

## THE LAST GLACIAL MAXIMUM AND DEGLACIATION IN CENTRAL PATAGONIA, 44° S-49° S

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**ABSTRACT.** *This paper reviews published geochronological data on glacier fluctuations and environmental changes in central Patagonia (44° S-49° S) from the Last Glacial Maximum (LGM) through to the Holocene. Well-dated glacial chronologies from the southern mid-latitudes can inform on the synchronicity of glacial advances worldwide and provide insight on the drivers of southern hemisphere glaciations. In central Patagonia, two large outlet lobes of the former Patagonian Ice Sheet advanced in broad synchrony with the global LGM. In contrast to other parts of Patagonia, there is no convincing evidence for a more extensive local LGM advance during Marine Isotope Stage 3. Deglaciation initiated at ca. 19 ka, earlier than in other parts of Patagonia and regionally in the Southern Hemisphere, and rapid deglaciation saw ice margins retreat in places by at least 80-120 km within a few millennia. The Lateglacial glacier margins are poorly constrained, but an ice mass substantial enough to maintain a large regional proglacial lake must have persisted at this time. The timing of lake drainage and opening of the Río Baker drainage route to the Pacific Ocean is debated; the only directly dated shoreline suggests this occurred at the end of the Antarctic Cold Reversal at 12.7 ka. Palaeoecological evidence for cooling during the Antarctic Cold Reversal or Younger Dryas remains equivocal, which may reflect both the eurythermic nature of Patagonian vegetation and shifting Southern Westerly Winds. Eastern outlet glaciers appear to have advanced or stabilised at the Lateglacial/Holocene transition when palaeoenvironmental records indicate warmer and drier conditions, but the reason for this is unclear. Our review reveals both spatial and temporal gaps in available data that provide avenues for future research.*

***El Último Máximo Glaciar y la deglaciación en Patagonia central, 44° S-49° S***

**RESUMEN.** Este artículo revisa los datos geocronológicos publicados sobre las fluctuaciones de los glaciares y los cambios ambientales en la Patagonia central (44° S-49° S) desde el Último Máximo Glaciar (LGM) hasta el Holoceno. Las cronologías glaciales bien datadas de las latitudes medias del sur pueden informar sobre la sincronidad de los avances glaciares en todo el mundo y proporcionar una visión de los factores de las glaciaciones del hemisferio sur. En el centro de Patagonia, dos grandes lóbulos de salida del antiguo Campo de Hielo Patagónico avanzaron en una amplia sincronía con el LGM global. En contraste con otras partes de Patagonia, no hay evidencia convincente de un avance local más extenso que la que tuvo lugar durante el LGM durante la Etapa 3 de Isótopos Marinos. La deglaciación se inició hacia aprox. 19 ka, antes que en otras partes de la Patagonia y regionalmente en el hemisferio sur, y la rápida deglaciación vio los márgenes de hielo retroceder en algunos lugares por lo menos 80-120 km en pocos milenios. Los márgenes de los glaciares del Tardiglaciar están mal delimitados, aunque una masa de hielo sustancialmente suficiente para mantener un gran lago proglaciar regional tuvo que persistir en ese momento. Se discute cuándo ocurrió el drenaje del lago y la apertura de la ruta de drenaje de Río Baker hacia el Océano Pacífico. La única costa directamente datada sugiere que esto ocurrió al final de la Reversión Fría Antártica en 12.7 ka. La evidencia paleoecológica del enfriamiento durante la Reversión Fría Antártica o Younger Dryas sigue siendo equívoca, lo que puede reflejar tanto la naturaleza eutérmica de la vegetación patagónica como los vientos del oeste meridionales cambiantes. Los glaciares de salida oriental parecen haber avanzado o estabilizarse en la transición Tardiglaciar/Holoceno cuando los registros palaeoambientales indican condiciones más cálidas y más secas, pero la razón de esto no está clara. Nuestra revisión revela vacíos espaciales y temporales en los datos disponibles que proporcionan vías para la investigación futura.

**Keywords:** Last Glacial Maximum, deglaciation, Holocene, Patagonia, glacial chronology.

**Palabras clave:** Último Máximo Glaciar, deglaciación, Holoceno, Patagonia, cronología glacial.

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## 1. Introduction

The former Patagonian Ice Sheet (PIS) extended 2000 km along the southernmost Andes, between 38° S and 55° S during the Last Glacial Maximum (LGM; Fig. 1). The ice sheet left a rich glacial geomorphological record of ice limits (e.g. Glasser *et al.*, 2008) that are particularly well preserved east of the mountain front, and which offer the opportunity to constrain the history of ice fluctuations throughout the last glacial cycle. The location in the mid-latitudes of the southern hemisphere, distant from the influence of northern hemisphere ice sheets, makes the site suitable to investigate, for example, the geographic extent of the global LGM and its termination, the interhemispheric linkages of climate variability and the forcing behind southern mountain glaciation.

There is a long history of research on Quaternary glaciations in Patagonia that began in the 19<sup>th</sup> century leading to the pioneering work of Caldenius (1932), who established the first comprehensive mapping of glacier moraine limits throughout Patagonia. The establishment of glacial chronologies that utilised modern radiometric dating techniques, including radiocarbon (<sup>14</sup>C) and potassium-argon (K-Ar), began in the 1960's through

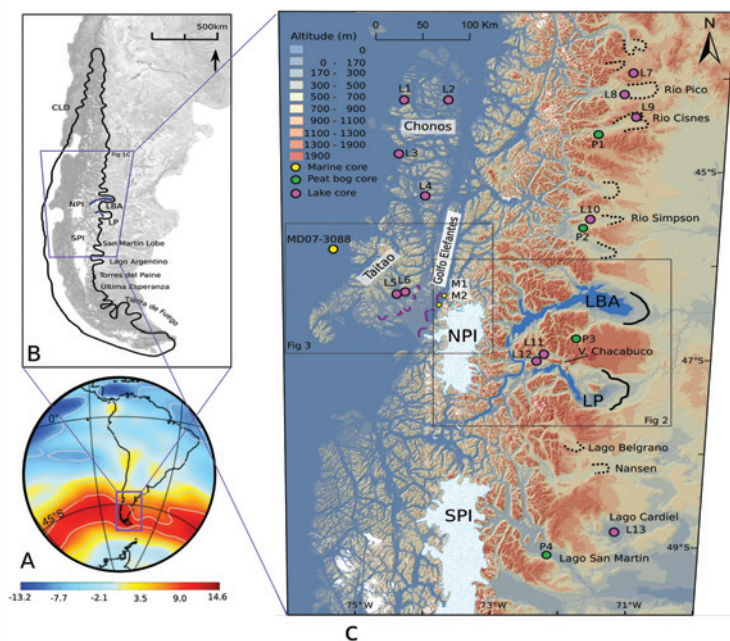


Figure 1. (A) Location of Patagonia, South America, with mean monthly zonal wind strength ( $m s^{-1}$ ) at 850 hPa based on NCEP/NCAR re-analysis data (modified from Vanneste *et al.*, 2015). (B) Estimated extent of the PIS at the LGM showing sites mentioned in the text; CLD- Chilean Lake District. (C) Map of Central Patagonia showing locations of palaeoenvironmental proxies, which are summarised in Table 1 and ice limits: black - chronologically constrained ice limits, black dashed- assumed LGM limits, purple – undated moraines on the Taitao Peninsula discussed in the text and moraines of the Gualas/Reicher, Rafael, and San Quintín Glaciers (Image: SRTM).

the work of John Mercer (e.g. Mercer, 1968, 1976, 1982, 1983). Subsequent efforts to develop the glacial history of the PIS have used a wide range of geochronological techniques and are well summarised by several review papers on Quaternary (McCulloch *et al.*, 2000; Rabassa and Clapperton, 1990; Rabassa and Coronato, 2009; Kaplan *et al.*, 2008a; Glasser *et al.*, 2004; Rodbell *et al.*, 2009; Clapperton, 1993; Harrison and Glasser, 2011; Rabassa *et al.*, 2000) and pre-Quaternary (Rabassa *et al.*, 2011) glaciations.

This paper aims to synthesize existing geochronological data on ice limits and environmental changes during the LGM and subsequent deglaciation, in the central part of the former PIS. We review available dating of glacial moraine sequences as well as palaeoenvironmental proxies; the latter are more geographically dispersed in the region and provide insight into postglacial landscape development and climate. We focus on the area surrounding the present day North Patagonian Icefield (NPI), between latitudes 44° S and 49° S (Fig. 1c), but draw comparisons to other parts of Patagonia. This area has been considered a transitional zone influenced by both Antarctic and Northern Hemisphere climate (Sugden *et al.*, 2005), and is an area where climate models suggest a 'discontinuity' in several climatic components (e.g. wind speed, precipitation) (Rojas *et al.*, 2009). Our review reveals that glacial advances synchronous with the global LGM (gLGM; 26-19 ka; Clark *et al.*, 2009) are constrained at two sites in central Patagonia. Numerous moraines, likely correlative to the local LGM, are found in almost every previously glaciated basin, but have so far received little attention. Tentative evidence suggests pre-LGM advances during the last glacial cycle but further work is needed to constrain these. Deglaciation initiated at about 19 ka, somewhat earlier than in southern Patagonia. Ice marginal positions during the Lateglacial are largely unknown, and the occurrence of cooling coeval with the Antarctic Cold Reversal (ACR; 14.5-12.9 ka) or Younger Dryas (YD; 12.9-11.7 ka) is equivocal.

## 2. Regional Setting

The regional geology of the southern Andes is characterized by intrusions of the Patagonian batholith within the Andean Cordillera, Mesozoic and Cenozoic volcanics, Mesozoic marine sediments, and a Paleozoic metamorphic complex (Vandekerkhove *et al.*, 2016; Lagabrielle *et al.*, 2004). The Taitao Peninsula and Chonos Archipelago are formed of Triassic metamorphic rocks and mid-Cenozoic volcanic and sedimentary deposits, with the Patagonian batholith constraining the eastern edges. Basins to the east of the Andes are filled with Pliocene-Quaternary glacial and glaciofluvial sediments (Vandekerkhove *et al.*, 2016; Lagabrielle *et al.*, 2004). A gap in the active volcanic arc between the Southern Andean Volcanic Zone and Austral Andean Volcanic Zone exists in central Patagonia (46.5° S-49° S), near the Chile Triple Junction, where the Nazca, South American and Antarctic tectonic plates meet (Gutiérrez *et al.*, 2005; Stern, 2004).

The region contains the highest mountain in the Patagonian Andes, Mt. San Valentin at 4058 m, which is situated on the northern flank of the NPI. The otherwise topographically continuous chain of the Andes exhibits a break in central Patagonia between latitudes 47° S and 48° S. This gap is occupied by the Río Baker, which drains the eastern outlet glaciers of the NPI, including the large cross-border lakes of Lago Buenos Aires/Lago

General Carrera (LBA) and Lago Pueyrredón/Lago Cochrane (LP) westwards into the Pacific Ocean (Fig. 1c). Therefore, the water divide is unusually located 170 km to the east of the mountain chain. The landscape east of the Andes is characterized by west-east trending over-deepened valleys separated by volcanic plateaus. To the west of the Andes, the landscape is dominated by an extended fjord and island system. The ice flow during successive glaciations exploited pre-existing structural faults, and over-deepened and widened the fjords and valleys through selective erosion (Glasser and Ghiglione, 2009).

The Patagonian Andes of southern South America are situated in the mid-latitudes of the Southern Hemisphere, where the landmass intersects important global atmospheric and oceanic systems. These include the Southern Hemisphere Westerly Winds (SWWs) (Fig. 1a), which play a role in modulating global climate through their influence on ocean currents and ventilation of the deep Southern Ocean (Toggweiler *et al.*, 2006; Garreaud *et al.*, 2013). The SWWs bring precipitation that feeds the largest temperate ice fields in the Southern Hemisphere, the NPI and the South Patagonian Icefield (SPI) (Warren and Sugden, 1993). In addition to these icefields, two smaller icefields, the Cordillera Darwin and Gran Campo Nevado, along with numerous smaller glaciers are all that remain of the former PIS. The present day NPI covers an area of 3976 km<sup>2</sup> (Davies and Glasser, 2012), contains 1234.6 km<sup>3</sup> of ice (3.10 mm sea-level equivalent) (Carrivick *et al.*, 2016) and has a mean ice thickness of 305 m. The equilibrium line altitude ranges from ~900 m in the west to ~1300 m in the east (Rivera *et al.*, 2007).

The strong circulation at the core of the SWWs (49° S-53° S) is due to limited land at these latitudes, resulting in low friction and a strong pressure gradient between the cold Antarctic and warm subtropics (Aravena and Luckman, 2009; Garreaud *et al.*, 2009). The Patagonian Andes act as the only significant barrier to the SWWs, causing a strong orographic effect, as a result forming an extreme climate divide. Precipitation in the west exceeds 3000 mm yr<sup>-1</sup>, reaching 6000 mm yr<sup>-1</sup> (in places up to 10,000 mm yr<sup>-1</sup>) on the mountain tops, and dropping to less than 100-200 mm yr<sup>-1</sup> within 100 km east of the Andean foothills (Garreaud, 2009). The precipitation gradient creates distinct vegetation zones. In central Patagonia, the North Patagonian rainforest dominates humid western islands and mountain slopes, along with moorland covering some islands. On the leeward mountain side, subantarctic deciduous forest transitions through shrubland to the Patagonian steppe in the semi-arid zone. Subantarctic rainforest fills the gap between the NPI and SPI.

### **3. Methods**

For the purpose of this review we have re-calculated all geochronological data from central Patagonia. Radiocarbon ages were recalculated using the OxCal 4.3 online calibration tool (Bronk Ramsey, 2009) and the ShCal13 Southern Hemisphere calibration curve (Hogg *et al.*, 2013), and are presented in the text as median cal ka BP. For cosmogenic <sup>10</sup>Be surface exposure ages, we used the online exposure age calculator formerly known as the CRONUS-Earth online calculator (v. 2.2; constants v.2.2.1; Balco *et al.*, 2008) and the local <sup>10</sup>Be production rate for southern Patagonia (Kaplan *et al.*, 2011). Cosmogenic <sup>36</sup>Cl ages have been re-calculated using the CRONUScalc web calculator (Marrero *et al.*, 2016a) and production rates presented by Marrero *et al.* (2016b). Ages for moraines

Table 1. Palaeoenvironmental proxies discussed in the text.

Site ID*	Core	Site	Lat (DD)	Long (DD)	Elev. (m)	Core length (m)	Stratigraphic range (ka)	Age control	References
<b>Lake cores</b>									
L9	LS0604A	Lake Shaman	46.43	71.18	919	6.13	19 - 0	14C	de Porras <i>et al.</i> , 2012
L7	composite	Lago Los Ninos	44.01	71.29	1015	8.12	17 - 0.1	14C	Iglesias <i>et al.</i> , 2016
L8	LP09B	Laguna La Pava	44.28	71.31	731	4.8	13.2 - 0.5	14C	Iglesias <i>et al.</i> , 2016
L10	CAST01	Lago Castrol	45.6	71.8	699	15.4	41 - 0	14C, floating varve, tephra	Van Daele <i>et al.</i> , 2016
L11	composite	Lago Augusta	47	72.38	440	11.53	16 - 0.65	14C, tephra	Villa-Martinez <i>et al.</i> , 2012
L12	composite	Lago Editá	47.13	72.42	570	17.26	19 - 9.0	14C, tephra	Henriquez <i>et al.</i> (in review)
L1	-	Laguna Facil	44.33	74.28	10	7.63	15.8 - 0	14C	Haberle and Bennet, 2004
L2	-	Laguna Oprasa	44.35	73.66	50	7.26	16.3 - 0	14C	Haberle and Bennet, 2004
L4	-	Laguna Lincoln	45.35	74	< 50	-	-	14C	Bennett <i>et al.</i> , 2000
L3	-	Laguna Lofel	44.88	74.41	< 50	-	-	14C	Bennett <i>et al.</i> , 2000
L6	-	Laguna Six Minutes	46.42	74.34	< 50	-	-	14C	Bennett <i>et al.</i> , 2000
L5	-	Laguna Stibnite	46.45	74.42	< 50	-	-	14C	Bennett <i>et al.</i> , 2000
L5	-	Laguna Stibnite	-	-	20	4.75	17.4 - 0	14C	Lumley and Switsur, 1993
<b>Peat cores</b>									
P1	composite	Mallín El Embudo	44.66	71.7	686	8.44	13 - 0	14C	De Porras <i>et al.</i> , 2014
P2	-	Mallín Pollux	45.69	71.84	640	14.07	18 - 0	14C	Markgraf <i>et al.</i> , 2007
P3	-	La Frontera	46.87	71.88	997	3	8.2 - 0.35	14C, tephra	McCulloch <i>et al.</i> , 2016
P4	-	La Tercera	49.18	72.36	-	2.19	11.3 - 0	14C	Bamonte and Mancini 2011
<b>Marine cores</b>									
MD07-3088	MD07-3088	-	-	-	-	18.9	22 - 0	δ18O, 14C, tephra	Siani <i>et al.</i> , 2010; Montade <i>et al.</i> , 2013
M1	JPC14	Golfo Elefantes	-	-	-	-	~ 11.3 - 0	14C	Fernandez <i>et al.</i> , 2012
M2	JPC12	Golfo Elefantes	-	-	-	-	~ 11.3 - 0	14C	Fernandez <i>et al.</i> , 2012

\*refers to site ID in Fig. 1

based on multiple cosmogenic nuclide analyses were derived from camel plots and are presented as the probabilistic peaks. The outliers identified in the original publications were removed from the camel plots. All exposure ages assume zero rock surface erosion and no shielding (snow, soil, loess, vegetation) and thus are minimum ages. However, inclusion of an erosion rate of  $0.2 \text{ m Ma}^{-1}$  derived in the LBA valley on the basis of paired  $^{36}\text{Cl}/^{10}\text{Be}$  concentrations (Douglass *et al.*, 2007), increases an LGM age by less than 1%, and thus is well within the reported internal uncertainties. Individual exposure ages are presented with  $1\sigma$  external uncertainties to allow comparison to radiocarbon ages. The data used to calculate the exposure ages are presented in Supplementary Table 1 (Available at: [http://publicaciones.unirioja.es/revistas/cig2017/Mendelova\\_TableS1.xlsx](http://publicaciones.unirioja.es/revistas/cig2017/Mendelova_TableS1.xlsx)). The original publications should be consulted for the original data and further information.

#### **4. Glacial History**

Mapping and dating of glacial limits has been done primarily along the eastern flank of the Patagonian Andes. The geomorphological record here is exceptionally well preserved due in part to the arid climate on the eastern side of Andes. Here, moraine systems can be found in nearly every valley between  $38^\circ \text{ S}$  and  $56^\circ \text{ S}$ , and often are associated with, or confine large lakes. In the past decade the application of cosmogenic nuclide surface exposure dating and optically stimulated luminescence (OSL) has led to an increase in the concentration and geographic range of glacial chronologies. Numerous well-dated chronologies have been developed in southern Patagonia and Tierra del Fuego (e.g. Darvill *et al.*, 2015; Kaplan, *et al.*, 2008b; McCulloch *et al.*, 2005b). With the exception of two well-dated chronologies from LBA and LP, glacial chronologies remain sparse in central Patagonia, and at the time of writing there are no published chronologies for NE Patagonia. To the west of the Andes, the PIS terminated in the ocean during the LGM and most geomorphological evidence is therefore submarine, with the exception of north-western Patagonia including the Chilean Lake District, where glaciers terminated on land and their advances over the last glacial cycle have been well documented (Denton *et al.*, 1999).

Regional distribution of palaeoenvironmental records largely depends on the availability of suitable natural archives (e.g. peat bogs, lakes). However, the location of study sites is also biased towards the east in part due to the limited accessibility of the Chilean fjords, with the exception of Isle Grande Chiloé in north-western Patagonia, and the Chonos Archipelago and Taitao Peninsula in central-western Patagonia. Swath bathymetry can reveal submarine geomorphological evidence and it has been applied in the southern Chilean fjords (e.g. Dowdeswell *et al.*, 2016; Dowdeswell and Vásquez, 2013). Less is known about the bathymetry of the fjords between  $44^\circ \text{ S}$  and  $46^\circ \text{ S}$  (Völker *et al.*, 2013).

Between latitudes  $44^\circ \text{ S}$  and  $49^\circ \text{ S}$ , the former PIS drained eastwards by ten topographically controlled outlet glaciers and ice lobes, which extended onto the Andean plains during the LGM (Glasser and Jansson, 2005; Glasser *et al.*, 2008). Each of these outlet glaciers left behind a series of moraine limits (Fig. 2). To date, however, most research efforts have focused on just two of these formerly glaciated basins: LBA and LP. Here, argon-argon ( $^{40}\text{Ar}/^{39}\text{Ar}$ ) dating of basaltic lava flows interbedded with the glacial deposits,

cosmogenic nuclide surface exposure dating, soil formation rates and OSL dating have been used to date glacial advances spanning from the Greatest Patagonian Glaciation (ca. 1.1 Ma) through to the last glacial cycle (e.g. Singer *et al.*, 2004; Douglass and Bockheim, 2006; Kaplan *et al.*, 2004; Smedley *et al.*, 2016; Hein *et al.*, 2010). Moraine systems in the other basins, which probably relate to the last glacial cycle, or older, have received little attention. Younger, likely Lateglacial to Holocene moraines are found within the Andean valleys closer to the NPI and present day mountain glaciers. On the western side of the Andes, terrestrial geomorphological evidence is confined to the Taitao Peninsula (Fig. 1c).

The locations of environmental proxies discussed in the paper are shown in Figure 1c and are listed in Table 1. On the western side of the Andes, several peat bog and lake cores have been obtained from the Taitao Peninsula and the Chonos Archipelago, which cover the Lateglacial and Holocene period. A marine core (MD07-3088) recovered off the Taitao Peninsula (46° S) provides a high-resolution pollen, clay mineralogy and major geochemistry record covering the last 22 ka (Siani *et al.*, 2010; Montade *et al.*, 2013). Multiple peat bog and lake cores have been sampled from valleys along the eastern flank of the Andes and provide a record covering the Lateglacial and Holocene. Minimum radiocarbon ages for deglaciation are listed in Table 2. In the following, we describe the data that constrain the LGM and the onset of deglaciation, the Lateglacial and the Holocene from the western and eastern side of the Andes.

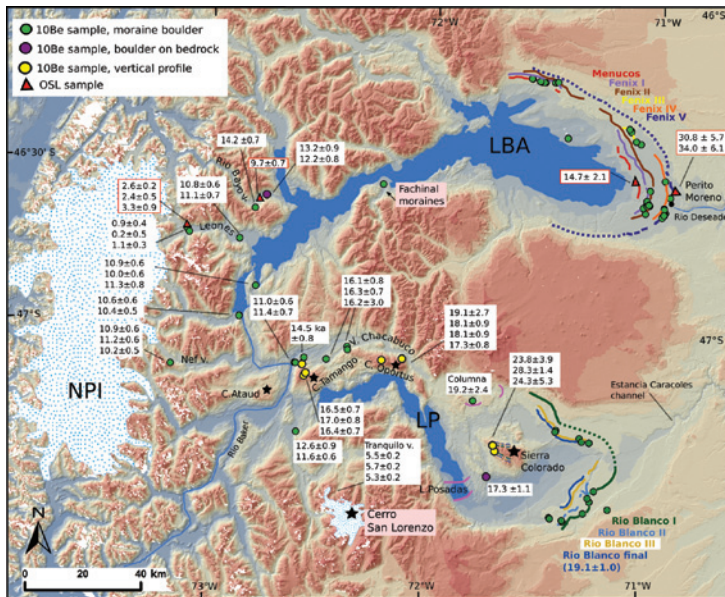


Figure 2. North Patagonian Ice Field (NPI), Lago Buenos Aires/Lago General Carrera (LBA) and Lago Pueyrredón/Lago Cochrane (LP) showing geochronological data (in ka) and delineated moraine limits, Fenix and Río Blanco, corresponding to the global Last Glacial Maximum, Columna and Lago Posadas moraines (pink), and lateral moraines on Sierra Colorado. Camel plots showing age distribution for Fenix, Río Blanco limits, and Fachinal moraines are presented in Figs. 4 - 6, respectively.



Table 2. Minimum radiocarbon ages for deglaciation.

References	Site	Site ID*	Lat (DD)	Long (DD)	Elev. (m)	14C Age	cal BP (median ± σ)	cal range BP (99.7 %)	Archive	Material	Notes
de Porras <i>et al.</i> , 2012	Lake Shaman	L9	46.43	71.18	919	15740 ± 50	18934 ± 72	19187 - 18745	wetland core	macrofossils	
Van Daele <i>et al.</i> , 2016	Lago Castrol	L10	45.6	71.8	699	36 600 ± 250	41204 ± 272	41905 - 40332	lake core	bulk	below hiatus
Villa-Martinez <i>et al.</i> , 2012	Lago Augusta	L11	47.01	72.38	440	16445 ± 45 12 785 ± 35	19792 ± 105 15183 ± 75	20070 - 19530 15452 - 14862	lake core lake core	bulk bulk	outlier
Henríquez <i>et al.</i> (in review)	Lago Edita	L12	47.13	72.41	570	16250 ± 90	19565 ± 142	20001 - 19172	lake core	bulk	
Turner <i>et al.</i> , 2005	Cerro Ataud	-	47.28	72.65	-	16020 ± 50 13550 ± 95	19282 ± 106 16267 ± 157	19572 - 18989 16795 - 15820	lake core kettle hole	bulk peat	
Haberle and Bennet, 2004	Cerro Ataud	-	47.28	72.65	-	13107 ± 81	15651 ± 158	16069 - 15231	kettle hole	macrofossils	
	Laguna Facil	L1	44.32	74.28	10	13230 ± 140	15832 ± 221	16453 - 15187	lake core	bulk	
	Laguna Oprasa	L2	44.35	73.65	50	13560 ± 125	16287 ± 197	16928 - 15751	lake core	bulk	
Lumley and Switsur, 1993	Laguna Stibnite	L5	46.44	74.41	20	14335 ± 145	17401 ± 209	17980 - 16731	lake core	bulk	

\*Site ID refers to locations in Fig.1.

\*\*Radiocarbon ages were recalculated using the OxCal 4.3 online tool (Bronk Ramsey, 2009) and the ShCal13 curve (Hogg *et al.*, 2013).

#### 4.1. Last Glacial Maximum

##### 4.1.1. Western Andes

At the LGM, the PIS is thought to have extended over the western islands and terminated in the ocean near the continental shelf edge (Hubbard *et al.*, 2005). As a consequence, there is no direct field evidence on the extent and timing of glacier advances during the LGM. However, a terrigenous sediment record recovered from immediately offshore of the Taitao Peninsula (Fig. 1c) was used to infer glacial advances at 21.6-20.9 ka, 20.2-20 ka, 19.6-19.1 ka and 18.5-18 ka, based on rapid short-term fluctuations of smectite/ (illite + chlorite) and titanium/potassium (Siani *et al.*, 2010). The record indicates increased erosion of the Coastal Range of Chile due to glacier dynamics until ca. 18 ka. An increase in the sediment input from the Andean Cordillera after 18 ka suggests the withdrawal of the ice margin from the western island zone by this time (Siani *et al.*, 2010). Minimum radiocarbon ages from Laguna Stibnite suggest that the Taitao Peninsula was ice free by  $17.4 \pm 0.2$  cal ka BP (Lumley and Switsur, 1993), and the northern parts of the Chonos Archipelago by  $16.3 \pm 0.2$  cal ka BP (Haberle and Bennett, 2004).

On the Taitao Peninsula, a series of arcuate terminal moraines indicate that valley glaciers flowing east and south from the centre of the peninsula drained an independent ice cap (Fig. 3) (Heusser, 2002; Glasser *et al.*, 2008). In one locality (Laguna Elena), at least twelve distinct moraines indicate multiple re-advances or still-stands (Heusser, 2002). The age of these moraines is unknown, but on the basis of a minimum radiocarbon age for deglaciation of the peninsula ( $17.4 \pm 0.2$  cal ka BP) (Lumley and Switsur, 1993),

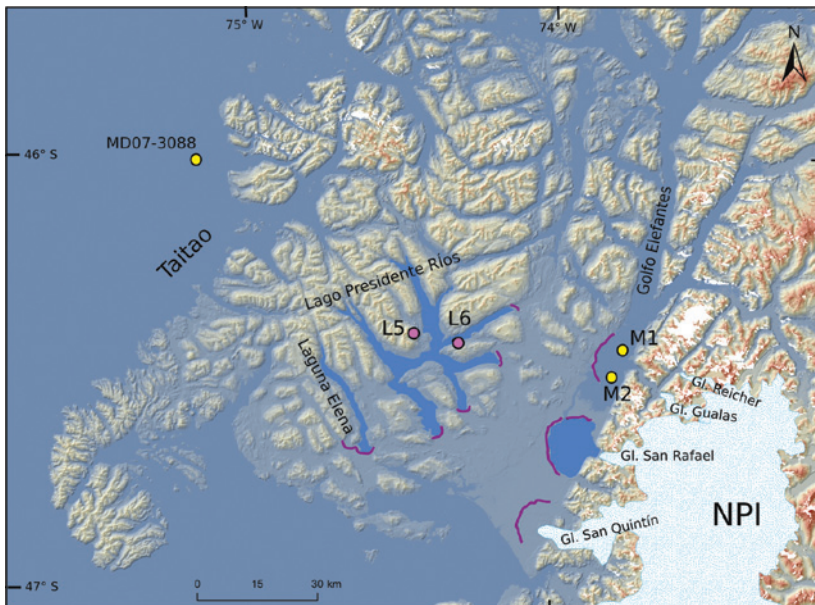


Figure 3. The Taitao Peninsula and adjacent Cordillera showing the locations of undated moraines (purple) discussed in the text, and offshore and lake sediment cores.

it has been suggested that they might date to the LGM (Heusser, 2002). If correct, this would imply either the limited expansion of the PIS to the west at the LGM, or that the deep Golfo Elefantes prevented the ice sheet from overrunning the Taitao Peninsula (Glasser *et al.*, 2008). In the former case, the climatological conditions that would allow for growth of an ice cap on the low-lying peninsula, but not support ice expansion from the Cordillera are unclear (Glasser *et al.*, 2008), and the scenario is difficult to reconcile with LGM ice expansion further north in the Chilean Lake District (Denton *et al.*, 1999). The latter case also seems improbable given that there is morainic evidence to show that the NPI outlet glaciers (Gualas/Reicher Glaciers) advanced over Golfo Elefantes (Heusser, 2002). An alternative is that the moraines on the Taitao Peninsula were deposited during late LGM still-stands, shortly after the separation of the Taitao ice cap from the PIS, and prior to rapid deglaciation on the peninsula.

#### *4.1.2. Eastern Andes*

Ten outlet glaciers drained the former PIS between 44° S and 49° S (Glasser *et al.*, 2008). Several moraine systems can be found at the eastern end of these formerly glaciated basins. While predominantly undated, the innermost moraines have been tentatively correlated to the LGM (Fig. 1c) (Glasser *et al.*, 2008). Rabassa *et al.* (2011) provided a summary of earlier studies that described the drifts and moraines in these basins. In the absence of direct dating of the moraines, radiocarbon ages from lake and peat bog cores provide a minimum deglaciation age (Table 2). In the Río Cisnes valley (44.7° S; Fig. 1), a minimum radiocarbon age from an intermorainic lake (Lago Shaman) indicates that ice abandoned the innermost moraine system before  $18.9 \pm 0.1$  cal ka BP (de Porras *et al.*, 2012). Further upstream, a core from Mallín El Embudo, suggests that the middle part of the Río Cisnes valley was occupied by a proglacial lake until ca. 13 cal ka BP (de Porras *et al.*, 2014). Further south in the Río Simpson basin (45.6° S), a minimum radiocarbon age of  $41.2 \pm 0.3$  cal ka BP was obtained from Lago Castor. However, a hiatus in the stratigraphy suggests that a later readvance over the site might have occurred (Van Daele *et al.*, 2016).

Cosmogenic  $^{10}\text{Be}$  ages from moraine boulders revealed that the LBA outlet lobe (46.5° S) reached its maximum extent sometime between 27-25 ka as marked by the Fenix V moraine (Figs. 2, 4). Four subsequent readvances or still-stands are indicated by the Fenix IV-I moraines, with the final advance at ca. 19.3 ka ( $^{10}\text{Be}$  peak age, Fenix I) (Douglass *et al.*, 2006; Kaplan *et al.*, 2004). Cosmogenic  $^{10}\text{Be}$  ages from moraine boulders on the Río Blanco moraine system showed that the LP ice lobe (47.5° S) reached its maximum extent at ca. 28.0 ka ( $^{10}\text{Be}$  peak, Río Blanco I), a little earlier than the LBA outlet (Hein *et al.*, 2010; Hein *et al.*, 2009). Three subsequent re-advances or still-stands are evident by the Río Blanco II-III moraines, with a final advance dated at ca. 21.6 ka ( $^{10}\text{Be}$  peak; Fig. 5) (Hein *et al.*, 2009, 2010).  $^{10}\text{Be}$  ages from boulders and bedrock along a mountain transect separating the LP and Chacabuco valleys were used to reconstruct ice surface profiles (Boex *et al.*, 2013). Lateral moraines on Sierra Colorado (~1170 m a.s.l.) (Fig. 2) yielded  $^{10}\text{Be}$  ages of  $23.8 \pm 3.9$ ,  $28.3 \pm 1.4$  and  $24.3 \pm 5.3$  ka, thus corresponding to the Río Blanco terminal limits as indicated by

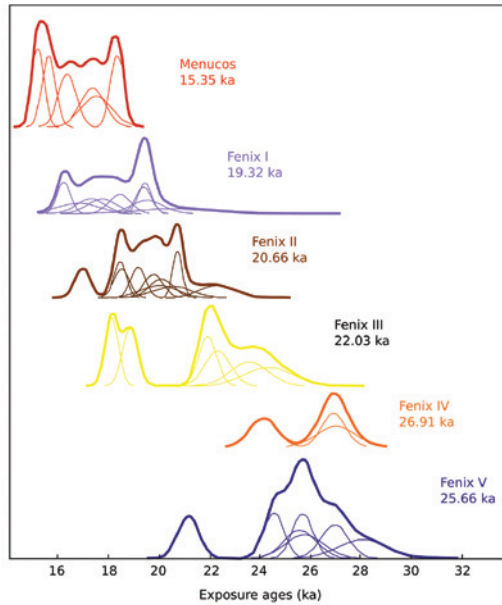


Figure 4. Camel plots summarising distribution of  $^{10}\text{Be}$  ages for each of the Fenix moraine limits and Menucos limit, outliers identified in the original publications were removed. Thin lines represent individual Gaussian probability distributions for  $^{10}\text{Be}$  ages with  $1\sigma$  internal error; thick line represents the sum of the individual probability distributions.

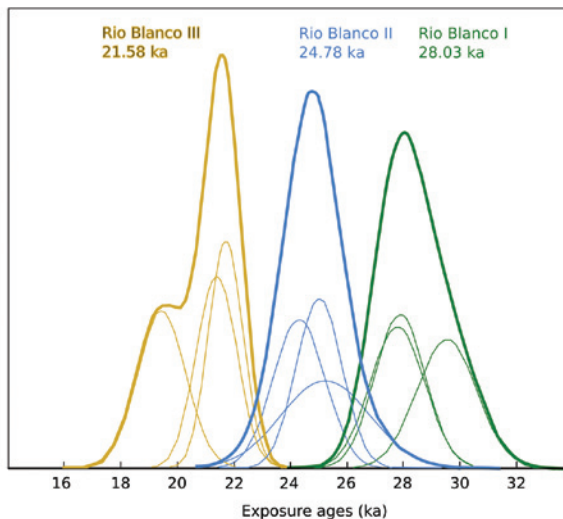


Figure 5. Camel plots summarising distribution of  $^{10}\text{Be}$  ages for each of the Río Blanco moraine limits, outliers identified in the original publications were removed. Thin lines represent individual Gaussian probability distributions for  $^{10}\text{Be}$  ages with  $1\sigma$  internal error; thick line represents the sum of the individual probability distributions.

geomorphological mapping. The profiles revealed that the LP lobe had a low surface gradient and a maximum thickness of ~1100 m at this location, a result consistent with previous modelling (Hubbard *et al.*, 2005).

Recently, Smedley *et al.* (2016) reported OSL ages of  $30.8 \pm 5.7$  ka and  $34.0 \pm 6.1$  ka from glacial outwash related to the Fenix moraines (LBA). They concluded that an earlier ice advance occurred and deposited the outwash, but the corresponding moraine was not preserved. It is likely the moraine was overrun by a subsequent ice advance or removed by meltwater. These results suggest the local LGM may have occurred earlier than indicated based on the preserved moraine record. Supporting evidence for an earlier advance comes from a single  $^{10}\text{Be}$  age from a boulder on a small moraine crest just outside of the outermost Río Blanco limit in the LP valley that gave an age of  $35.1 \pm 1.2$  ka (Hein *et al.*, 2010). The retreat of the ice margin initiated at approximately the same time in both the LBA and LP valleys at ca. 19 ka, and coincided with the onset of ice surface lowering revealed by the vertical reconstruction of the ice sheet surface in the LP/Chacabuco valleys (Boex *et al.*, 2013). The initial deglaciation in both valleys was interrupted by subsequent re-advances. The LBA outlet lobe initially retreated to an unknown distance, leading to the formation of a proglacial lake. A subsequent re-advance then deposited the Menucos moraine above varved lake sediment (Douglass *et al.*, 2006; Kaplan *et al.*, 2004; Turner *et al.*, 2005). Cosmogenic  $^{10}\text{Be}$  ages from moraine boulders on the Menucos moraine yielded range between 18.3 ka and 15.2 ka (Douglass *et al.*, 2006). Given that most geological processes act to reduce apparent exposure ages, the older ages of 18.3 ka are probably more representative of the true age of the moraine. Cemented-carbonate concretions in the stratigraphically older varved sediments gave three radiocarbon ages of  $15.3 \pm 0.3$ ,  $15.3 \pm 0.2$ , and  $17 \pm 0.5$  cal ka BP (Kaplan *et al.*, 2004; Sylwan, 1989). However, these ages should be interpreted with caution as the origin of carbonate concretions is uncertain (Hein *et al.*, 2010). Smedley *et al.* (2016) reported an OSL age of  $14.7 \pm 2.1$  ka from glaciofluvial sediment stratigraphically older than Menucos moraine, and thus there remains an element of uncertainty on the age of the Menucos moraines. After their deposition, the glacier most likely retreated permanently back into the LBA basin.

A re-advance or still-stand also occurred in the LP and Chacabuco valleys depositing the Lago Posadas and Lago Columna moraines (Hein *et al.*, 2010). There are limited data to constrain the age of these moraines. A boulder on the Columna moraine yielded a  $^{10}\text{Be}$  age of  $19.2 \pm 2.4$  ka (Boex *et al.*, 2013), and a boulder on bedrock in the LP valley, which provided a minimum age for the readvance, had a  $^{10}\text{Be}$  age of  $17.3 \pm 1.1$  ka (Hein *et al.*, 2010). The latter age, which came from an erratic that was submerged in a pro-glacial lake, was also used to infer ice retreat of more than 80 km and draining of the lake by this time. Rapid thinning initiated at ca. 18 ka, as revealed by  $^{10}\text{Be}$  ages constraining the ice surface elevation, which lowered to expose both Cerro Oportus and Cerro Tamango completely in approximately 1000 years, equivalent to about 1000 m of ice-surface lowering (Boex *et al.*, 2013). The Lago Augusta (440 m) and Lago Edita (570 m) lake records indicate that the Chacabuco Valley was ice free and no longer inundated by a proglacial lake by  $15.2 \pm 0.1$  cal ka BP

(Villa-Martínez *et al.*, 2012), or as early as  $19.6 \pm 0.1$  cal ka BP (Henríquez *et al.*, 2017). Two radiocarbon ages obtained from a kettle hole near Cerro Ataud, interpreted as minimum deglaciation ages, indicate that the ice had receded close to the Andean foothills by ca. 16 cal ka BP (Turner *et al.*, 2005). Cosmogenic  $^{10}\text{Be}$  ages for the Río Blanco moraines (Hein *et al.*, 2010), Columna moraine, vertical ice surface profiles, minimum radiocarbon deglaciation ages and the Lago Augusta stratigraphy all appear consistent, within uncertainties, suggesting rapid deglaciation of more than 100 km occurred between ca. 19 ka and 16 ka.

There is growing evidence to suggest that in north-western and southern Patagonia, the local LGM occurred during Marine Isotope Stage (MIS) 3, earlier than in central Patagonia. On Tierra del Fuego, the Bahía Inútil-San Sebastián ice lobe (53-54° S) advanced about 100 km beyond the global LGM limit at 45.6 ka and 30.1 ka (Darvill *et al.*, 2015). Sagredo *et al.* (2011) dated the local maxima of the Última Esperanza lobe (51-52° S) to ca. 39 ka, and found no limits corresponding to the global LGM. Two exposure dates from a moraine in the Lago San Martín valley (49° S) tentatively indicate an advance at about 34-37 ka (Glasser *et al.*, 2011). In north-western Patagonia, the radiocarbon chronology from the Chilean Lake District revealed advances at 33.6 cal ka BP and 30.8 cal ka BP (Denton *et al.*, 1999). Pre-LGM advances have been documented elsewhere in the Southern Hemisphere as well; in New Zealand, the Ohau Glacier (44° S) advanced at ~32 ka (Putnam *et al.*, 2013), the Pukaki Glacier (44° S) advanced at ~65 ka (Schaefer *et al.*, 2015) and again at ~42 ka (Kelley *et al.*, 2014), and the Mt. Field glaciers in Tasmania advanced at 44-41 ka (Mackintosh *et al.*, 2006).

Deglaciation of the LBA, LP and Chacabuco valleys preceded the onset of warming (18 ka) identified in a marine core off Taitao peninsula (Montade *et al.*, 2013; Siani *et al.*, 2010) and field evidence suggests that deglaciation occurred earlier and more rapidly than predicted by modelling experiments (Hubbard *et al.*, 2005). Deglaciation in central Patagonia appears to have initiated about 1000 years earlier than in southern Patagonia or the Chilean Lake District. Recession of the Última Esperanza ice lobe started after 17.5 ka (Sagredo *et al.*, 2011), deglaciation in Torres del Paine was underway by ca. 17.4 ka (García *et al.*, 2014), the Strait of Magellan and Bahía Inútil were underway by ca. 17 ka (Kaplan *et al.*, 2008b; McCulloch *et al.*, 2005a), and in the Chilean Lake District, ice recession was underway by ca. 17.8 ka (Moreno *et al.*, 2015; Denton *et al.*, 1999). The timing is similar to the onset of deglaciation in New Zealand (Rother *et al.*, 2015; Putnam *et al.*, 2013; Kelley *et al.*, 2014; Doughty *et al.*, 2015; Schaefer *et al.*, 2006). The cause of early, rapid deglaciation in central Patagonia likely stems from many processes. Rapid deglaciation here has been linked to southward migration of the SWWs and regional warming (Boex *et al.*, 2013), which could also help to explain the comparatively early deglaciation. Modelling experiments indicate the PIS in central Patagonia is stable in two states, full glacial and full interglacial (Hubbard *et al.*, 2005). At these times, the ice sheet is relatively insensitive to changes in climate. However, during deglaciation, the modelling suggests the ice sheet reached a tipping point where it rapidly switched between stable states and deglaciated, losing some 85% of its volume in 800 years (Hubbard *et al.*, 2005). Thus, while the trigger for initial deglaciation will relate to an external driver, the feedbacks within the system,

including the change in dynamics associated with ice calving into deep pro-glacial lakes, must play a significant role in the rapidity of ice decline following the LGM.

## 4.2. The Lateglacial

### 4.2.1. Western Andes

On the western side of the Andes there are no directly dated Lateglacial ice limits. There are a series of moraines indicating an expansion of the NPI outlet glaciers (San Quintín, San Rafael and Gualas/Reicher Glaciers) to form three piedmont lobes. It has been suggested that these moraines might be correlative (Heusser, 2002). Tentatively dated sedimentological records suggest that Golfo Elefantes remained ice free since ca. 11.3 ka, and thus the moraines deposited along its western edge by the Gualas/Reicher Glacier are probably older than this age (Fernandez *et al.*, 2012). The San Rafael Glacier advanced into a pre-existing moraine system during the early Holocene (see section 4.3.1.). The exact timing for the formation of these piedmont lobes remains to be established. The existing evidence suggests the Taitao Peninsula and the Chonos Archipelago likely remained ice free for most of the Lateglacial.

In the marine pollen record, gradual *Nothofagus* forest expansion was recorded from ca. 17.6 cal ka BP, representing a first step of warming (Montade *et al.*, 2013). Closed *Nothofagus-Pilgerodendron-Podocarpus* forest developed across the Chonos Archipelago and Taitao Peninsula between 14-13 cal ka BP and showed no reductions in the forest cover once a closed forest became established (Haberle and Bennett, 2004; Bennett, 2000; Lumley and Switsur, 1993). Humid rainforests, once established, tend to be insensitive to smaller climatic changes. However, a high-resolution marine pollen record indicated there was an increase in Magellanic moorland (*Astelia*) at 14.5 cal ka BP, interpreted as a sudden precipitation increase accompanied by a pause in the warming trend and coinciding with the ACR. The subsequent retreat of the Magellanic moorland after 12.8 cal ka BP marks the second warming step during the Lateglacial and provides a strong evidence for the end of the ACR (Montade *et al.*, 2013). A Chironomid record from Laguna Stibnite (Taitao Peninsula) shows a peak in abundance of *Podonominae* between 13.1-10.5 cal ka BP, interpreted as indicative of cool and oligotrophic lake waters. The changes in the chironomid assemblages were small, however, and any cooling during this period, coeval with the YD would have been minimal (Massaferro and Brooks, 2002).

### 4.2.2. Eastern Andes

#### 4.2.2.1. Glacier advances

Evidence for the extent of glaciers during the Lateglacial is sparse on the eastern side of the Andes. Four moraines deposited by Tranquilo Glacier, Cerro San Lorenzo, were dated to the ACR using  $^{10}\text{Be}$  (E. Sagredo, personal communication, 2017). These provide the only direct evidence for an ACR glacier re-advance in central Patagonia. A single  $^{10}\text{Be}$  age of  $14.2 \pm 0.7$  ka from the Lago Tranquilo cirque moraine, and two  $^{10}\text{Be}$  ages of  $13.2 \pm 0.9$  and  $12.2 \pm 0.8$  ka, from boulders on bedrock in the Río Bayo valley have also been reported (Glasser *et al.*, 2006). Unlike moraines, exposure ages from

boulders on bedrock do not represent advances or still-stands, but in this case they do suggest that the ice margin withdrew to within ~30 km of the present day margin by ca. 13 ka. Glasser *et al.* (2012) obtained  $^{10}\text{Be}$  ages from moraines in several valleys east of the present day NPI (Fig. 2), ranging from ca. 11.4–10.0 ka, which represent advances near the Lateglacial/Holocene transition. Although most of these moraines have apparent early Holocene exposure ages, within uncertainties, they might represent the culmination of YD advances/still-stands (Glasser *et al.*, 2012). We note that in certain locations east of the NPI, such Lateglacial advances may be intricately linked with the development of regional lakes that have the potential to complicate surface exposure ages if samples had been submerged.

In contrast, Lateglacial glacier advances are well-documented in southern Patagonia. Robust radiocarbon and cosmogenic nuclide chronologies from Lago Argentino (50° S; Fig. 1) showed that outlet glaciers advanced during the ACR, deposited the Puerto Bandera moraines, and subsequently retreated during the YD, reaching within 20 km of the present day glacier margins by 12.2 ka (Strelin *et al.*, 2011; Kaplan *et al.*, 2011; Ackert *et al.*, 2008).  $^{10}\text{Be}$  and radiocarbon chronologies from Torres del Paine (51° S), revealed that the outlet glaciers of the SPI re-advanced during the ACR and ice recession was underway by about 12.5 ka (García *et al.*, 2012; Moreno *et al.*, 2009; Fogwill and Kubik, 2005). There is geomorphic and stratigraphic evidence for a re-advance of the Última Esperanza lobe (52° S), tentatively assigned to 14.8–12.8 ka (Sagredo *et al.*, 2011). In Tierra del Fuego (54.6° S),  $^{10}\text{Be}$  ages from cirque moraines showed deposition contemporaneous with the ACR (Menounos *et al.*, 2013). A re-advance of the Cordillera Darwin outlet glaciers into the Strait of Magellan, coeval with the ACR, has been suggested by McCulloch *et al.* (2005a, 2005b). Still-stands or minor re-advances coeval with the YD interrupted the overall ice retreat from the maximum ACR position at numerous sites in southern Patagonia (Strelin *et al.*, 2011; Moreno *et al.*, 2009; Kaplan *et al.*, 2011). A similar pattern of glacier advances culminating at the end of the ACR, followed by retreat during the YD has been documented in New Zealand (Kaplan *et al.*, 2010; Kaplan *et al.*, 2013; Putnam *et al.*, 2010; Turney *et al.*, 2007). These glacier advances occurred during the northern hemisphere Bølling/Allerød interstadial.

#### 4.2.2.2. Lake evolution

The over-deepened nature of the glacially-eroded basins in central Patagonia led to the formation of large proglacial lakes as ice retreat progressed. These lakes initially drained eastward into the Atlantic Ocean because the low-lying structural gap in the Andes (48° S) was blocked by the PIS. As ice retreat progressed, several of these lakes are thought to have become interconnected as a large regional lake east of the present NPI (Turner *et al.*, 2005; Bell, 2008; Hein *et al.*, 2010; Glasser *et al.*, 2016; Bourgois *et al.*, 2016). The position of ice margins to the west and moraine dams to the east determined the location of low overspill routes and hence controlled the former lake levels (Glasser *et al.*, 2016). The retreating ice margin caused a gradual re-organization of the drainage, leading to a final sudden drainage of the lake into the Pacific as the ice sheet separated and the Río Baker Valley opened up (Turner *et al.*, 2005; Bell, 2008; Hein *et al.*, 2010; Glasser *et al.*, 2016). This forced a 200 km migration of the water divide, a major re-



organization of the drainage and released between 160 km<sup>3</sup> and 1150 km<sup>3</sup> of freshwater into the Pacific Ocean (Glasser *et al.*, 2016; Hein *et al.*, 2010). Therefore, the palaeo-lake history has implications for understanding of the wider deglacial history of the region, the dynamics of glaciers terminating in such lakes, and the impact of freshwater release on the regional oceanic and climate system in the Pacific. However, the timing and sequence of events relating to the evolution of palaeo-lakes in the LBA/LP basins, is the subject of debate.

Evidence of former lake levels in the form of raised shorelines, deltas and beaches has been documented in the LBA, LP and Chacabuco valleys (Turner *et al.*, 2005; Bell, 2008; Hein *et al.*, 2010; Glasser *et al.*, 2016; Bourgois *et al.*, 2016). The shorelines can be observed as continuous features at the eastern ends of LBA and LP, but become discontinuous toward the interior of the range. The elevations vary within and between each valley, but there appears to be enough coherence to suggest a series of palaeo-lake levels with an overall tilt toward the east; the tilt and variance are thought to be a consequence of differential isostatic rebound (Turner *et al.*, 2005).

Turner *et al.* (2005) mapped three distinct shorelines in the LP and LBA valleys, on the basis of which a three-stage lake evolution was inferred. One shoreline mapped at 654-633 m a.s.l., which is confined to the LP basin, represented an upper lake level which drained eastwards via the Estancia Caracoles channel (475 m). Retreat of the ice margin exposed a part of Río Baker basin, allowing LP to drain into LBA to create a single lake at a lower elevation of 512-489 m a.s.l. and 484-390 m a.s.l. in LP and LBA valley respectively, and leading to the abandonment of the Estancia Caracoles channel. The minimum age for the abandonment of this channel is ca. 13 cal ka BP on the basis of radiocarbon ages from peat within the channel (Turner *et al.*, 2005; Mercer, 1976), but probably closer to  $17.3 \pm 1.1$  ka on the basis of a cosmogenic <sup>10</sup>Be age for the timing of lake drainage (Hein *et al.*, 2010). The above mentioned radiocarbon age is unlikely to be a close minimum age as organic matter is limited and slow to develop in the arid environment east of the Andes. The single pro-glacial lake drained via a channel (near Perito Moreno) eastwards into the Río Deseado and to the Atlantic Ocean. The higher united lake must have formed and drained sometime before ca. 16 cal ka BP, based on the two <sup>14</sup>C ages from a kettle hole near Cerro Ataud (Turner *et al.*, 2005) or by ca. 17.3 ka, based on the above mentioned <sup>10</sup>Be age from a boulder on bedrock (350 m a.s.l.) in the LP valley (Hein *et al.*, 2010). A shoreline recorded at 397-375 m a.s.l. and 355-305 m a.s.l. in LP and LBA basins, respectively, provides evidence for a lower united pro-glacial lake. This shoreline is constrained by two <sup>10</sup>Be ages from boulders on a delta in Nef Valley ( $14.2 \pm 0.6$  and  $15.4 \pm 0.6$  ka), two direct radiocarbon ages obtained from *in situ* organic material from a kame delta at Río Nef ( $12.7 \pm 0.04$  cal ka BP) and a lake terrace at Mirador Chile Chico ( $13.4 \pm 0.1$  cal ka BP), and several minimum radiocarbon ages from the lake floor (Table 3) (Turner *et al.*, 2005).

Three OSL ages ranging from 12.9 to 10.3 ka have been obtained from deltas above LBA at ~400 m a.s.l. (Glasser *et al.*, 2016), equivalent to the higher single proglacial lake. However, the ages are significantly younger than the ages for the abandonment of the Estancia Caracoles channel (Turner *et al.*, 2005; Mercer, 1976) and for the drainage

Table 3. List of ages associated with former shorelines of a palaeo-lake.

Site and geomorphological context	Shoreline association and age significance	Sample elev. (m)	Dating method*	Age	Calibrated age (BP)** cal median $\pm$ cal range (99.7 %)	Sample number	Reference
Rio Nef, boulder on kame delta	lower united lake, direct age	365	$^{10}\text{Be}^{**}$	14195 $\pm$ 554		DES/00/C1	Turner <i>et al.</i> , 2005
Rio Nef, boulder on kame delta	lower united lake, direct age	368	$^{10}\text{Be}^{**}$	15412 $\pm$ 602		DES/00/C2	Turner <i>et al.</i> , 2005
Rio Nef, kame delta	lower united lake, direct age		14C	14C Age			
Chile Chico Mirador	lower united lake, direct age		14C	10780 $\pm$ 50	12687 $\pm$ 35		Turner <i>et al.</i> , 2005
Cerro Ataud I, kettle hole	lower united lake, min. age		14C	11636 $\pm$ 73	13420 $\pm$ 77		Turner <i>et al.</i> , 2005
Cerro Ataud I, kettle hole	lower united lake, min. age		14C	13550 $\pm$ 95	16267 $\pm$ 157		Turner <i>et al.</i> , 2005
Cerro Ataud II, kettle hole	lower united lake, min. age		14C	13107 $\pm$ 81	15651 $\pm$ 158		Turner <i>et al.</i> , 2005
Cerro Ataud II, kettle hole	lower united lake, min. age		14C	12105 $\pm$ 80	13915 $\pm$ 106		Turner <i>et al.</i> , 2005
Cerro Ataud II, kettle hole	lower united lake, min. age		14C	11578 $\pm$ 71	13372 $\pm$ 72		Turner <i>et al.</i> , 2005
Lago Esmeralda I, kettle hole	lower united lake, min. age		14C	9805 $\pm$ 70	11184 $\pm$ 121		Turner <i>et al.</i> , 2005
Lago Esmeralda II, kettle hole	lower united lake, min. age		14C	10975 $\pm$ 80	12812 $\pm$ 85		Turner <i>et al.</i> , 2005
Lago Esmeralda III, kettle hole	lower united lake, min. age		14C	10752 $\pm$ 87	12654 $\pm$ 77		Turner <i>et al.</i> , 2005
El Mattien, kettle hole	lower united lake, min. age		14C	10472 $\pm$ 72	12284 $\pm$ 150		Turner <i>et al.</i> , 2005
Rio Cochrane, kettle hole	lower united lake, min. age		14C	10840 $\pm$ 95	12712 $\pm$ 85		Turner <i>et al.</i> , 2005
Cordon Esmeralda, kettle hole	lower united lake, min. age		14C	11205 $\pm$ 85	13017 $\pm$ 101		Turner <i>et al.</i> , 2005
Estancia Caracoles palaeochannel	channel abandonment, min. age		14C	11850 $\pm$ 75	13635 $\pm$ 84		Turner <i>et al.</i> , 2005
Estancia Caracoles palaeochannel	channel abandonment, min. age		14C	11245 $\pm$ 245	13069 $\pm$ 226		Mercer, 1976
Puente Santa Maria, aeolian sand capping of a delta	300 - 320 m a.s.l.	333	OSL	8000 $\pm$ 500		PSM01	Glasser <i>et al.</i> , 2016
Península Levican, delta	300 - 320 m a.s.l.	301	OSL	10200 $\pm$ 700		PL01	Glasser <i>et al.</i> , 2016
Rio Ibanez, delta	300 - 320 m a.s.l.	315	OSL	9500 $\pm$ 800		RII02	Glasser <i>et al.</i> , 2016
Las Mercedes, palaeo-beach	300 - 320 m a.s.l.	330	OSL	8500 $\pm$ 1100		MCS01	Glasser <i>et al.</i> , 2016
Bahía Murta, delta	300 - 320 m a.s.l.	315	OSL	10600 $\pm$ 900		BMG02	Glasser <i>et al.</i> , 2016
Fachinal, delta	300 - 320 m a.s.l.	303	OSL	11000 $\pm$ 700		FD02	Glasser <i>et al.</i> , 2016
Jenemén Valley, delta fragment	300 - 320 m a.s.l.	303	OSL	9800 $\pm$ 700		JV02	Glasser <i>et al.</i> , 2016
Bertrand, delta fragment	300 - 320 m a.s.l.	-	OSL	10900 $\pm$ 1600		SB01	Glasser <i>et al.</i> , 2016
Rio Ibanez, flat topped feature	370 - 440 m a.s.l.	396	OSL	12900 $\pm$ 2100		BC02	Glasser <i>et al.</i> , 2016
Puerto Ibanez, delta front foresets	370 - 440 m a.s.l.	396	OSL	10300 $\pm$ 900		RII01	Glasser <i>et al.</i> , 2016
Chile Chico, delta fragment	370 - 440 m a.s.l.	396	OSL	11100 $\pm$ 200		PII02	Glasser <i>et al.</i> , 2016
Los Tres Hermanos, palaeo-beach	370 - 440 m a.s.l.	403	OSL	16800 $\pm$ 1500		WCC01	Glasser <i>et al.</i> , 2016
Las Mercedes, aeolian sand capping, palaeo-beach	460 - 520 m a.s.l.	452	OSL	11800 $\pm$ 1200		LTH01	Glasser <i>et al.</i> , 2016
	460 - 520 m a.s.l.	444	OSL	6500 $\pm$ 700		MCS02	Glasser <i>et al.</i> , 2016
				12000 $\pm$ 1300		MCS02	Glasser <i>et al.</i> , 2016

\*  $^{10}\text{Be}$  - cosmogenic surface exposure dating using  $^{10}\text{Be}$ ,  $^{14}\text{C}$  - radiocarbon dating, OSL - optically stimulated luminescence.\*\* Exposure ages were calculated with the online calculator formerly known as the CRONUS online calculator (v.2.2; constants v. 2.2.1; Balco *et al.*, 2008), southern South America  $^{10}\text{Be}$  production rate (Kaplan *et al.*, 2011), and the time-dependent Lal (1991)/Stone (2000) scaling model, see supplementary table for other scaling models.\*\*\* Radiocarbon ages were recalculated using the OxCal 4.3 online tool (Bronk Ramsey, 2009) and the ShCal13 curve (Hogg *et al.*, 2013), range at (99.7%).

of the higher united lake (Hein *et al.*, 2010). OSL ages from deltas above LBA at ~300-315 m a.s.l., equivalent to the lower united lake, ranged from 11 to 9.5 ka (Glasser *et al.*, 2016), and are also significantly younger than direct and minimum radiocarbon ages for the lower united lake (Turner *et al.*, 2005). The persistence of the lower regional proglacial lake until ca. 8.5 ka, proposed on the basis of the two youngest OSL ages from deltas above LBA (Glasser *et al.*, 2016), or 6.7 ka as inferred from the absence of volcanic ash on top of the lowest delta at Fachinal (Bell, 2008), appears at odds with the radiocarbon and cosmogenic nuclide age constraints. It is more likely that the final drainage of the lower pro-glacial lake into the Pacific Ocean occurred at ca. 12.7 ka on the basis of direct radiocarbon ages from the shoreline (Turner *et al.*, 2005). Yet, several interpretations on the exact sequence and timing of events exist (Turner *et al.*, 2005; Bell, 2008; Hein *et al.*, 2010; Glasser *et al.*, 2016; Bourgois *et al.*, 2016) and further work is required to resolve this.

#### 4.2.2.3. Lateglacial palaeoenvironmental and climate records

Following ice retreat, most records from the eastern side of the Andes indicate an open landscape with sparse scrub-steppe dominated by *Poacea*, *Asteraceae*, and *Ericaceae* under persistent colder and drier conditions at most sites until ca. 14 cal ka BP (Markgraf *et al.*, 2007; de Porras *et al.*, 2012; Iglesias *et al.*, 2016). Steppe taxa such as *Plantago* and *Acaena* was often present indicating disturbance and unstable post-glacial conditions (de Porras *et al.*, 2012; de Porras *et al.*, 2014; Markgraf *et al.*, 2007; Iglesias *et al.*, 2016). Small abundances of hygrophilous arboreal taxa in records from the eastern side of the Andes were interpreted as a long distance transport from forest refugia along the Pacific coast (Markgraf *et al.*, 2007; de Porras, 2014). Iglesias *et al.* (2016) inferred ~200 mm annual precipitation required to support the reconstructed vegetation community (44° S 71° W). Subsequently, a gradual increase in *Nothofagus* taxa suggests grass-steppe vegetation with scattered trees and an increase in effective moisture, but still less than modern values (de Porras *et al.*, 2012; Markgraf *et al.*, 2007; de Porras *et al.*, 2014). The pollen record from Lago Los Niños suggests open forest between 14.3-13.2 cal ka BP, with a subsequent trend towards more open landscape and drier conditions until 11.5 cal ka BP (Iglesias *et al.*, 2016). Fluctuations of aquatic taxa in pollen records attest for high millennial environmental variability between ca. 14-11 cal ka BP (Markgraf *et al.*, 2007; Villa-Martínez *et al.*, 2012)

Contrary to the above, wetter and colder conditions throughout the Lateglacial period are inferred from the Lago Augusta and Lago Edita pollen record on the basis of arboreal taxa, such as *Nothofagus*, hygrophilous *Fitzroya/Pilgerodendron* and *Drimys* present immediately after the onset of sedimentation and persisting until 11.8 cal ka BP (Villa-Martínez *et al.*, 2012; Henríquez *et al.*, 2017). The pollen assemblages reflect patches of evergreen forests in an open landscape, likely a forest refugia in Chacabuco Valley (Henríquez *et al.*, 2017), sustained by the enhanced influence of the SWW at this latitude. A decline in hygrophilous and cold-resistant taxa at ca. 11.8 cal ka BP is concurrent with an increase in *Nothofagus* and likely reflects warming and reduced precipitation due to the decreased influence of the SWWs.

### 4.3. Holocene

#### 4.3.1. Western Andes

On the western side, the only evidence of Holocene glacier advances comes from the Rafael Glacier, a large outlet of the NPI. OSL ages from the Tempanos moraines (9.7 ka, 9.3 ka and 5.7 ka) and glacial outwash (7.7 ka and 7.8 ka) suggested that Rafael Glacier was either more extensive between 9.7 ka and 5.7 ka, or advanced multiple times during this time period (Harrison *et al.*, 2012). However, non-climatic factors related to calving dynamics may have driven the Holocene fluctuations of Rafael Glacier. Previous sedimentological work demonstrated the complex composition of these moraines, resulting from sediment reworking by a glacier re-advance into its former proglacial area (Glasser *et al.*, 2006). Holocene advances of the Rafael Glacier correlate with short term variations in offshore terrigenous sediment input at 10.1-9.1 ka and 7.8-6.5 ka, probably indicating glacier re-advances (Siani *et al.*, 2010). Palaeoenvironmental archives, however, indicate warmer (~1° to 2° C above present mean temperature) and drier conditions during the early Holocene, as suggested by forests in the Chonos Archipelago dominated by *Tepualia* and *Weinmannia* between 12.4-6.7 ka (Haberle and Bennett, 2004). Palynological analysis of a marine core recovered off the Taitao Peninsula ~46° S (MD07-3088) showed an increase in heliophytic taxa, such as *Myrtaceae* and *Tepualia*, from 11.5 to 7.4 ka, indicating the warmest and driest conditions of the Holocene (Montade *et al.*, 2013). This is further supported by a chironomid record from Laguna Stibnite which suggested conditions drier than today between 10.5-7.2 cal ka BP (Massaferro and Brooks, 2002).

With the exception of one OSL date from the Tempanos moraines (see above), no moraines directly dated to the mid-Holocene have been reported. Sedimentological and geochemical analysis of two fjord sediment cores suggested that Gualas Glacier advanced towards the Golfo Elefantes shore between ca. 4 ka and 0.8 ka (Bertrand *et al.*, 2012). The chironomid record from Laguna Stibnite suggests a trend towards wetter conditions during the mid-Holocene indicating intermediate precipitation between 7.2-5.2 cal ka BP and higher precipitation between 5.2-2.4 cal ka BP (Massaferro and Brooks, 2002). Slightly cooler and wetter conditions were also inferred from an increase in arboreal taxa, such as *Pilgerodendron*, in the marine pollen record from ca. 7.4 cal ka BP (Montade *et al.*, 2013).

#### 4.3.2. Eastern Andes

The Fachinal moraines were deposited on top of a raised delta above LBA and have <sup>10</sup>Be and <sup>36</sup>Cl ages of 8.9 ka (peak, inner moraine) and 10.9 ka (peak, outer moraine) (Douglass *et al.*, 2005). The ages, including repeat measurements, in particular for the outer moraines, show a considerable scatter (Fig. 6). <sup>10</sup>Be ages from moraines east of the NPI (Leones and Nef valleys, Lago Negro, Lago Bertrand) might represent early Holocene advances (see section 4.2.2.1) (Glasser *et al.*, 2012), correlative to the outer Fachinal moraines.

Palaeoenvironmental records suggest a drier and warmer period in the early Holocene. At ca. 11 ka, most records detected an increase in the frequency and magnitude

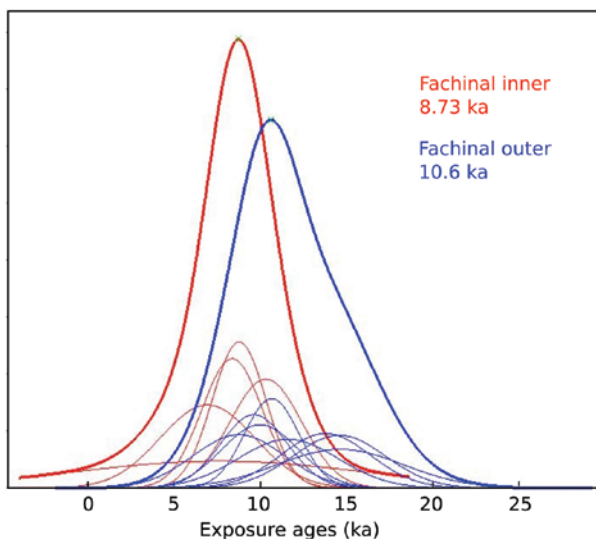


Figure 6. Camel plots summarising distribution of  $^{10}\text{Be}$  ages for the inner and outer Fachinal moraines, outliers identified in the original publications were removed. Including re-calculated  $^{36}\text{Cl}$  ages (Douglass *et al.*, 2005) in the probability distribution (not shown) makes the peak age of the outer Fachinal moraine slightly younger (9.73 ka). Thin lines represent individual Gaussian probability distributions for  $^{10}\text{Be}$  ages with  $1\sigma$  internal error; thick line represents the sum of the individual probability distributions.

of fires (de Porras *et al.*, 2012; de Porras *et al.*, 2014; Markgraf *et al.*, 2007; Villa-Martínez *et al.*, 2012), accompanied by a sudden change from aquatic to a dominance of marsh/bog taxa (de Porras *et al.*, 2012), including a short-lived drying of a fen at Mallín Pollux (Markgraf *et al.*, 2007). However, the concurrent expansion of *Nothofagus* forest suggests these changes represent an increase in effective moisture, but below modern values, and summers that were drier and/or warmer than present (de Porras *et al.*, 2012; de Porras *et al.*, 2014; Markgraf *et al.*, 2007). The palynological record from the southern side of Lago San Martín is characterized by dominance of *Poaceae-Ephedra-Chenopodiaceae* taxa between 11.3-9.5 cal ka BP. This pollen assemblage is analogous to modern dwarf-shrub steppe and was interpreted to represent precipitation lower than today, likely around  $200 \text{ mm a}^{-1}$  (Bamonte and Mancini, 2011). The increased frequency of fires, a peak in laminated carbonate deposition, decline of hygrophilous and cold-resistant taxa and the increase of *Nothofagus* taxa at Lago Augusta at ca. 11.8 cal ka BP was interpreted as a temperature increase and precipitation decrease, driven by a shift in the SWWs (Villa-Martínez *et al.*, 2012). A major transgression was documented at Lago Cardiel during early Holocene ca. 11.8 cal ka BP (Gilli *et al.*, 2005), attributed to easterly moisture advection to the extra-Andean region due to decreased SWW influence, which would support decreased precipitation within the Andes.

The cosmogenic  $^{10}\text{Be}$  ages on moraines in the Río Tranquilo valley, indicate that an outlet glacier of the San Lorenzo ice cap advanced at ca. 5.5 ka (Sagredo *et al.*, 2016). Three OSL ages ( $2.6 \pm 0.2 \text{ ka}$ ,  $2.4 \pm 0.5 \text{ ka}$  and  $3.3 \pm 0.9 \text{ ka}$ ) were obtained from glacio-

lacustrine sediments and a section in a terminal moraine of the Leon Glacier (Harrison *et al.*, 2008). The OSL ages are, however, inconsistent with  $^{10}\text{Be}$  ages obtained from the terminal moraine (1.1 ka and 0.9 ka) (Harrison *et al.*, 2008). Several records suggest wetter and/or colder conditions during the mid-Holocene. From 8-3 cal ka BP the Lake Shaman record suggests the highest effective moisture for the whole record (de Porras *et al.*, 2012). A shift to colder, periglacial conditions between 7.4-6.5 cal ka BP was inferred from sediment stratigraphy and a pollen record at a high altitude site at  $\sim 47^\circ\text{S}$ , followed by drier temperate conditions until 3.7 cal ka BP, and a subsequent shift to more humid, temperate conditions (McCulloch *et al.*, 2016). Similarly, a pollen record from Lago San Martín  $\sim 49^\circ\text{S}$  indicated a decrease in moisture levels ( $< 300\text{ mm a}^{-1}$ ) between 8-3 cal ka BP (Bamonte and Mancini, 2011). At other sites, no major changes in the pollen record were observed once the closed *Nothofagus* forest established (Villa-Martínez *et al.*, 2012; Markgraf *et al.*, 2007).

There is evidence from both sides of the Andes for Lateglacial/early Holocene advances in central Patagonia (Glasser *et al.*, 2012; Harrison *et al.*, 2012; Douglass *et al.*, 2005). Variations in the terrigenous sediment record, interpreted as a proxy for glacier activity, supports early Holocene (10 - 9 ka) glacier expansion (Siani *et al.*, 2010). These glacier advances occurred despite the drier and warmer conditions inferred from the palaeoenvironmental proxies. It also contrasts with well-dated chronologies from southern Patagonia, where there was extensive recession of the Upsala Glacier from its Lateglacial maximum to within the present-day limit by  $\sim 9.2$  ka (Kaplan *et al.*, 2016; Strelin *et al.*, 2014; Strelin *et al.*, 2011). The SPI outlet glaciers were most extensive between 6.1-4.5 ka (Kaplan *et al.*, 2016; Strelin *et al.*, 2014; Strelin *et al.*, 2011). Model simulations suggested a northward shift of low level SWWs at and before 9 ka, leading to substantial precipitation increases on the western side of Andes and at high elevation north of  $49^\circ\text{S}$  (Pollock and Bush, 2013).

## 5. Discussion and Conclusions

### 5.1. Local LGM

The LGM is only constrained chronologically at LBA and LP where the Fenix and Río Blanco ice limits represent advances contemporaneous with the global LGM. However, there exists a growing body of evidence from elsewhere in Patagonia and New Zealand to suggest that glaciers reached their maxima earlier in the last glacial cycle, during MIS 3 when southern insolation was more favourable (García *et al.*, in review). The MIS 3 advances are coeval with a peak in the EPICA Dome C dust record between ca. 31-37 ka (Lambert *et al.*, 2008). Such dust peaks have previously been correlated with glacier advances in southern Patagonia (Sugden *et al.*, 2009). There is also a trend of decreasing sea surface temperatures in the marine offshore record, near the coast of Chile ( $41^\circ\text{S}$ ), during the late MIS 3 (Kaiser *et al.*, 2005; Lamy *et al.*, 2004). The apparent absence of an MIS 3 glacier advance in central Patagonia (with the exception of some tentative evidence; Smedley *et al.*, 2016; Hein *et al.*, 2010) is therefore puzzling. The absence of this signal in central Patagonia might be due to: 1) lack of discovery 2) the large LBA/LP outlet glaciers failing to respond to this climate signal; 3) calving into

large proglacial lakes that might have decoupled these outlet glaciers from climate; 4) pre-LGM advances were later obliterated by a more extensive LGM advance; or 5) pre-LGM moraines were removed by focused meltwater erosion as a consequence of the over-deepened nature of the big glacial basins of central Patagonia (e.g. Hein *et al.*, 2017). Determining the geographic extent of the MIS 3 advance and the global LGM should be a priority for resolving the drivers of southern hemisphere mountain glaciation. Thus, in central and, in particular, northern Patagonia, there is a need to derive detailed glacial chronologies for what are assumed to be LGM and pre-LGM moraines. Ideally, moraines should be targeted for dating in valleys where complication by points 3) and 5) can be avoided. Robust chronologies are required to confidently resolve the occurrence, or absence, of the MIS 3 advance in central Patagonia, to determine its geographic extent relative to the global LGM limits, and to explore the drivers of southern hemisphere glaciation through modelling.

## *5.2. Lateglacial*

The Lateglacial ice marginal positions are poorly constrained and the occurrence of the ACR and/or YD is equivocal in central Patagonia. Only at one site, the Tranquilo Glacier, moraines have been dated to the ACR (E. Sagredo, personal communication, 2017). The offshore pollen record is the only palaeoenvironmental proxy, which firmly shows a vegetation change indicative of cooling coeval with the ACR (Montade *et al.*, 2013). Other palaeovegetation proxies, however, do not indicate any significant vegetation changes from which cooling contemporaneous with either the ACR or YD may be inferred. An exception to this is a chironomid record, from which minor cooling at the time of YD was inferred (Massaferro and Brooks, 2002). Proxies from the eastern side of the Andes show high variability in effective moisture during the Lateglacial (15-11 ka), which together with the degree of insensitivity of proxies to minor oscillations in temperature might have masked any less pronounced cooling coeval with either the ACR or YD. Moraines deposited by the eastern outlet glaciers of the NPI, were dated at the Lateglacial/Holocene transition, which might represent YD readvances or stillstands (Glasser *et al.*, 2012). Obtaining minimum and/or maximum radiocarbon ages for some of these moraines would provide further chronological control needed to improve dating resolution. Obtaining chronological constraints from undated, likely Lateglacial, moraines on the western side of the Andes, would fill the spatial and temporal gap and help to evaluate the relative extent of glaciers during ACR and YD. On the eastern side of the Andes moraines unaffected by regional lake inundation should be targeted for dating. The existence of the regional pro-glacial lake implies substantial ice mass and ice margin stabilization to maintain the ice dam. The lake was probably in existence since the retreat of ice from the position of the Menucos moraines and the youngest Río Blanco moraines at LP. The inconsistency in the interpretation of radiocarbon and OSL ages from shorelines renders the history of the lake open to debate. Resolving the lake history is important for understanding the wider deglacial history, especially since such lakes may have decoupled glacier response from climate by invoking calving dynamics.

### 5.3. Holocene

Much of the early work on Holocene glacial advances often relied on single minimum/maximum radiocarbon age from a given locality (see Glasser *et al.*, 2004). It is only in last decade and a half that a wider range of dating techniques such as OSL, more robust radiocarbon chronologies, and cosmogenic nuclide dating have been applied to date Holocene glacial advances in Patagonia (e.g. Strelin *et al.*, 2011, 2014; Kaplan *et al.*, 2016; Harrison *et al.*, 2012). Yet, evidence for Holocene glacier advances in central Patagonia remains sparse and poorly constrained. The available evidence suggests that the glaciers in central Patagonia advanced or stabilised at the Lateglacial/Holocene transition, while the contemporary palaeoenvironmental proxies revealed warm and dry conditions throughout central Patagonia. The explanation for this is unclear, within dating uncertainties, these advances might represent Lateglacial events. Alternatively, more easterly penetration of the SWWs and precipitation due to a diminishing NPI could have promoting advances of glaciers in the higher mountains east of the NPI as suggested by modelling experiments (Hubbard *et al.*, 2005). There is far less evidence for moraines deposited during the mid-Holocene, with only one moraine (Tranquilo valley) directly dated (Sagredo *et al.*, 2016), when palaeoenvironmental records showed colder and/or wetter conditions. The duration of these climatic changes evident from palaeoenvironmental proxies varied between sites. The timing and pattern of Holocene glacial advances in central Patagonia remains elusive. Improvements in cosmogenic nuclide dating methods have enabled increasingly younger surfaces to be dated. In combination with bracketing radiocarbon ages, this approach would provide robust chronological control in order to determine the pattern of Holocene glacier advances.

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