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Citation for published version:

Mendelova, M, Hein, A, Rhodes, A & Xu, S 2020, 'Extensive mountain glaciation in central Patagonia during Marine Isotope Stage 5', *Quaternary Science Reviews*, vol. 227.
<https://doi.org/10.1016/j.quascirev.2019.105996>

Digital Object Identifier (DOI):

[10.1016/j.quascirev.2019.105996](https://doi.org/10.1016/j.quascirev.2019.105996)

Link:

[Link to publication record in Edinburgh Research Explorer](#)

Document Version:

Peer reviewed version

Published In:

Quaternary Science Reviews

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Extensive mountain glaciation in central Patagonia during Marine Isotope Stage 5

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Abstract

The timing and magnitude of glacial advances throughout a glacial cycle can give insight on the underlying drivers of climate change. Our knowledge of glacial activity early in a glacial cycle, however, is limited by incomplete preservation of the geomorphological record. Here, we present a ^{10}Be -dated glacial chronology from early in the last glacial cycle from the Belgrano valley, east of Monte San Lorenzo in central Patagonia. Our chronology reveals the most extensive glacial advance of the last glacial cycle occurred at ~ 75 ka, at the end of Marine Isotope Stage 5. A second, less-extensive advance occurred at ~ 25 ka during the global Last Glacial Maximum. The topographic setting is such that ice growth would have initiated on and around the San Lorenzo massif at the start of the glacial cycle. We suggest reduced southern hemisphere summer insolation at the time played a significant role in driving this early advance. The comparatively smaller glacier extent during the global Last Glacial Maximum is likely a consequence of the build-up of

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the full Patagonian Ice Sheet, which would have caused ice-divide migration, ice-flow re-routing, and increased aridity on its lee side.

Keywords: last glacial cycle, cosmogenic nuclide exposure dating, ice-divide migration, South America, glacial chronology

1. Introduction

1 There is a long standing debate on the timing of glaciation in the southern
2 mid-latitudes, its (a-) synchrony with the northern hemisphere and its forcing
3 mechanisms (Mercer, 1984; Denton et al., 1999a; Sugden et al., 2005). Recent
4 studies in the southern mid-latitudes demonstrated numerous substantial
5 glacial advances, in many cases local maxima, prior to the global Last Glacial
6 Maximum (gLGM, ca. 26 - 19 ka, Clark et al., 2009) (e.g. Zech et al.,
7 2011; Darvill et al., 2015; García et al., 2018; Schaefer et al., 2015; Kelley
8 et al., 2014; Shulmeister et al., 2019; Sutherland et al., 2007). While an
9 early maximum is not uncommon for many mountain glaciers and sectors of
10 large ice sheets (Gillespie and Molnar, 1995; Hughes et al., 2013), it suggests
11 that our understanding of glaciation in the southern mid-latitudes and, in
12 particular its drivers, is incomplete. Orbital parameters, including regional
13 insolation and season duration, latitudinal shifts in southern westerly winds
14 (SWW) and associated oceanic systems, as well as local ice dynamics have
15 been invoked as potential drivers of pre-LGM ice expansion (Doughty et al.,
16 2015; García et al., 2018; Darvill et al., 2015, 2016). Evaluating the drivers
17 of these pre-LGM advances and their role throughout the last glacial cycle
18 is, however, limited by a lack of dated moraines spanning the entire glacial
19 period. Moraine preservation is inherently incomplete due to a range of

erosional processes, including the removal of pre-existing moraines by an advancing glacier. As a result, moraine preservation in a landscape is biased toward the most extensive and the more recent glacial advances. Here, we present a ^{10}Be -dated chronology from the Lago Belgrano valley on the eastern side of Monte San Lorenzo (47.6° S , 3706m) in central Patagonia, Argentina, where a unique topographic setting has allowed preservation of a moraine record from the early stages of glaciation.

2. Physical setting

Monte San Lorenzo is an isolated granitoid massif that is situated ~ 90 km east of the main spine of the Andes on the border with Chile (Chilean name: Monte Cochrane) (Fig. 1). At this latitude, faulting has created a topographic depression that breaks the main chain of the Andes (Lagabriele et al., 2004) to separate the North Patagonian Icefield (NPI) from the South Patagonian Icefield (SPI), and the San Lorenzo massif from the Pacific coast. The Río Baker and Río Pascua exploit this ~ 80 km gap in the mountain chain to drain much of the eastern side of the Andes, including the large cross-border lakes of Lago General Carrera/Buenos Aires (LBA), Lago Cochrane/Pueyrredón (LP), and Lago O’Higgins/San Martín (SM), westward to the Pacific Ocean. The San Lorenzo and southern massifs constitute an area of high topography stepped east of the main Andean chain (Fig. 1 C). They are enclosed on their southern, western and northern sides by deep glacial valleys, some of which have been eroded to below sea level (e.g. LP and LBA; Murdie et al., 1998). By comparison, their eastern side is less affected by glacial erosion. Lago Belgrano lies at over 800 m asl. and Lago

44 Burmeister at 900 m asl. The modern drainage divide runs just east of Lago
45 Belgrano (Fig. 1 C), with this lake and the eastern outlet glaciers of San
46 Lorenzo draining via a series of interconnected lakes westward into the Pa-
47 cific Ocean. Lago Burmeister and the eastern part of the Belgrano valley
48 instead drain eastward into the Atlantic Ocean via the Río Belgrano.

49 The geology of the area comprises Paleozoic basement rocks, intruded and
50 overlain by Jurassic volcanic rocks and sedimentary sequences of Cretaceous
51 and Jurassic age (Ramos et al., 1982; Ramos and Kay, 1992). The eastern
52 part of the Belgrano valley is mainly characterized by Neogene and Paleogene
53 volcanic sequences, the latter is represented by the Meseta Belgrano, which
54 separates the Belgrano and Pueyrredón valleys (Ramos et al., 1982; Ramos
55 and Kay, 1992). Monte San Lorenzo is part of the Patagonian Batholith
56 intrusion, and is the nearest source of granitic boulders.

57 The climate of Patagonia is dominated by the SWW, which deliver precip-
58 itation to the present glaciers and icefields. The SWW are strongest during
59 austral summer with the core situated between $45^{\circ} - 55^{\circ}$ S. In winter, they
60 weaken and expand equatorward (Garreaud et al., 2009). The Andes create
61 an orographic barrier to the westerly storms resulting in a strong west-east
62 precipitation gradient. On the Pacific coast, precipitation reaches 5000 - 10
63 000 mm a^{-1} (Garreaud et al., 2013). East of the mountain divide precipi-
64 tation decreases sharply to less than 300 mm a^{-1} within tens of kilometres
65 from the mountain front (Garreaud et al., 2013). Air temperatures do not
66 exhibit such a strong west-east gradient, and are pre-dominantly controlled
67 by the SWW (Kilian and Lamy, 2012).

68 The San Lorenzo massif, situated about 160 km from the Pacific coast,

69 has a transitional maritime to continental climate (Falaschi et al., 2013).
70 At this latitude, Monte San Lorenzo is the first orographic barrier to the
71 SWW, which sustain this easternmost extensively glaciated area in the region
72 (Falaschi et al., 2013). Glaciers of Monte San Lorenzo and the surrounding
73 mountains cover an area of $\sim 207 \text{ km}^2$ (Falaschi et al., 2013). The snowline
74 is estimated to be $\sim 1700 \text{ m}$ asl on the western side and about 100 m higher
75 on the eastern side of San Lorenzo (Falaschi et al., 2013). Glaciers of cen-
76 tral Patagonia can be characterized as warm and wet, and are considered
77 particularly sensitive to temperature changes (Sagredo and Lowell, 2012).

78 During the last glacial period, several tributary glaciers draining the San
79 Lorenzo and southern massifs coalesced and flowed down the Belgrano valley
80 (here referred to as the Belgrano glacier). By comparison with other outlets
81 glaciers in central Patagonia, the Belgrano glacier was considerably smaller
82 and occupied a relatively high-altitude valley. Three main moraine systems
83 exist in the valley east of Lago Belgrano. Based on local landmarks, we
84 informally name these moraine belts, Klementek, Menelik and Belgrano from
85 the oldest to the youngest (Fig. 2). Here, we focus on the Klementek and
86 Menelik moraine systems.

87 **3. Glacial history and previous work**

88 *3.1. Previous work at Lago Belgrano*

89 The glacial history of the Belgrano valley has received little scientific
90 attention and is therefore largely unknown. Even Caldenius (1932), who
91 mapped nearly all of Patagonia in his pioneering work, did not map the Bel-
92 grano valley. The area of Lago Belgrano was included in a Patagonia-wide

93 glacial geomorphological map (Glasser et al., 2008), and partly covered by a
 94 more detailed geomorphological mapping by Bendle et al. (2017b). The only
 95 field-based mapping and chronological work in the Belgrano valley was done
 96 by Wenzens (2005) and Horta et al. (2017). Wenzens (2005) documented
 97 the glacial geomorphology of the valley and tentatively correlated the outer-
 98 most moraine belt (the Klementek Moraines) to the LGM. He attributed the
 99 Menelik moraine system to the Late Glacial based on a radiocarbon age of
 100 14.85 ± 0.12 ^{14}C ka from a kettle hole (18.0 ± 0.2 cal ka BP ¹). Horta et al.
 101 (2017) focused on the palaeolake history and obtained radiocarbon ages from
 102 lacustrine sediments inboard of the Menelik moraines. The Late Glacial and
 103 Holocene glacier history has been investigated by several workers, mainly on
 104 the western and northern flanks of Monte San Lorenzo (Mercer, 1968; Davies
 105 et al., 2018; Glasser et al., 2012; Sagredo et al., 2017, 2018; Martin et al.,
 106 2019).

107 *3.2. Pre- gLGM glacial history of Patagonia*

108 In Patagonia, the evidence for pre-gLGM ice expansion is largely re-
 109 stricted to Marine Isotope Stage (MIS) 3 and earlier glacial cycles. In south-
 110 ern Patagonia ($49^\circ - 55^\circ$ S), several studies indicate MIS 3 advances that were
 111 significantly more extensive than the MIS 2 advances. On Tierra del Fuego
 112 ($53-54^\circ$ S), Darvill et al. (2015) determined that the Bahía Inútil - San Se-
 113 bastián ice lobe advanced at $45.6 (+139.9/-14.3)$ ka and $30.1 (+45.6/-23.1)$
 114 ka, and was up to 100 km more extensive than the subsequent gLGM lim-

¹Recalibrated using the OxCal online calibration program (version 4.3; Bronk Ramsey,
 2009) and the SHCal13 calibration curve (Hogg et al., 2013)

115 its (Kaplan et al., 2004; McCulloch et al., 2005). The Torres del Paine and
 116 Última Esperanza outlet glaciers (50° S) reached their maximum at $48.0 \pm$
 117 1.6 ka with subsequent advances at 39.2 ± 2.0 ka and 34.0 ± 1.3 ka dated by
 118 ^{10}Be exposure ages (García et al., 2018; Sagredo et al., 2011). Here, the MIS
 119 2 ice advances (21.5 ± 1.8 ka) were half the extent of the MIS 3 maximum
 120 (García et al., 2018). Tentative exposure dates from the San Martín valley
 121 (49° S) follow a similar pattern (Glasser et al., 2011).

122 In northern Patagonia (40° – 44° S), glaciers also advanced during MIS
 123 3 and MIS 2, but in contrast to southern Patagonia, the relative magnitudes
 124 were similar. In the CLD, stratigraphic evidence and radiocarbon chronology
 125 suggest ice expansion at $33.6 \sim 0.2$ ka, but this was similar in magnitude or
 126 smaller than later advances (Denton et al., 1999a; Moreno et al., 2015). Zech
 127 et al. (2011) reported an advance at 39° S on the eastern side of the Andes
 128 dated by ^{10}Be at ~ 39 ka. Much of the glacial deposits on the eastern side
 129 of the Andes in northern Patagonia remains undated, which precludes firm
 130 conclusions on the magnitude and timing of the maximum ice expansion in
 131 the north.

132 In central Patagonia (44° – 49° S), two chronologies from large former
 133 outlet glaciers suggest maximum ice expansion roughly during the gLGM,
 134 between $\sim 27 - 25$ ka at LBA (Kaplan et al., 2004; Douglass and Bockheim,
 135 2006), and ~ 28 ka and perhaps ~ 35 ka at LP (Hein et al., 2010, 2009).
 136 In both valleys, moraines outboard of the MIS 2 limits pre-date the last
 137 glacial cycle (Kaplan et al., 2005; Hein et al., 2017, 2009). Smedley et al.
 138 (2016) identified MIS 3 advances at LBA (34.0 ± 6.1 , 30.8 ± 5.7 ka) using
 139 optically stimulated luminescence (OSL) dating of buried outwash sediment.

140 However, there are no preserved moraines related to this advance, which
141 likely only reached to within the MIS 2 limits.

142 More widely, robust ^{10}Be chronologies from New Zealand indicate glacial
143 advances at ~ 42 ka (Pukaki valley, Kelley et al., 2014), ~ 35.5 ka (Pukaki
144 valley, Doughty et al., 2015) and ~ 32 ka (Ohau valley, Putnam et al., 2013).
145 Overall, ice expansion during MIS 3 appears to be a regional phenomena
146 (Darvill et al., 2016). Despite this, we are still far from fully establishing
147 the magnitude of the MIS 3 glacial expansion throughout Patagonia and its
148 climatic significance.

149 Chronostratigraphic evidence for glacial expansion prior to MIS 3 in Patag-
150 onia is fragmentary. At the time of publishing, there are no firmly dated
151 glacial advances during MIS 4 or 5. In the Chilean Lake District (CLD),
152 the outermost moraine belts of three outlet lobes have been tentatively as-
153 signed to MIS 4 (Denton et al., 1999b; Heusser et al., 1999; Mercer, 1976) but
154 this assignment is not based on firm geochronological data. Palaeoclimate
155 proxies from the southern mid-latitudes (e.g. Kaiser and Lamy, 2010) and
156 Antarctica (Johnsen et al., 1972; EPICA Community Members, 2004) indi-
157 cate cold conditions comparable to the gLGM during MIS 4. Dust fluxes to
158 Antarctica (Lambert et al., 2008) and the Southern Ocean (Martínez-García
159 et al., 2011), which are considered a proxy for glacial activity in Patagonia
160 (Sugden et al., 2009; Kaiser and Lamy, 2010), suggest expansion of Patag-
161 onian glaciers at the time. It is likely that MIS 4 glacial records exist in
162 Patagonia, but they have not yet been firmly dated. By comparison, in New
163 Zealand, full glacial conditions during MIS 4 are well documented (e.g. Mc-
164 Carthy et al., 2008; Shulmeister et al., 2018; Schaefer et al., 2015). The only

165 tentative evidence for ice expansion in Patagonia during MIS 5 comes from
166 two individual ^{10}Be ages (91.4 ± 4.1 ka, 96.8 ± 4.8 ka) from Lago San Martin
167 (49° S; Glasser et al., 2011). As a consequence, little is known about the
168 development of the Patagonian Ice Sheet (PIS) from the onset of the last
169 glaciation towards the gLGM.

170 4. Methods

171 4.1. Geomorphic mapping

172 Geomorphological mapping was done using high resolution imagery (2.5
173 m to sub-meter) available through the ESRITM World Imagery service, along
174 with 10 m Sentinel-2 imagery. An ALOS PALSAR radiometrically terrain-
175 corrected digital elevation model with a resolution of 12.5 m was used to
176 aid landform identification. Geomorphological features were identified and
177 mapped following well established criteria (Glasser et al., 2008; Bendle et al.,
178 2017b; Darvill et al., 2014; Martin et al., 2019). Mapping was supported by
179 ground-truthing during two field campaigns in the austral summers of 2016
180 and 2018.

181 4.2. Dating approach and sampling

182 Cosmogenic nuclide surface exposure ages from glacial boulders are com-
183 monly used to establish the timing of glacial advances/still-stands in a range
184 of environmental settings (e.g. Heyman et al., 2011). The technique works
185 well in most settings on timescales relevant to the gLGM, but it can break
186 down on older moraine systems due to site specific environmental factors that
187 can affect nuclide concentrations, such as moraine degradation and boulder

erosion (Putkonen and Swanson, 2003; Hallet and Putkonen, 1994; Briner et al., 2005; Hein et al., 2009, 2017; Heyman et al., 2011; Gillespie and Bierman, 1995; Applegate et al., 2010). Depending on the environmental conditions, direct dating of glaciofluvial outwash sediments directly linked to a moraine system can circumvent some of these issues to be an effective way of constraining the timing of a glacial advance. This is because outwash terraces are flat, unlike moraine ridges, and less susceptible to gravity-driven surface lowering, which minimises the potential for exhumation of surface clasts. Furthermore, the outwash cobbles rooted in the terraces are subject to less wind erosion, which is a significant source of boulder erosion in Patagonia. The outwash method has proved effective at dating pre-LGM glacial advances as old as one million years in this region (Hein et al., 2009, 2011, 2017; Darvill et al., 2015; Cogež et al., 2018). To develop our chronology, we measured cosmogenic ^{10}Be in both moraine boulders and glacio-fluvial outwash cobbles.

To date the Menelik moraines, we sampled six boulders on the outermost moraine ridges, one boulder on the inner moraine and two samples came from moraines near the northern valley side. In addition, we sampled five cobbles from its associated outwash terrace. To date the Klementek system, we sampled six surface cobbles from two sites on the outwash terrace associated with the outermost moraines. Surface boulders on this moraine system are rare; those we found were significantly eroded and unsuitable for dating. The sample details are presented in Table 1. When sampling, we preferentially targeted large (ideally >1 m) stable boulders on moraine crests that showed minimal signs of surface erosion (cf. Heyman et al., 2016; Putkonen

213 and Swanson, 2003). Samples were collected by hammer and chisel from the
214 centre of the boulder surface (Gosse and Phillips, 2001; Darvill, 2013). Out-
215 wash cobbles were collected whole from locations where the outwash terrace
216 could be unambiguously linked to the moraine system, but far enough from
217 the moraine ridges to avoid material transported down its slopes and onto
218 the terrace surface (Hein et al., 2011, 2009).

219 4.3. Preparation

220 Samples were prepared at two cosmogenic nuclide laboratories: the Nat-
221 ural Environment Research Council’s Cosmogenic Isotope Analysis Facility
222 (NERC-CIAF) and the University of Edinburgh’s Cosmogenic Nuclide Lab-
223 oratory. The samples were crushed, milled and then sieved to obtain the 250
224 μm to 500 μm (CIAF), or 250 μm to 710 μm (Edinburgh) size fraction. Large
225 cobbles (>10 cm) were cut to an appropriate thickness parallel to the sur-
226 face prior to crushing, while the majority of the cobbles were fist-sized and
227 these were crushed whole. All measurements of ^{10}Be concentrations were
228 done at the Scottish Universities Environmental Research Centre (SUERC)
229 Accelerator Mass Spectrometry (AMS) facility.

230 4.3.1. NERC-CIAF Preparation

231 The 250 μm to 500 μm fraction was magnetically separated using a Franz
232 machine and the non-magnetic fractions were recovered. To remove carbon-
233 ates and soluble oxides, the samples were first treated with a 6:1:1 solution
234 of water, HCl (37%) and HNO_3 (68%, aqua regia), on a hot-plate ($\sim 50^\circ\text{C}$)
235 and the supernatant was discarded. The samples were then treated with a
236 100:1 mixture of water and concentrated (40%) HF on a shaker table for one

237 hour to slightly etch feldspar and mica grain surfaces. Supernatant liquid
 238 was discarded. Feldspar and mica minerals were then removed through froth
 239 floatation (Herber, 1969): the samples were repeatedly treated with eucalyp-
 240 tus oil, to make the etched grains hydrophobic, and a carbonated surfactant
 241 solution. Floating minerals were then discarded and quartz grains that did
 242 not float were recovered. The samples were then etched three times in a
 243 1 litre solution of water, HF (40%) and HNO_3 (150:2:1) in a high power
 244 ultrasonic tank to isolate the cores of the quartz grains. An aliquot of the
 245 final quartz sand was dissolved and its purity was tested for Al , Be , Fe , Ca
 246 and Ti by ICP-OES at SUERC. The samples were dissolved in pure HF
 247 and each sample and process blanks ($n=2$) were spiked with ~ 0.2 mg of
 248 9Be carrier. After dissolution, the HF was evaporated and replaced by
 249 HCl . The solutions were first passed through anion exchange chromatogra-
 250 phy columns to remove Fe . The Fe -free fraction was then evaporated and
 251 the HCl was replaced by diluted H_2SO_4 . The sulphate solutions were then
 252 passed through cation exchange chromatography columns to remove Ti and
 253 B , and to separate Be and Al fractions. The ^{10}Be fractions were precipi-
 254 tated as hydroxides and oxidised at $\sim 900^\circ C$. Resulting BeO were mixed
 255 with Nb (1:6) and pressed into copper cathodes for AMS analysis. All mea-
 256 surements were normalized to NIST SRM4325 standard with nominal ratios
 257 of $2.79 \times 10^{-11} \text{ } ^{10}Be/^9Be$ (Nishiizumi et al., 2007) and half-life of 1.36 Ma.
 258 The blank corrections ranged between 1 and 4% of the sample $^{10}Be/^9Be$
 259 ratios.

260 4.3.2. *Edinburgh Preparation*

261 At the University of Edinburgh’s Cosmogenic Nuclide Laboratory, quartz
262 was isolated through repeated acid etching following standard procedures
263 (Nishiizumi et al., 2007; Bierman et al., 2002). The samples and the process
264 blanks (n=2) were spiked with ~ 0.25 mg ^9Be carrier. The samples were
265 prepared as BeO targets for AMS analysis following procedures detailed by
266 Hein (2009), which are similar to those described above. $^{10}\text{Be}/^9\text{Be}$ mea-
267 surements were normalized to the standards of Nishiizumi et al. (2007) with
268 nominal ratios of 2.79×10^{-11} . The blank corrections were under 1% of the
269 sample $^{10}\text{Be}/^9\text{Be}$ ratios.

270 4.3.3. *Age Calculations*

271 The ^{10}Be ages were calculated using the online exposure age calculator
272 formerly known as the CRONUS-Earth online calculator (version 3, Balco
273 et al., 2008), with the Patagonia production rate (Kaplan et al., 2011) ob-
274 tained from the ICE-D online database (<http://calibration.ice-d.org/>).
275 Throughout the paper we report ages calculated using the time-dependent
276 scaling scheme of Lal (1991) modified by Stone (2000). Using the New
277 Zealand ^{10}Be production rate (Putnam et al., 2010), our ages decrease by
278 less than 1%, and using the global production rate (Borchers et al., 2016)
279 the ages decrease by $\sim 6\%$. Summary statistics for each moraine system are
280 provided in Table 2. We base our discussion on the arithmetic mean and 1
281 σ standard deviation for each moraine system.

282 The ages presented here assume zero erosion and are therefore minimum
283 ages. Topographic shielding was measured in the field and is negligible for
284 all samples (~ 0.99). The ages are not corrected for snow, vegetation or

285 soil shielding. Vegetation is sparse in the area, mostly comprises steppe
286 grasses and is unlikely to introduce significant shielding. Owing to high winds
287 and the semi-arid environment, we expect snow shielding to be minimal.
288 Soils in the field area are thin (normally 0-10 cm) and spatially variable,
289 and a significant lag is expected before soil develops following ice retreat.
290 Quantifying soil shielding over the exposure history is therefore difficult, but
291 we expect the impact on ages to be minimal.

292 **5. Results**

293 *5.1. Moraine Record*

294 The Klementek moraine system comprises of well-defined outermost moraine
295 ridges with 25 m relief and slopes up to 12° deposited in the eastern part of
296 the valley, ~ 65 km from present day glacier margins (Fig. 2). The outermost
297 moraine ridges are discontinuous and up-valley transition into a hummocky
298 terrain, which extends for 10 km on the northern side of the Río Belgrano
299 (Fig. 3). Hummocks and hollows are interspersed with streamlined, ice-flow
300 parallel landforms of 250 - 450 m in length. Small narrow ridges of up to one
301 kilometer in length, perpendicular to former ice flow, are superimposed on
302 the hummocky terrain, and are interpreted as recessional moraines. Occa-
303 sional, significantly eroded boulders with fluting and ventifacts are scattered
304 on the hummocky terrain. A glacio-fluvial outwash terrace emanates from
305 the distal margin of the outermost moraine ridges. Relict braided channels
306 can be traced from the surface of the outwash to the moraines. The out-
307 ermost moraine ridges are interpreted to mark a former ice margin position
308 of the Belgrano glacier, while the hummocky terrain was likely formed by a

309 combination of subglacial and ice-marginal processes during ice retreat. The
310 morphology of the Klementek moraine system is suggestive of a pre-LGM
311 age.

312 The Menelik moraines (Fig. 4 and 5) are 22 km up valley and are sep-
313 arated from the Klementek moraines by a >10 km long broad glacio-fluvial
314 outwash terrace. The double-lobate shape of the Menelik moraines and the
315 presence of a medial moraine indicate the coalescence of the Belgrano and
316 Burmeister glaciers. The outermost Menelik moraines are prominent, cross-
317 valley ridges with broad crests, and 30 m relief and slopes up to 15° (Fig. 4
318 and 5). Clusters of exposed gravel and abundant boulders can be found on
319 moraine crests. Inboard, moraines diminish in size and in places are inter-
320 spersed with numerous ponds and depressions. Numerous small (1-2 m relief)
321 moraine ridges are preserved on higher ground (Fig. 4 A). The sequence of
322 smaller moraines extends up valley over a distance of ~ 10 km, and suggests
323 deposition during ice recession. Palaeoshorelines indicate formation of shal-
324 low proglacial lakes as the ice margin retreated (Fig. 2). Near the northeast
325 shore of Lago Belgrano, the Menelik moraines are cross-cut by the younger
326 Belgrano moraines (Fig. 2). This indicates that the glacier retreated into
327 the Belgrano trough prior to a subsequent re-advance, which also removed
328 part of the inner Menelik moraine system. Glacio-fluvial outwash terraces
329 emanate from the Menelik moraines with preserved relict surface channels.
330 The outwash plain has been incised by two rivers, leaving the area sampled
331 isolated and unaffected by post deposition fluvial processes.

332 5.2. Exposure ages

333 We present 20 new exposure ages from the Klementek and Menelik moraine
334 systems. The ^{10}Be data are presented in Figures 3 B and 5, and Tables 1-2.
335 The probability density diagrams ('camelplots') are shown in Figure 6. In
336 the following, we report individual ^{10}Be ages with 1σ internal uncertainties.

337 The ages from the Klementek outwash range from 67.1 ± 1.7 ka to $85.9 \pm$
338 1.9 ka ($n = 6$; Fig. 3 B and 6 A). The distribution of ages contains a cluster
339 of four ages, with one younger and one older age. The ages yield a mean of
340 75.7 ± 5.9 ka (all data), or 75.2 ± 2.8 ka with the outliers excluded.

341 The boulder ages from the Menelik outermost moraines range from $18.4 \pm$
342 0.6 ka to 26.0 ± 0.8 ka ($n=6$; Fig. 5 and 6 B). We consider the youngest sam-
343 ple (18.4 ± 0.6 ka) an outlier because it was taken from a short, partly buried
344 boulder and was likely exhumed from the moraine matrix post-deposition.
345 The remaining samples range from 23.0 ± 0.6 ka to 26.0 ± 0.8 ka and yield
346 a mean age of 24.8 ± 1.1 ka. The two boulders from moraines near the Bel-
347 grano valley side yielded ages of 22.9 ± 0.9 and 18.3 ± 0.6 ka. The boulder
348 from the inner Menelik moraine gave an age of 17.0 ± 0.7 ka. Cobble ages
349 from the Menelik outwash (Fig. 5 and 6 C) range from 23.7 ± 0.9 ka to 26.0
350 ± 1.0 ka, excluding an older age of 36.6 ± 1.1 ka, and yield a mean age of
351 24.8 ± 0.9 ka.

352 6. Discussion

353 6.1. The age of the Klementek outwash

354 6.1.1. Geomorphic considerations

355 The distribution of the Klementek outwash cobble ages suggests that
356 the samples may have been affected by pre- and post-depositional processes.
357 One possibility is that the cobbles are erroneously young because they were
358 exhumed through bio- or cryo-turbation, or as the terrace surface lowered
359 through time. We do not expect bio- or cryo-turbation processes to signifi-
360 cantly affect the exposure ages because soils at the site are thin, which should
361 limit this process. Terrace deflation is also expected to be insignificant based
362 on a study in the neighboring Lago Pueyrredón valley to the north. Here,
363 Hein et al. (2009) estimated outwash terrace surface lowering rates at 0.53
364 mm ka⁻¹ using a depth-profile approach. This rate would imply about 45
365 mm of total terrace surface lowering over the past 86 ka (the age of oldest
366 outwash sample), which would not significantly decrease the apparent ex-
367 posure ages. Rock surface erosion is another potential source of geological
368 scatter and erroneously young apparent exposure ages. Our ages assume no
369 rock surface erosion, but minor surface erosion is evident on most Klementek
370 cobble samples (Fig. 3 F). The erosion depth is variable but less than ~ 3
371 mm (Figures 3 F). This amount of erosion integrated over ~ 86 ka (age of
372 oldest sample) yields an erosion rate of 0.035 mm ka⁻¹. Applying this erosion
373 rate to the age calculations would increase the ages by less than 1 %, which
374 is within the analytical uncertainty.

375 Nuclide inheritance is a potential source of geological scatter that would
376 make the ages erroneously old. We assume inheritance is negligible because

the outwash sediment was shielded, transported and eroded by warm-based ice before deposition. This assumption is supported by depth-profile measurements in the neighbouring Lago Puyeredón valley where average inheritance was demonstrated to be negligible (Hein et al., 2009), and by the similarity of outwash cobble and moraine boulder exposure ages. Given the similarity of ages obtained from the Menelik moraine boulders and outwash surface cobbles, we too conclude inheritance is low.

6.1.2. Age interpretation

There are several ways to interpret the outwash exposure ages: 1) If we assume that the oldest sample remained on the surface the longest and the other samples were exhumed later as the outwash terrace deflated, then the oldest age approach (cf. Hein et al., 2009) would provide the closest age for the deposition of the outwash at ~ 86 ka. 2) The youngest and oldest ages were affected by pre- (85.9 ± 1.9 ka, e.g. inheritance) and post-depositional processes (67.1 ± 1.7 ka, exhumation). If we exclude them as outliers, we obtain a mean age of 75.2 ± 2.8 ka ($n=4$), which would be the best possible age for the deposition of the outwash. 3) All ages are correct and the outwash is of composite age, and was deposited during ice expansion between ~ 86 and 67 ka. Given the clustering of ages and our assessment of the significance of terrace deflation, exhumation and inheritance, we tentatively suggest the mean age of 75.2 ± 2.8 ka (scenario 2) as the best estimate for the culmination of the ice advance that deposited the outwash. However, since we cannot exclude other scenarios, we acknowledge that the outwash might be as old as 86 ka or it could have been deposited between 86 - 67 ka.

Our mapping suggests the Klementek moraines are age-equivalent to the

402 dated outwash and therefore represent a glacial advance at 75.2 ± 2.8 ka
403 ka; the morphology of the moraine system is consistent with this age. We
404 acknowledge that the hummocky terrain with superimposed moraine ridges
405 could be a palimpsest feature younger than the dated outwash and could
406 have been deposited during MIS 4 and 3.

407 *6.2. The age of Menelik moraines and outwash*

408 The Menelik outermost moraine boulders yield a mean age of 24.8 ± 1.1
409 ka, excluding one young outlier. The outwash cobbles yield a mean age of
410 24.8 ± 0.9 ka, excluding one older outlier. As with the Klementek outwash,
411 there are several ways to interpret the Menelik outwash ages: 1) the oldest
412 age approach (36.6 ± 1.1 ka, LB1623); 2) the oldest age is an outlier (e.g.
413 contains inherited nuclides or reworked from previous advance) 3) all ages
414 are correct, and the outwash is of composite age. The relative clustering of
415 the boulder and outwash ages indicates that scenario 2) is most likely. Our
416 field observations, however, provide some support for scenario 3, since there
417 is a remnant of a smaller moraine just outboard of the outermost Mene-
418 lik moraine sampled. It is possible the oldest outwash sample (36.6 ± 1.1
419 ka) represent this earlier, slightly more extensive advance, and therefore we
420 tentatively suggest an advance at this time. The remaining four samples,
421 which produce a central peak (Fig. 6 C), would therefore relate to the ma-
422 jor advance that deposited the prominent outermost moraines and (part-)
423 deposited the outwash. Excluding two ages (LB1618 and LB1623), the re-
424 maining outwash and moraine boulder exposure ages are indistinguishable
425 and when combined together yield a mean age of 24.8 ± 1.0 ka, which we
426 interpret as the timing of this major advance.

427 6.3. *The last glacial cycle at Lago Belgrano*

428 Our ^{10}Be ages suggest the most extensive advance of the Belgrano glacier
429 occurred early in the last glacial cycle at 75.2 ± 2.8 ka toward the end
430 of MIS 5. Given the scatter in our exposure ages and the extent of the
431 hummocky terrain with recessional moraines, it is possible the glacier oscil-
432 lated near its maximum position between 86 - 67 ka. The advance precedes
433 major cooling during MIS 4, when some New Zealand glaciers had reached
434 full glacial conditions (e.g. McCarthy et al., 2008; Shulmeister et al., 2018;
435 Schaefer et al., 2015). Whilst we find no unambiguous evidence for an ad-
436 vance during MIS 4, we recognize the youngest age of 67.1 ± 1.7 ka overlaps
437 in timing. We only dated outwash associated with the outermost moraines,
438 it is therefore possible that some of the recessional moraines were deposited
439 during MIS 4.

440 We have tentative evidence for an advance during MIS 3 (36.6 ± 1.1 ka),
441 but this requires further chronological constraint. Significant MIS 3 glacial
442 advances have been identified at several locations elsewhere in Patagonia
443 (Darvill et al., 2015; Smedley et al., 2016; García et al., 2018; Denton et al.,
444 1999b). A second major, but considerably smaller, advance of the Belgrano
445 glacier is represented by the Menelik moraines and occurred at 24.8 ± 1.1
446 ka, coeval with the gLGM.

447 6.4. *Evidence for MIS 5 glaciation*

448 The Belgrano glacier reached its maximum extent much earlier than else-
449 where in Patagonia, where MIS 3 is the more dominant signal. Our data pro-
450 vide the first robust evidence for MIS 5 glaciation in Patagonia. Elsewhere in
451 the southern mid-latitudes, evidence for MIS 5 glaciation is fragmentary. In

southern Patagonia, the maximum expansion of the Lago San Martin outlet glacier may have occurred during late MIS 5 (Glasser et al., 2011). In New Zealand, ^{10}Be ages from the Cascade valley, suggest an advance at ~ 90 ka (Sutherland et al., 2007). Shulmeister et al. (2010) reported glacier advances at $\sim 100 - 90$ ka and ~ 80 ka in the Rakaia Valley based on stratigraphic evidence and OSL ages. The advance of Belgrano glacier could correlate with ice expansion evident in New Zealand during MIS 5b-a.

In the northern hemisphere, there is evidence for expansion of mountain glaciers and ice sheets during MIS 5. For example, glaciers in central Asia (Röhringer et al., 2012; Koppes et al., 2008), NE Siberia (Zech et al., 2011), as well as in NW Washington and in the Pyrenees (Pallàs et al., 2010; Lewis et al., 2009) reached their maximum during MIS 5. Sectors of large northern hemisphere ice sheets expanded to their maximum position early in the last glacial cycle. For example, the Barents-Kara ice sheet reached its maximum extent during MIS 5 (90 - 80 ka), and was less extensive during subsequent glaciations (Svendsen et al., 2004).

These studies demonstrate substantial glaciation of mid-latitude mountain areas globally, as well as the growth of ice sheets, during MIS 5. Our data demonstrates extensive glaciation of the eastern massifs of central Patagonia at this time. We speculate that the onset of glaciation throughout the Patagonian Andes also occurred during MIS 5, but the geomorphic and stratigraphic evidence has largely been destroyed, or remains undiscovered.

6.5. gLGM and deglaciation in Patagonia

Our data show that the Belgrano glacier advanced at 24.8 ± 1.0 ka during the gLGM, but this advance was substantially smaller than the MIS 5

477 advance. The timing of this advance is in good agreement with other glacial
 478 records in Patagonia (e.g. Hein et al., 2010; Kaplan et al., 2004, 2008; McCul-
 479 loch et al., 2005; Denton et al., 1999b). For example, in central Patagonia,
 480 the LP outlet glacier achieved its maximum extent at ~ 28 ka with a subse-
 481 quent advance at ~ 25 ka (Hein et al., 2010). The LBA outlet lobe was at its
 482 maximum at $\sim 27 - 25$ ka followed by several smaller re-advances/still-stands
 483 (Kaplan et al., 2004; Douglass and Bockheim, 2006). Advances during the
 484 gLGM in southern Patagonia were substantially less extensive than the MIS
 485 3 advances (García et al., 2018; Darvill et al., 2015; McCulloch et al., 2005;
 486 Kaplan et al., 2008).

487 The presence of numerous, regularly spaced recessional moraines indicate
 488 that deglaciation from the 25 ka limits at Lago Belgrano was, at least initially,
 489 gradual. We speculate that the Belgrano glacier occupied the eastern portion
 490 of the valley for at least 8 ka, between ~ 25 ka and ~ 17 ka, based on the
 491 age from the inner Menelik moraine (17.0 ± 0.7 ka), but further work is
 492 needed to confirm this. Shallow proglacial lakes formed as the ice margin
 493 retreated from the inner Menelik moraines. Once the glacier retreated into
 494 the Belgrano trough, it likely experienced a more dynamic retreat associated
 495 with calving. We have no evidence to constrain the maximum recession prior
 496 to the re-advance that deposited the Belgrano moraines.

497 Although we cannot firmly constrain the timing and rate of deglaciation,
 498 our youngest age (17.0 ± 0.7 ka) from the inner Menelik moraine broadly
 499 matches the onset of deglaciation elsewhere in Patagonia. At LP, deglaciation
 500 began as ice withdrew from the youngest Río Blanco moraines dated at $\sim$
 501 21 - 19 ka (Hein et al., 2010). Rapid deglaciation of the LP and Chacabuco

valleys accompanied by ice thinning is supported by a vertical reconstruction (Boex et al., 2013) and a pollen analysis of a lake core (Henríquez et al., 2017). At LBA, the onset of deglaciation was dated at 18.1 ± 0.2 ka, followed by a short still-stand at 17.7 ± 0.1 ka, which deposited the Menucos moraines (Bendle et al., 2017a; Kaplan et al., 2004; Douglass and Bockheim, 2006). Deglaciation was initially steady, but accelerated after 17.3 ka when the LBA outlet lobe started to calve into a deep lake (Bendle et al., 2017a). Overall, it appears that in central Patagonia deglaciation was underway by 19 - 18 ka (García et al., 2019; Markgraf et al., 2007; Hein et al., 2010; Henríquez et al., 2017; Bendle et al., 2017a), while in southern Patagonia and in the CLD ice retreat initiated about 1 ka later (McCulloch et al., 2005; García et al., 2014; Kaplan et al., 2008; Sagredo et al., 2011; Denton et al., 1999a; Moreno et al., 2015; Mendelova et al., 2017).

6.6. *Causes of glaciation*

In the following we consider potential mechanisms that could lead to ice expansion during MIS 5 and the pattern of successively smaller ice extent at Lago Belgrano.

6.6.1. *Insolation*

Earth's orbital parameters exert a primary control on glacial-interglacial cycles and modulate global climate (Hays et al., 1976; Milankovitch, 1941). Insolation in the southern hemisphere is almost completely out-of-phase with the northern hemisphere, and it has been proposed that southern hemisphere insolation modulates regional climates, and therefore controls glacial advances in the southern mid-latitudes (Vandergoes et al., 2005; Sutherland

et al., 2007; Darvill et al., 2015, 2016).

A minimum in southern hemisphere summer insolation occurred between 85 - 75 ka (Fig. 7 B), and this may have promoted ice advance at Lago Belgrano. As noted by Hughes and Gibbard (2018), insolation played a particularly important role in glacial inception, including the insolation minimum at the end of the previous interglacial. The amplitude of southern hemisphere insolation change is greatest during MIS 5, with a minimum of the last glacial cycle between 115 ka and 100 ka. This would have likely created favorable conditions for glaciation of the Patagonian Andes. Glaciers in mid- and high-northern latitudes are thought to have expanded in response to reduced summer insolation early in the last glacial cycle (e.g. Abramowski et al., 2006; Porter and Swanson, 2008; Farquharson et al., 2018).

Reduced southern hemisphere summer, as well as winter, insolation have been invoked to explain glacier expansion in Patagonia and New Zealand during MIS 3 (Darvill et al., 2015, 2016; García et al., 2018). However, Doughty et al. (2015) and Darvill et al. (2016) noted that southern hemisphere insolation alone cannot explain the timing of glacier advances observed in the southern mid-latitudes. While insolation was important in driving the onset of glaciation, later in the glacial cycle it may have only played a comparatively minor role (Hughes and Gibbard, 2018). Further, variations in insolation would modulate regional climates differently (Gillespie and Molnar, 1995). The timing and extent of ice advances along the Patagonian Andes was then ultimately determined by local-to-regional factors and feedback mechanisms (cf. Darvill et al., 2016; García et al., 2018).

An alternative hypothesis suggests that the precessional cycle might af-

551 fect the timing of glacial advances in the southern mid-latitudes by control-
552 ling southern hemisphere seasonality and the pole-to-equator thermal gradi-
553 ent, which in turn drives westerly circulation (Shulmeister, 1999; Shulmeister
554 et al., 2004). Glacial advances would therefore occur at southern hemisphere
555 precessional maxima, i.e. 23 ka, 48 ka, 72 ka, and 96 ka (Shulmeister et al.,
556 2010). In this model, ice expansion at Lago Belgrano might have been driven
557 by a precessional maximum at 96 ka and at 72 ka. There is evidence for ice
558 advances in Patagonia at ~ 48 ka (García et al., 2018), and in NZ at ~ 42
559 ka (Kelley et al., 2014) that could be linked to a precessional maximum at
560 48 ka.

561 6.6.2. *Moisture availability*

562 While the initial growth of the San Lorenzo ice cap was likely driven
563 by reduced southern hemisphere insolation, we suggest the early maximum
564 and successively smaller ice extent at Lago Belgrano was, in part, related
565 to decreasing moisture delivery over the course of the glacial cycle. The
566 San Lorenzo massif is the easternmost glaciated area in Patagonia today,
567 sustained by westerly precipitation. Early in the glacial cycle, expansion
568 of an independent ice cap over San Lorenzo would have been fostered by
569 westerly precipitation that was able to penetrate through the gap in the
570 Andes. Once the full PIS developed to the west, it blocked this gap and
571 created a lee-side effect that led to a reduction in snowfall over San Lorenzo.
572 As a consequence, the Belgrano glacier was substantially smaller during the
573 gLGM.

574 Availability of moisture and the role of continentality has been highlighted
575 as an important factor contributing to an early local LGM and a pattern of

576 successively smaller extent of mountain glaciers (Hughes and Gibbard, 2018;
 577 Gillespie and Molnar, 1995). For example, maximum glaciation of the moun-
 578 tain ranges in central Asia early in the glacial cycle was attributed to higher
 579 precipitation related to the strength and advection of the South Asian mon-
 580 soon and/or westerlies (e.g. Röhringer et al., 2012; Dong et al., 2018; Zech,
 581 2012; Owen et al., 2002; Finkel et al., 2003; Owen et al., 2003; Abramowski
 582 et al., 2006). Development of large northern hemisphere ice sheets would
 583 have affected atmospheric circulation and created lee side effects and precip-
 584 itation starvation in other glaciated areas, thus indirectly contributing to a
 585 pattern of successively smaller glaciation elsewhere that was asynchronous
 586 with the global ice volume (e.g. Thackray, 2001; Porter and Swanson, 2008;
 587 Svendsen et al., 2004; Zech et al., 2011; Farquharson et al., 2018).

588 6.6.3. *Antarctic stadials*

589 On millennial time-scales, glacial advances in Patagonia and New Zealand
 590 have also been attributed to cooling in the southern mid-latitudes during
 591 Antarctic stadials (García et al., 2018; Kelley et al., 2014; Schaefer et al.,
 592 2015). An Antarctic millennial-scale climate signature across the southern
 593 mid-latitudes is likely an expression of antiphased interhemispheric coupling
 594 commonly attributed to the bipolar seesaw (Pedro et al., 2018; Stocker and
 595 Johnsen, 2003). Meridional oceanic heat transfer causes the Southern Ocean
 596 and Antarctica to cool during warm interstadial conditions in the North At-
 597 lantic region, and vice versa (Blunier, 2001; EPICA Community Members,
 598 2006). Likely forced from the North Atlantic, the oceanic teleconnections are
 599 accompanied by an atmospheric reorganization (Markle et al., 2017; Buiz-
 600 ert et al., 2018), which causes equatorward migration of the SWW during

601 Antarctic stadials (Markle et al., 2017; Lamy, 2004; Buizert et al., 2018).
602 Further, the Antarctic Circumpolar Current weakens (Lamy et al., 2015),
603 and cold water masses are advected northward resulting in lower SST off
604 Chile (Kaiser et al., 2005) and New Zealand (Barrows et al., 2007). Together
605 this could promote glacier expansion in Patagonia and New Zealand.

606 Antarctic millennial-scale climate variations have been correlated across
607 the southern mid-latitudes by a range of marine proxies (e.g. Barrows et al.,
608 2007; Lamy, 2004; Caniupán et al., 2011). Southern high- to- mid latitude
609 millennial-scale cooling evident in Antarctic ice-cores and several marine
610 records might have also facilitated the glacial advance at our site at ~ 75
611 ka. Cold stadial conditions at this time are evident in Antarctic ice cores
612 following the prolonged A7 warm event (7 F; Blunier, 2001; EPICA Commu-
613 nity Members, 2006. A Mg/Ca-based SST reconstruction from east of New
614 Zealand indicates a temperature decrease towards ~ 75 ka (45° S, Pahnke
615 et al., 2003; Fig. 7 D). Although the resolution of exposure dating and as-
616 sociated geological uncertainties do not allow for precise correlations, it is
617 reasonable to expect that mountain glaciers in the southern mid-latitudes
618 responded to such cooling events.

619 *6.6.4. Build-up of the Patagonian Ice Sheet and ice-divide migration*

620 In the light of the above discussion, we propose a conceptual model of
621 the build-up of the PIS (Fig. 8). The San Lorenzo massif and surrounding
622 high topography (and present day NPI and SPI) likely acted as nucleation
623 points for the ice sheet. Ice inception here allowed for the development of an
624 independent ice cap at the time of reduced southern hemisphere insolation
625 and higher moisture availability early in the glacial cycle. Growth of the San

626 Lorenzo ice cap would have been sustained by westerly precipitation that
 627 was able to reach the eastern massif through the gap in the Andes (Fig. 8
 628 A). Later in the glacial cycle, the individual ice masses that were building
 629 up on high topography, including the San Lorenzo massif, the NPI and the
 630 SPI, coalesced to form a continuous ice sheet (Fig. 8 B). Approaching peak
 631 glacial conditions (Fig. 8 C), we expect the ice-divide migrated westward
 632 towards the precipitation source, and in the process caused re-organization
 633 of ice-drainage routes to focus ice flow into the major outlets of the PIS that
 634 surround the San Lorenzo and southern massifs (LP and SM). Furthermore,
 635 increasing the ice-sheet elevation and migrating the ice-divide would create
 636 an effective barrier to westerly precipitation, causing a reduction in snowfall
 637 reaching the San Lorenzo massif (cf. Sugden et al., 2002). The successively
 638 smaller glacial extent of the Belgrano glacier can therefore be explained by
 639 reduced moisture on the lee side of the PIS and ice-drainage re-organization
 640 as a consequence of ice-divide migration. This explains why the Belgrano
 641 glacier was relatively small at ~ 25 ka, at a time when the major regional
 642 outlets of the PIS (LP and LBA) were at their local maximum (Hein et al.,
 643 2010, 2009; Kaplan et al., 2004).

644 7. Conclusion

645 Our ^{10}Be ages and geomorphological mapping demonstrate that the Bel-
 646 grano glacier reached its maximum extent of the last glacial cycle at ~ 75 ka
 647 towards the end of MIS 5. A second, substantially smaller advance occurred
 648 at ~ 25 ka during the gLGM. The topographic setting prevented overprinting
 649 of the MIS 5 moraine record by later advances. We suggest that an indepen-

650 dent ice cap developed over San Lorenzo early in the last glacial cycle, likely
651 in response to reduced southern hemisphere summer insolation. Fostered by
652 relatively high westerly precipitation, this led to the most extensive advance
653 of the Belgrano glacier early in the last glacial cycle. Successively smaller
654 ice extent at Lago Belgrano can be explained by snow starvation, ice-divide
655 migration and ice-flow re-organization as the full PIS developed to the west
656 of San Lorenzo. Our data provide first robust evidence for glaciation of the
657 Patagonian Andes during MIS 5. However, more work is needed to elucidate
658 the drivers of glaciation in Patagonia, from the onset early in the glacial cycle
659 through to the full glacial conditions.

660 **Acknowledgements**

661 This work was supported by a NERC PhD studentship (NE/L002558/1)
662 to MM. ^{10}Be analysis was supported by NERC CIAF grant (9167/0416).
663 The British Society for Geomorphology, Quaternary Research Association,
664 Royal Geographical Society (with IBG) and Explorers Club supported the
665 2018 field campaign. We also acknowledge funding from the Scottish Al-
666 liance for Geoscience, Environment and Society (Small Grants Scheme), and
667 the Moray Endowment Fund from the University of Edinburgh. We are
668 grateful to Andrew Gray for his help with fieldwork and Patricio Parada for
669 arranging export of samples. The national park authorities of Argentina are
670 thanked for granting permission to conduct fieldwork, and the rangers of the
671 Perito Moreno National Park are thanked in particular for their assistance
672 in the field. We would like to thank D. Sugden and R. McCulloch for in-
673 sightful discussions. We are grateful for the constructive comments from five

674 anonymous reviewers that greatly improved this manuscript.

Figure 1

Figure 1: A) Location of Patagonia in the southern hemisphere along with Southern Ocean and Antarctic proxies discussed in the text and approximate location of the southern westerlies (SWW). B) Location of our study site in Patagonia and approximate extent of the Patagonian Ice Sheet during the gLGM (Singer et al., 2004). Image: ESRI World Imagery. (LBA: Lago Buenos Aires, TdP: Torres del Paine, UE: Última Esperanza, BI-SS: Bahí Inútil-San Sebastián lobe). C) The topography of central Patagonia including the tectonic gap that breaks the main chain of the Andes to separate the North Patagonian Icefield (NPI) from the South Patagonian Icefield (SPI), and the San Lorenzo massif (SL) from the Pacific Ocean (not shown). The Lago Belgrano study area is indicated. Blue arrows depict drainage directions described in the text, blue dotted line represents hydrological drainage divide. (LP: Lago Puyeredón, SM: Lago San Martín). The hillshade basemap was derived from the Shuttle Radar Topography Mission (SRTM) Digital Elevation Model (v3, 30m). Glacier outlines are from the GLIMS database (Falaschi et al., 2017; De Angelis et al., 2015).

Figure 2: Geomorphological map showing the main moraine systems in the Belgrano valley, informally named the Klementek, Menelik and Belgrano moraines from the oldest to youngest. Sample locations are also shown. The hillshade basemap was derived from the Shuttle Radar Topography Mission (SRTM) Digital Elevation Model (v3, 30m).

Figure 3: A) Satellite image (Sentinel-2) and B) mapped moraine ridges of the Klementek moraine system and the associated hummocky terrain. Sample locations and ^{10}Be exposure ages are also shown. C) Photograph showing part of the hummocky terrain, with a vehicle in the bottom left corner for scale. D) Photograph showing the outermost Klementek moraine ridges. E) Photograph showing the outermost Klementek moraine ridges and the associated outwash in foreground that was sampled for surface exposure analysis. F) An example of an outwash cobble sampled, which still exhibits fluvial shape, but some surface erosion is evident (maximum 3 mm).

Figure 4: A) Photograph showing the Menelik moraines (dashed yellow lines) draped on higher ground. B-E) Photographs showing the outermost Menelik moraines (dashed yellow lines) along with examples of sampled boulders. F) Example of a cobble sampled from the outwash associated with Menelik moraines. Ventifacts on the surface suggest long surface exposure (cf. Hein et al., 2011).

Figure 5: Enlarged geomorphological map showing the Menelik moraine system along with sample locations and ^{10}Be exposure ages.

Figure 6: Probability density diagrams ('camelplots') for A) the Klementek cobble ages, A) Menelik boulder ages, and C) Menelik cobble ages. Thin purple curves represent Gaussian curves for individual samples. Thick purple curve represent the summed probability distribution for the data, excluding outliers. Outliers are represented by dashed curves. Vertical lines denote an arithmetic mean (dark green, excluding outliers), and 1 and 2 σ (paired brown and yellow lines respectively) confidence intervals of the mean. Wmean: uncertainty-weighted mean.

Figure 7

Figure 7: Comparison of our ^{10}Be ages and other proxies. A: July insolation intensity for 60° N (Berger and Loutre, 1991). B: July (orange) and December (blue) insolation intensity for 50° S (Berger and Loutre, 1991). C: EDC dust record (Lambert et al., 2012). D: SST (Mg/CA) record from off New Zealand (MD97-2120) (Pahnke et al., 2003). E: EDC deuterium record (EPICA Community Members, 2004) plotted on AICC2012 timescale (Veres et al., 2013). F: Oxygen isotope record from Byrd ice core (Johnsen et al., 1972) plotted on GISP2 timescale (Blunier, 2001). G: Summed probability density functions ('camelplots') for the Klementek outwash, and the Menelik boulders and outwash cobbles combined (plotted on individual y-axis; dashed curves represent outliers). Circles indicate the arithmetic means of the Klementek and Menelik ages (after outliers have been excluded) and error bars indicate uncertainty of the mean, which includes propagated individual internal uncertainties of the ^{10}Be ages and production rate uncertainty (Kaplan et al., 2011).

Figure 8

Figure 8: A conceptual model of the build-up of the last Patagonian Ice Sheet (PIS). A) The initial inception of the Patagonian Ice Sheet begins with nucleation centres over the present day North and South Patagonian Icefields (NPI, SPI), and the San Lorenzo massif (SL), likely triggered by reduced summer insolation in the southern hemisphere. At this time, the Southern Westerly Winds (SWW) can freely penetrate through the topographic depression between the present icefields to bring snowfall east to the San Lorenzo ice cap. Ice expansion is greatest down the high-elevation eastern Belgrano valley, since glaciers draining to the south, west and north of the massif descended into deep valleys. B) The individual ice masses coalesce, causing reorganization of ice (and fluvial) drainage routes, and reducing ice flux down the Belgrano valley. C) Under peak glacial conditions, the ice-divide increased in elevation, and migrated westward towards the precipitation source. This caused snow starvation to the San Lorenzo massif, and significant re-organization of ice-drainage routes to focus ice flow into the major outlets of the Patagonian Ice Sheet that surround the San Lorenzo massif (LP and SM). These processes limited ice expansion in the Belgrano valley when the major outlet to its north, LP (and also LBA, which is not shown) was near its maximum last glacial extent indicated by the Río Blanco moraines (Hein et al., 2010).

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