample. Pollen percentages of terrestrial taxa were based on the sum of trees, shrubs, herbs and grasses. Aquatic and paludal and fen taxa were calculated from a superset that included the basic pollen sum and the sum aquatic and paludal taxa or the sum Pteridophytes, respectively. CONISS cluster analysis (Grimm, 1987) was performed to divide the pollen diagram in zones considering all local terrestrial pollen taxa ≥2%. Pollen percentages and pollen accumulation rate calculations were complementarily used to reconstruct past vegetation.

Macroscopic charcoal analysis was performed to reconstruct the local fire regime at Lake Shaman. 2 cm^3 of sediment at contiguous 1 cm-interval were processed following sieving methods outlined by Whitlock and Larsen (2001). The charcoal fractions (125 and 250 μm) were tallied in gridded Petri dishes under a stereomicroscope. Charcoal concentration (particles cm^-3) was calculated from the result of particle counting. The charcoal concentration and deposition time obtained from the age model were interpolated at 31 yrs bins (the median temporal resolution of the record; yrs sample^-1) using the Char Analysis software (Higuera et al., 2009) followed by the calculation of charcoal accumulation rate (CHAR; particles cm^-2 yr^-1). Two components of the charcoal series were distinguished: (1) the background component (Cback), which documents extra local or regional fire episodes and secondary charcoal mixed with the sediment in years without fires, and (2) the peak component (Cpeak) which represents the local fire-episodes from high frequency of charcoal deposition in the record (Power et al., 2008). Cback was calculated applying the regression method of locally weighted scatterplot smoothing (LOWESS), robust to outliers considering a window width of 700 years. The Cpeak was calculated as the difference between the charcoal interpolated and the Cback. As Cpeak presents a noise component that proceeds from analytical and sedimentological variables, a Gaussian mixture model was used to identify the noise distribution choosing the 99th percentile (Higuera et al., 2009). Then, fire-episode magnitude and frequency were inferred based on a 1000 yr moving window. Finally, the macroscopic charcoal record was reported according to three major categories based on fire-episode frequency values: null or low (N–L; 0–3 peaks 1000 yr^-1), moderate (M; 3–5 peaks 1000 yr^-1), and high (H; >5 peaks 1000 yr^-1).

4. Results

4.1. Chronology and deposition time

The chronology was based on eight AMS radiocarbon dates obtained from plant macro-remains inside the sediment cores (Table 1) and assignment of a modern age to the surface of the wetland. The depth–age model suggests a continuous deposition without hiatus or erosive discontinuities at Lake Shaman since 19 cal ka BP (Fig. 2). Nevertheless, some sedimentation rate variations are found along the core.

The highest (lowest) sedimentation rate (resolution) values around 0.9 cm yr^-1 (5 yrs cm^-1) are recorded between 19 and 18.5 cal ka BP and decrease (increase) to 0.02 cm yr^-1 (40 yrs cm^-1) around 18 cal ka BP. Between 18 and 3.5 cal ka BP, sedimentation rate (resolution) remains almost steady around 0.02–0.03 cm yr^-1 (30–50 yrs cm^-1) peaking at 3500 cal yr BP up to 0.05 cm yr^-1 (15 yr cm^-1) and decreasing (increasing) to 0.04 cm yr^-1 (20 yrs cm^-1) the last 3 cal ka BP (Fig. 2).

4.2. Sediment description: lithology and loss on ignition

LS0604A core sediments consist of dusky brown silty gyttja between 613 and 595 cm; brown gyttja between 595 and 356 cm intermingled with olive silt layers which decrease in frequency up-core and with six tephra layers; moderate brown gyttja with black spots between 356 and 38 cm; peat between 38 and 2 cm; and the modern wetland soil the last top 2 cm (Fig. 3).

Changes in the values of loss-on-ignition analysis are closely related to the lithology (Fig. 3). Organic matter percentages

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Table 1
Radiocarbon dates of LS0604A core from Lake Shaman. *Adjusted depth without tephra and clastic layers.

<table>
<thead>
<tr>
<th>Lab Code</th>
<th>Depth (cm)</th>
<th>Depth * (cm)</th>
<th>Material</th>
<th>Age (14C yr BP)</th>
<th>±1σ</th>
<th>13C (‰)</th>
<th>Age (cal yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beta237634</td>
<td>85</td>
<td>80</td>
<td>Plant macro-remains</td>
<td>1940</td>
<td>40</td>
<td>-28.4</td>
<td>1827</td>
</tr>
<tr>
<td>Beta234011</td>
<td>122</td>
<td>114</td>
<td>Plant macro-remains</td>
<td>3000</td>
<td>40</td>
<td>-28.0</td>
<td>3111</td>
</tr>
<tr>
<td>Beta237635</td>
<td>195</td>
<td>183</td>
<td>Plant analysis of</td>
<td>3910</td>
<td>50</td>
<td>-26.7</td>
<td>4275</td>
</tr>
<tr>
<td>Beta234012</td>
<td>260</td>
<td>245</td>
<td>Plant macro-remains</td>
<td>7590</td>
<td>40</td>
<td>-26.3</td>
<td>8357</td>
</tr>
<tr>
<td>Beta237636</td>
<td>316</td>
<td>288</td>
<td>Plant macro-remains</td>
<td>9560</td>
<td>70</td>
<td>-27.4</td>
<td>10,824</td>
</tr>
<tr>
<td>Beta234010</td>
<td>392</td>
<td>361</td>
<td>Plant macro-remains</td>
<td>11,370</td>
<td>40</td>
<td>-15.5</td>
<td>13,241</td>
</tr>
<tr>
<td>Beta237617</td>
<td>489</td>
<td>457</td>
<td>Plant macro-remains</td>
<td>15,120</td>
<td>100</td>
<td>-15.7</td>
<td>18,474</td>
</tr>
<tr>
<td>Beta224300</td>
<td>599.5</td>
<td>555.5</td>
<td>Plant macro-remains</td>
<td>15,740</td>
<td>50</td>
<td>-15.2</td>
<td>18,951</td>
</tr>
</tbody>
</table>

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Author's personal copy
associated with the dusky brown silty gyttja and the brown gyttja between 613 and 356 cm range between 3.2 and 32%. From 356 cm to 38 cm (moderate brown gyttja), organic matter percentages increases up to 78% but is significantly variable due in part to the presence of tephra layers which are actually represented as maximum values (peaks) of inorganic density (gr cm⁻³). Organic matter percentages rise almost up to 100% at the top of the core associated with the peat layer and the modern wetland soil. Inorganic density is low (average value ~0.09 gr cm⁻³) except for those values related to the tephra layers and some clastic levels (Fig. 3).

The six sediment sub-samples from the white bands seen in the X-radiographs and represented as inorganic peaks in LOI analysis (Fig. 3) were identified as tephra layers. The origin of the tephras is related to several volcanoes from the Andean Southern Volcanic Zone in the study area (Fig. 1). The Barium (Ba) versus Lanthanum (La) contents of the six tephra are plotted in Fig. 4 along with the composition fields as indicated by available data, which are only preliminary in nature, for the likely possible tephra source volcanoes Melimoyu, Mentolat and Hudson west of Coyhaique (Fig. 1). The identification of the specific source volcano and eruption for each of the analyzed tephra were determined by a combination of both this chemical data as well as the age of the tephra. Tephra α is considered to be derived from the Melimoyu volcano, and its age and thickness is consistent with the age of eruption MEL2 dated by Naranjo and Stern (2004) as <1750 ± 80 ¹⁴C yrs BP (Fig. 2). Tephra β is considered to be derived from the Hudson volcano, and its age and thickness is consistent with the age of eruption H2 dated by Naranjo and Stern (1998) as ~3600 ± 200 ¹⁴C yrs BP (Fig. 2). The other tephra (c, d, e and h) are considered to be derived from the Mentolat volcano based on their chemistry (Fig. 4) (López-Escobar et al., 1993; Naranjo and Stern, 1998, 2004; Stern, unpublished data). Alternatively, some of these later tephra may have been derived from the eruptions that produced small basaltic cones located along the Liquiñe-Ofqui fault east of Puyuhuapi, but it is unlikely that these eruptions generated regionally widespread tephra layers, and there is no trace-element data available from these cones to compare with.
Mulinum present the highest PAR values (15 grains cm$^{-2}$ yr$^{-1}$) since 7 cal ka BP. Asteraceae subf. Asteroideae (84 grains cm$^{-2}$ yr$^{-1}$) and Acaena (20 grains cm$^{-2}$ yr$^{-1}$) highest PAR is recorded during the last 0.3 and 0.1 ka BP, respectively.

*N. dombeyi*-type PAR increases around 14.3 cal ka BP with two peaks between 5 and 3 cal ka BP (800–130 grains cm$^{-2}$ yr$^{-1}$) and 1.3–0.5 cal ka BP (450–170 grains cm$^{-2}$ yr$^{-1}$). A decrease of *N. dombeyi*-type PAR (up to 130–50 grains cm$^{-2}$ yr$^{-1}$) is recorded between 2.2 and 1.5 cal ka BP. Poaceae shows the highest PAR values (up to 522 grains cm$^{-2}$ yr$^{-1}$) between 4.7 and 2.3 cal ka BP while two other peaks (200–30 and 380–120 grains cm$^{-2}$ yr$^{-1}$) are recorded at 12.9–9.1 and 1.7–0.6 cal ka BP. A significant decrease of Poaceae PAR is recorded since 2 cal ka BP (≈ 120 grains cm$^{-2}$ yr$^{-1}$) (Fig. 6).

4.4. Fire record

Charcoal record points out very low magnitude and frequency of fires in Lake Shaman between 19 and 10.5 and particularly at 8–7.1 cal ka BP when the time between fires was maximum and the lowest CHAR were recorded (<4 particles cm$^{-2}$ yr$^{-1}$) (Fig. 7). Moderate fire activity was recorded at 10.5–8, 7.1–3.5, and 0.3–0 cal ka BP with fire-episode frequency values between 3–5 peaks 1000 yr$^{-1}$ and CHAR values up to 44 pieces cm$^{-2}$ yr$^{-1}$ (Fig. 7). High fire activity was recorded during 3.5–0.3 cal yr BP associated with CHAR values up to 176 pieces cm$^{-2}$ yr$^{-1}$ at 2.8 cal ka BP, the highest fire-episode frequency (7.66 episodes 1000 yr$^{-1}$) and magnitude (14,800 pieces cm$^{-2}$ peak$^{-1}$) (Fig. 7).
5. Discussion

5.1. Palaeoenvironmental history

5.1.1. Stratigraphy and chronology

Sedimentological, X-radiograph, loss-on-ignition and chronological data indicate a continuous deposition in Lake Shaman during the last 19 cal ka BP. Gyttja deposition between 19 and 1 cal ka BP (614–38 cm depth) suggests the development of a permanent lake. Transitions between silty gyttja (614–590 cm depth; 19–18.8 cal ka BP), to olive-colored silty gyttja layers (590–360 cm depth; 18.8–12.2 cal ka BP) and to moderate brown gyttja (360–38 cm depth; 12.2–1 cal ka BP) reflect gradual increases of organic matter and decline of inorganic density (Fig. 3). The presence of silty gyttja and gyttja with olive silt layers, coinciding with the highest deposition rates within the core, could represent aeolian transport of silt-sized particle into the lake. Nevertheless, transport via fluvial processes (runoff) is not discarded because of the still unstable environment and sparse vegetation that surrounded the lake at this time (see Pollen and charcoal records section below). The onset of peat accumulation around 1 cal ka BP (38–0 cm depth) together with increasing values of organic matter indicates a shallowing of the shore of Lake Shaman into a boggy/swampy environment.

5.1.2. Pollen and charcoal records

Lake Shaman record shows a highly dynamic picture of vegetation and fire history since 19 cal ka BP.

The pollen record of Lake Shaman shows dominance of Poaceae and shrubs such as Ericaceae, Pergiza-type, Mulinum, Adesmia and Asteraceae subf. Asteroideae and herbs such as Plantago between 19 and 14.8 cal ka BP (Fig. 5) reflecting the development of a grass steppe with high proportions of shrubs. Modern pollen surveys southwards (50°S) of Lake Shaman area indicate that similar Poaceae-shrub pollen assemblages are found on high plateaus (>1000 m asl) located at the present grass steppe-dwarf shrub steppe ecotone under mean annual precipitation values around 300–350 mm (Mancini et al., 2012). Hence Lake Shaman pollen record suggest colder and drier conditions than present between 19 and 14.8 cal ka BP. High values of Acaena at Lake Shaman at this time (Fig. 5) suggest ground soil conditions under unstable post-glacial environments as also recorded in Vega Nándú (Villa-Martinez and Moreno, 2007) and Mallin Pollux (Fig. 1a; Markgraf et al., 2007) after the initial phase of deglaciation. The low values of Nothofagus and Misdendrum (an exclusive parasite of Nothofagus species) between 19 and 14.8 cal ka BP would reflect the incipient development of Nothofagus forest in the upper Rio Cisnes valley but under the dominance of open or tree-less landscapes. Other possible interpretation of the incipient increase of Nothofagus could be an extra-local signal (long distance dispersion) from small populations of Nothofagus that would have persisted in unglaciated areas close to the Andes along Central Patagonia, supporting the idea of glacial refugia (Markgraf et al., 1995; Premoli, 1997).

A progressive increase of Nothofagus along with a slight decline of shrubs (Ericaceae, Acaena, Perizia type and Mulinum) and steady values of Poaceae occurred from 14.8 to 13.3 cal ka BP (Fig. 5) suggesting the existence of a similar grass steppe with scattered shrubs as before but with the appearance of some trees that points to an effective moisture increase. The appearance (trace values) of Fitzroya/Pilgerodendron at 14.8 cal ka BP and the subsequent increase of P. nubigena (Fig. 5) would indicate the development of the Nord-patagonic forest westwards of Lake Shaman, (eg. at the lower Río Cisnes valley and the Chonos Archipelago; Fig. 1a) supporting the idea of increased effective moisture (Haberle and Bennett, 2004). Vegetation and climate inferences at Lake Shaman between 19 and 13.3 cal ka BP are supported by the N–L fire activity (Fig. 7) probably related to a low biomass accumulation and/or fuel discontinuity that would have prevented the occurrence of fires.

A decline in the steppe shrubs (Ericaceae, Acaena, Pergiza-type) occurred between 13.3 and 11.2 cal ka BP at expense of Poaceae and Nothofagus (Fig. 5) suggesting the development of a parkland (grass steppe with scattered shrubs matrix with stands of Nothofagus trees) around Lake Shaman (Fig. 5) that would indicate a slight increase in effective moisture but still under modern values.

At 11 cal ka BP, the aquatic and paludal taxa show a major change when Myriophyllum percentages suddenly drop and Cyperaceae begins to dominate (Fig. 5). Increase of the Cyperaceae may be a consequence of an extension of Lake Shaman’s shore areas to be colonized by them. These changes may be a consequence of (1) a progressive filled up of the lake’s margin and towards the plain and/or (2) variable increases of temperatures that would have led to short term lower lake levels and therefore exposing larger shore areas than before. The idea of increased temperatures around 11 cal ka BP is supported by the replacement of a cold-dry steppe by the Nothofagus forest at Lake Shaman at 11 cal ka BP (Fig. 5). Besides, the increased fire frequency recorded around 10.5 cal ka BP (Fig. 7) supports the latter and in particular suggests warmer summers than present given that fire dynamics in north Central Patagonia are mainly conditioned today by summer droughts (Veblen et al., 2008).

After 11 cal ka BP, the slight but continuous development of the forest shown by the increase in Nothofagus values together with the presence of the grass steppe and the decline of the shrub taxa percentages and PAR to minimum values (Figs. 5 and 6), reflect the establishment of the forest-steppe ecotone with modern analogues within the Río Cisnes valley (Maldonado, unpublished data). The late changes would reflect the onset of the Holocene.

Moderate fire frequency and intensity are recorded at Lake Shaman (Fig. 7) from 10.5–8 cal ka BP. The increased fire activity could have been triggered by an increase in fuel availability by Nothofagus expansion and fuel continuity by the increase of grasses (Poaceae, 60%) at expense of shrubs in the steppe (Figs. 5 and 6). Besides, the record of the first human occupations in the upper Río Cisnes valley around 11.5 cal ka BP strongly suggest them as additional fire ignition agents (Méndez et al., 2009; Reyes et al., 2009).

Between 8 and 3 cal ka BP, values of Nothofagus up to 65% and the continuous presence of Misdendrum (Fig. 5) suggest the
Fig. 5. Percent pollen diagram for main pollen types from Lake Shaman and CONISS analysis.
development of a dense forest coexisting with a grass steppe with *Mulinum* and Asteraceae around Lake Shaman (Figs. 5 and 6). In fact, the pollen record suggests the easternmost position of the forest-steppe ecotone during the Holocene in the Río Cisnes valley since similar modern pollen assemblages are found westwards Lake Shaman area, in the upper-middle Río Cisnes Valley (Maldonado, unpublished data). This easternmost position of the forest-steppe ecotone accompanied by null-low to moderate fire activity (peak magnitude and frequency) (Fig. 7) would suggest the highest effective moisture for the whole record (highest precipitation) under an unmarked seasonality until 5 cal ka BP. It is likely that fuel to be burned was abundant but the summer was not dry enough to allow burning. Accordingly, Mallín Pollux’s (Markgraf et al., 2007) and Laguna Fácil’s (Chonos Archipelago; Haberle and Bennett, 2004) records show a vegetation change to more mesic forests and reduced fires during the mid-Holocene in the western slope of the Andes at Central Patagonia. Between 5 and 3 cal ka BP, the maximum development of *Nothofagus* (Figs. 5 and 6) associated with an increased fire peak frequency trend that turned from moderate to high around 3.5 cal ka BP and continues throughout the late Holocene (Fig. 7) are recorded at Lake Shaman. This association of vegetation and fire activity suggest the gradual establishment of the dry season that under the same vegetation (fuel) and effective moisture (precipitation) conditions than during the mid-Holocene, would promote fire occurrence. Around 3 cal ka BP, the pollen record shows the retraction of the forest with the concomitant expansion of the steppe (decrease in *Nothofagus dombeyi*-type and increase of Poaceae %; Fig. 5). These changes suggest a decrease of effective moisture and the establishment of the modern forest-steppe ecotone under modern climate conditions. Fluctuating Cyperaceae percentages (ranging between 30 and 60%) since 3 cal ka BP indicates, however, an active lake’s shore dynamics associated at some degree with minor lake level changes that would be reflecting high climatic variability. Low *Nothofagus* values (percentage and PAR; Figs. 5 and 6) recorded around 1.8 cal ka BP would be consequence of a fire-induced change given its coincidence with the highest fire occurrence (high fire frequency) for the whole record (Fig. 7). In fact, widespread fires in the *Nothofagus* forests between (~38–55°S) have been shown to be dependent on drought at monthly, seasonal, annual and supra-annual time scales driven by either decreased precipitation or increased temperatures (Kitzberger and Veblen, 2003; Lara et al., 2003; González and Veblen, 2006). Therefore, decreased effective moisture associated with a high climatic variability and with a widespread human occupation along the Río Cisnes valley (Méndez and Reyes, 2008) since 3 cal ka BP may explain the highest frequency and intensity of fires for the whole Holocene in Lake Shaman and its consequent effect on the forest-steppe ecotone. Similar climate variability, vegetation changes, fire occurrence and human occupation relationships were inferred at Mallín Pollux and the Chonos Archipelago areas suggesting a regional common pattern at Central Patagonia during the Late Holocene.

During the last 0.3 cal ka BP and particularly during the last century, Lake Shaman pollen record indicates a decrease of *Nothofagus* and Poaceae, whereas *Acaena* and Asteraceae subf. Asteroideae increase markedly (Figs. 5 and 6). Even though both tendencies would have been naturally caused at first, they could be enhanced as a consequence of human activities, like forest clearance and sheep grazing after the European settlement (1902 AD; Martinic, 2005) in the Río Cisnes valley. According to Bertioller and Bisigato (1998), strong grazing pressures at grass steppes produce the replacement of plant species between and within functional
types leading to an increase in shrub cover at expense of the absolute cover of perennial grasses and the consequent changes in the plant community physiognomy. Bertiller and Bisigato (1998) have demonstrated that *F. pallescens* grassland is similar to that surrounding Lake Shaman, where perennial grasses are replaced by *M. spinosum*, *Senecio* spp. (Asteraceae subf. Asteorideae family) or by dwarf shrubs such as *Acaena* in more disturbed sites as it is reflected in the Lake Shaman pollen record (Fig. 5). According to this, the reduction of fire events in the steppe since 0.3 cal ka BP (Fig. 7) could be related to the shift of the vegetation’s structure given that the appearance of shrubs into the grass steppe would have interrupted the fuel continuity leading to less frequent and/or less intense fires.

5.2. Paleoclimatic implications

5.2.1. Interpretation assumptions

The effective moisture changes inferred from Lake Shaman record located at the forest-steppe ecotone along the lee side of the Andes have been mainly related to precipitation changes associated with the Southern Westerlies dynamics. This is based on the fact that the location of the forest-steppe ecotone is dependent on moisture balance (Paruelo et al., 1998) directly related to the amount of precipitation derived from the SW eastward over the Andes. However, some authors have recently questioned the paleoclimatic inferences based on pollen records located on the forest-steppe ecotone arguing that: (1) vegetation changes in this transitional area are strongly controlled by evaporation changes in addition to possible effects of precipitation changes and (2) long-distance pollen transport (extra-local signal) from the westward forests might mask the local vegetation changes (Lamy et al., 2010). Despite this disagreement, modern climate studies demonstrate that a positive correlation exists between the monthly anomalies of precipitation in southern Patagonia and the low-level zonal flow (700–850 hPa) which decreases from west to east (rainshadow effect) but still shows positive values on the lee side of the Andes (Garreaud, 2007; Moy et al., 2008; Moreno et al., 2010a). In fact, correlation between monthly precipitation anomalies and the low-
Central Patagonia during deglaciation has been already evidenced a locally nourished ice cap in the peninsula. The local ice cap would the west and the colder and drier climate on the eastern is suggested that the differential glacial dynamics could be quite well studied (e.g. Mancini, 1998; Haberle and Bennett, 2001; Heusser, 2003; Tonello et al., 2009; Bamonte and Mancini, 2011; Mancini et al., 2012) therefore local and extra-local pollen signals can be clearly identified when interpreting the fossil pollen records.

5.2.2. Lateglacial

The deglaciation for the Patagonian Ice Cap was initiated by a sudden rise in temperature across Patagonia at around 17.5–17.2 cal ka BP. Nevertheless, there is also evidence for out of synchrony for the deglaciation of different glacial lobes in Patagonia, which is believed to be related to changes in accumulation and ablation during decay (McCulloch et al., 2000). As the beginning of sedimentation at Lake Shaman record suggests, upper Rio Cisnes valley would have been already free of ice around 19 cal ka BP. A similar age (18.6 cal ka BP) was assigned to the end of glacial influence southwards at Mallín Pollux (Markgraf et al., 2007), indicating an early deglaciation timing for Lake Shaman and Mallín Pollux areas compared to those ages proposed for the Patagonian Ice Cap deglaciation (McCulloch et al., 2000). Besides, the end of glacial influence ages at Lake Shaman and Mallín Pollux areas match within the Lake Buenos Aires (LBA) LGM moraine chronology (ca 23–16 cal ka BP; Douglass et al., 2006; Kaplan et al., 2004) and the Lago Cochrane/Pueryrudón (20–18 cal ka BP; Hein et al., 2010). All these data provide therefore a consistent scenario for the age of deglaciation at the eastern flank of the Andes in Central Patagonia. Records located in the western side of the Andes at the Chonos Archipelago (Fig. 1a) suggest minimum deglaciation ages between 16.3 and 15.9 cal ka BP (Laguna Oprasa and Laguna Fácil, respectively; Haberle and Bennett, 2004) whereas in the Taitao Peninsula (Fig. 1a) the deglaciation occurred at around 17.2 and 17.5 cal ka BP (Laguna Stibnite and Laguna Six Minutes, respectively) (Bennett et al., 2000). According to Glasser et al. (2008), the differences between the Taitao Peninsula and Mallín Pollux deglaciation timing and the continental areas at either west or east sides of the Andean cordillera (44°–46°S), may be related to the existence of a locally nourished ice cap in the peninsula. The local ice cap would have been completely independent from the LGM Patagonian Ice Cap, which would have actually presented different dynamics (Glasser et al., 2008). The high variability in the glacial dynamics at Central Patagonia during deglaciation has been already evidenced by previous studies (Hulton et al., 2002; Glasser et al., 2008). Both, glacial erosional features (Glasser et al., 2008) as well as results of ice sheet modeling (Hulton et al., 2002) have shown differential glacial dynamics between west and east of the Andean cordillera. It is suggested that the differential glacial dynamics could be reflecting the contrast between the maritime template climate on the west and the colder and drier climate on the eastern flank of the Andes.

The presence of steppe vegetation at Lake Shaman indicates colder and drier conditions than the present until 14.8 cal ka BP. The cold/dry vegetation assemblages show large similarity to that recorded at Mallín Pollux between 18 and 15.4 cal ka BP (Markgraf et al., 2007) and time correlation to the presence of the plant community adapted to cold conditions such us assemblage of Ericaeae, grasses, sedges, and cool-wet herbaceous plants in the

Chonos Archipelago and the Taitao Peninsula until 14.8 (Haberle and Bennett, 2004). The colder than present conditions at around 14.8 cal ka BP is also evidenced by the chironomid assemblages of Laguna Fácil in the Chonos Archipelago (Massaferro and Brooks, 2002). Glacial geomorphology studies performed along the eastern flank of the Patagonian Andes indicate persisting glacial activity previous to 14.5 cal ka BP supporting the idea of cold conditions during the early deglaciation phase (e.g. Sugden et al., 2005; Douglass et al., 2006; Moreno et al., 2009).

A shift in the vegetation assemblages over Central Patagonia was recorded at both east and west of the Andes after 14.8 cal ka BP. Lake Shaman and Mallín Pollux (Markgraf et al., 2007) indicate steppe-dominated vegetation with scattered trees (Nothofagus) whereas rainforest taxa including Nothofagus, Pilgerodendron and Podocarpus developed in the Chonos Archipelago and Taitao Peninsula (Haberle and Bennett, 2004). All these pollen records located in Central Chilean Patagonia suggest an increase in effective moisture and a rise in temperature at around 14.8 (cal ka BP according to the establishment of similar-than-present rainforests in the Chonos Archipelago and the Taitao Peninsula (Fig. 8). Simultaneously, NordPatagonic forests dominated North Patagonia (39°–42°S) vegetation until 14.2 cal ka BP when they were replaced by Valdivian forests (Lago Mellí; Figs. 1a and 8) suggesting the shift from cold-wet LGM conditions to warmer and drier than LGM conditions (Moreno and León, 2003; Abarzúa et al., 2004; Moreno, 2004). On the other hand, at southern Patagonia, Lago Guanaco pollen record (51°S; Figs. 1a and 8) shows the dominance of steppe grasses and forbs from 13 cal ka BP (Paleovegetation Index below zero) but interrupted by the arboreal expansion reaching to a maximum until around 11.3 cal ka BP (Paleovegetation Index close to zero).

Taken together, the paleoecological and glaciological data indicates widespread colder conditions along Patagonia during the glacial-interglacial transition but wetter (drier) than present at North (Central and South) Chilean Patagonia. The evidence indicates a northern position of the Southern Westerlies compared to present-day. A displacement towards the equator by the Southern Westerlies at around 41°S during part of the LGM (21 cal ka BP) has been supported by model simulations (Rojas et al., 2009). Later on, the SW would have experimented a subsequent shift southwards of their present position around 14.3 cal ka BP, only 1.6 cal ka BP after the first warming pulse that would have initiated deglaciation (17.5–17.2 cal ka BP) along Patagonia (McCulloch et al., 2000).

5.2.3. Early Holocene

Overall changes in the vegetation occurred throughout Patagonia during the early Holocene (11.5–8 cal ka BP). Pollen records from Lake Shaman and Mallín Pollux (Markgraf et al., 2007) in eastern Chilean Patagonia exhibit similar vegetation changes at multi-millennial timescale. The beginning of forest-steppe ecotone development (Lake Shaman) and the shift from a steppe to a steppe-woodland vegetation (Mallín Pollux; Markgraf et al., 2007) indicate steady to slightly increased effective moisture between 11.5 and 8 cal ka BP. Nevertheless, the sudden and simultaneous dominance of paludal taxa and the increased fire frequency at both sites would suggest increased summer temperatures at around 11 cal ka BP. In addition, laminated carbonates in the sedimentological records of Lago Augusta (Fig. 1a) (Villa-Martínez et al., 2012) were interpreted as evaporation rates stronger-than-present related to warmer and drier summers than the present. The replacement of the NordPatagonic forest (decline of cold-tolerant taxa) by the Valdivian forest (appearance of thermophilous taxa) in North Chilean Patagonia (Lago Mellí, Fig. 8) (Moreno and León, 2003; Abarzúa et al., 2004; Moreno, 2004) and in the Chonos Archipelago and the Taitao Peninsula (Haberle and
Fig. 8. Integration of a. Lake Shaman Paleovegetation Index (PI, Nothofagus/steppe taxa + Poaceae), fire episode frequency and CHAR with b. summary of palaeofire activity in western South America (>30°S; Power et al., 2008); c. Mallín Pollux PI (Nothofagus/steppe taxa + Poaceae; Markgraf et al., 2007); d. Chonos Archipelago relative precipitation and temperature changes based on inferences of Haberle and Bennett (2004); e. Lago Melli Paleoforest Index (Eucryphia + Cladcluvia/Podocarps) and Valdivian and NordPatagonic forest percentages (Abarzúa, 2004) and f. Lago Guanaco NPI (Nothofagus/Poaceae; Moreno et al., 2010a).
Bennett, 2004) suggest warmer and drier conditions than during the Late Glacial and the present. A major increase in grasses and forbs indicates a retraction of the forest-steppe ecotone at Southern Chilean Patagonia (Fig. 8) (Lago Guanaco; Moreno et al., 2010a) suggesting a decline in effective moisture. In addition, charcoal data from almost all Patagonian sites south of 40°S (Whitlock et al., 2007; Abarzúa and Moreno, 2008; Power et al., 2008; Moreno et al., 2010b) showed a shift in fire regime associated with a widespread fire activity (positive anomalies; Fig. 8b) which correlates to changes in the vegetation between 11 and 8.5 cal ka.

In combination, vegetation and charcoal paleoecological records throughout Patagonia indicate drier and warmer conditions during Late Glacial — early Holocene (11.5–8 cal ka BP). The implications for the paleoecological data suggest weaker and poleward shift of the SW over the continent (Whitlock et al., 2007; Moreno et al., 2010b) which could be due to a reduction in latitudinal temperature gradients driven by higher-than-present annual insolation in midlatitudes (40°S) leading to warm dry conditions at midlatitudes (40–50°S) of western Patagonia whereas during a SAM positive phase, the SW deflect to 50–60°S resulting in dry conditions between 35–43°S and positive temperature anomalies along midlatitude eastern Andes. Hence, both the ENSO and SAM may have triggered climatic variability at Central Chilean Patagonia during the Late Holocene that, associated with a broad human occupation (ignition agent), would have resulted in high fire occurrence and the associated vegetation changes.

Vegetation changes have been strongly associated with a broad human occupation and the consequent increased anthropogenic activity at Central Patagonia from mid 19th until early 20th century (Cárdenas et al., 1993; Martinic, 2005). Pollen and charcoal records reflect the European settlement as a decline of arboreal taxa in forest environments associated with peaks of charcoal in some cases resulting in patchy landscapes due to forest clearance through burning (Szeicz et al., 2003; Haberle and Bennett, 2004; Markgraf et al., 2007). Moreover, the introduction of European exotic taxa such as R. acetosella and Taraxacum officinale is recorded particularly in those sites located within the deciduous forest (Markgraf et al., 2007) whereas an increase in shrubs cover at the expense of the perennial grasses was recorded at forest-steppe ecotone under strong (sheep) grazing pressures.

5.2.4. Mid and late-Holocene

The establishment of a trend to similar-to-modern vegetation and negative charcoal anomalies throughout Patagonia from ~8–7 to 3.5 cal ka BP (Fig. 8) was recorded. Central Chilean Patagonian records indicate the presence of the Nothofagus forests at Mallín Pollux and Lago Augusta, and the North Patagonic rainforest in the Taitao Peninsula and Chonos Archipelago, within this period. No major vegetation changes attributable to the establishment of seasonally equable conditions or low magnitude precipitation changes are reflected in any of these records (Haberle and Bennett, 2004; Markgraf et al., 2007; Villa-Martínez et al., 2012). On the other hand, the forest-steppe ecotone sensitive precipitation record of Lake Shaman point to wetter conditions than present under an unmarked seasonality up to 5 ka BP. At Northern Chilean Patagonia, Lago Melli record shows a trend towards colder and wetter conditions up to 3 ka BP (Abarzúa, 2004) whereas a multi-millenial trend to increased precipitation superimposed to lower magnitude precipitation changes are inferred at Southern Chilean Patagonia from Lago Guanaco record (Moreno et al., 2010a). All these data throughout Patagonia suggest colder and/or wetter conditions than in the early Holocene. According to paleoclimate modeling at 6 cal ka BP, these would respond to intensified SW north of ~50°S as a consequence of a small equatorward shift (1–3°) of the latitude of maximum wind speed over the entire Southern Hemisphere (Rojas and Moreno, 2009).

Around 3 cal ka BP, charcoal records from Central Chilean Patagonia (L. Fácil and L. Opra, Haberle and Bennett, 2004; L. Shaman) with exception of M. Pollux record (Markgraf et al., 2007) show higher fire activity than present associated with a high variability in the pollen record. Regional paleofire reconstructions (Whitlock et al., 2007; Moreno et al., 2010b) noted a spatially heterogeneous increase of fire activity south of 30°S that may have resulted from a breakdown of regional climate controls. This increase of fire spatially heterogeneous could be related to the greater interannual and interdecadal climate variability as well as increased use of fire by natives together with their widespread occupation during the Late Holocene. Climate variability may be associated with synoptic patterns related to the El Niño Southern Oscillation (ENSO) (actually more recurrent during the last 16,000 yr). Pollen and charcoal records during the mid and late Holocene that, associated with past SW major changes (latitudinal shifts and/or strengthening/weakening) along Patagonia. Late Holocene record variability, on the other hand, appears as a regional common pattern throughout Central Chilean Patagonia related probably to low magnitude SW changes associated with ENSO and/or SAM (or their interactions) or to the complex relationships between vegetation, fire and human occupations during this time.

6. Conclusive remarks

The record from Lake Shaman is the longest record that provides evidence of environmental and climatic dynamics since the beginning of deglaciation at Central Chilean Patagonia. Lake Shaman is proved to be one of the most sensitive sites concerning the Holocene climate variability in this area. Major vegetation and climate changes were recorded from the Late Glacial to the early Holocene, from a grass-shrub steppe to a forest-steppe ecotone similar to the present around 11.5 cal ka BP. Low magnitude climatic changes were reflected in the Lake Shaman pollen and charcoal records during the mid and late Holocene, contrary to other less-sensitive records located at the rainforest or deciduous Nothofagus forest at Central Chilean Patagonia. A maximum development of the forest in the mid-Holocene indicating the easternmost position of the forest-steppe ecotone since the Late Glacial was inferred from Lake Shaman's record whereas high pollen assemblage variability characterized the late-Holocene.

All these changes integrated at the regional (Central Chilean Patagonia) and extra-regional (Chilean Patagonia) scale and associated with past climate model simulations indicate that record changes from the Late Glacial to the mid-Holocene were mostly related to past SW major changes (latitudinal shifts and/or strengthening/weakening) along Patagonia. Late Holocene record variability, on the other hand, appears as a regional common pattern throughout Central Chilean Patagonia related probably to low magnitude SW changes associated with ENSO and/or SAM (or their interactions) or to the complex relationships between vegetation, fire and human occupations during this time.

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