Rapid glacier recession at Monte San Lorenzo (Patagonia) in response to abrupt Southern Hemisphere warming 13.0–12.0 ka BP

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ABSTRACT: Mid-latitude Patagonian glaciers are sensitive to changes in the complex coupled ocean-atmospheric climate system of the Southern Hemisphere. Here, we investigate glacier response to a period of rapid climate warming immediately post-dating the Antarctic Cold Reversal (ACR). We analyse a sequence of 13 ice margins, four of them dated, from an outlet valley of the Monte San Lorenzo ice cap (47.3°S). We constrain glacier recession of 31.7 km over a period of ~1 kyr from 13.2 ± 0.4 to 12.1 ± 0.4 ka. The average rate of recession was 35.2 m a⁻¹ over this 1-kyr period, increasing from 12.75 m a⁻¹ at the start of the record (moraines M1 to M4) to 50 m a⁻¹ from M9 to M12. This recession occurred during a period of rapid warming when the austral westerlies shifted polewards. It is likely that ice extent stabilized during the Holocene, with ice occupying the M13 moraine during repeated Holocene neoglacial, before anthropogenic warming caused recession at a rate of 55.5 m a⁻¹ from 1985 to 2016. We conclude that 20th century rates of recession were higher than those during post-ACR warming, though of a similar order of magnitude.

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KEYWORDS: Antarctic Cold Reversal; cosmogenic nuclide surface exposure dating; glacial geomorphology; Late Pleistocene; Southern Hemisphere palaeoclimate.

Introduction

Across central Patagonia there is now an improved understanding of the timing and extent of major Quaternary ice limits of the former Patagonia Ice Sheet (Davies et al., 2020). For example, mapping and dating has determined ice limits for the Great Patagonian Glaciation at 1.1 Ma (Hein et al., 2009, 2011; Tobal et al., 2021), the Last Glacial Maximum (Hein et al., 2010; Garcia et al., 2018), the Antarctic Cold Reversal (ACR) (Saggedo et al., 2018; Davies et al., 2018; Mendelova et al., 2020) and Holocene neoglacial (Nimick et al., 2016; Kaplan et al., 2016; Reynhout et al., 2019; 2020; Sagredo et al., 2021). Rates of glacier recession during the Last Glacial Interglacial Transition (LGIT), defined as 15–11.7 ka (Lowe and Hoek, 2001) can be difficult to constrain in Patagonia as most outlet glaciers terminated in large proglacial lacustrine environments or because mountain glaciers remain inaccessible, both limiting landform reconstructions. Bendle et al. (2017a) used thinning of annually laminated lake sediments at Lago Buenos Aires (Argentina) to precisely constrain an acceleration in glacier recession at the start of deglaciation (18.0–17.0ka), but ice margin position during the later centuries of this record was inferred as the ice-lobe had already receded into deep lake waters.

Quantifying rates of glacier recession during warming phases of the LGIT is an important goal to contextualize modern rates of glacier change in Patagonia (Dussaillant et al., 2019), and to understand how present-day rates of recession compare to the palaeo-record. Furthermore, these data can be used to test temperature and precipitation drivers of glacier response to climate change and provide empirical data to underpin glacier modelling efforts. One palaeo-warming phase during the LGIT in Patagonia followed the ACR, a Southern Hemisphere cooling event from 14.7 to 13.0 ka that influenced Patagonian climate as far north as 40°S (Pedro et al., 2016). A rapid temperature rise of 2.5°C during an ~1-kyr period following the ACR is recorded in the WAIS Divide ice cores in Antarctica (Friedel et al., 2015; Cuffey et al., 2016), while sea surface temperatures in the Pacific Ocean at 46°S also show rapid warming at this time (Haddam et al., 2018). Lago Cardiel (48°S, 71°W) reached its hydrological maximum (136% of modern) by 11.3 ka (Quade and Kaplan, 2017) because a southerly (poleward) shift in the austral westerlies allowed easterly moisture sources to reach the lake (Garreau et al., 2013). ACR ice limits have been identified in outlet glaciers of the main Patagonian icefields (Thorndycraft et al., 2019), and around smaller satellite ice caps such as Monte San Lorenzo (MSL) (Davies et al., 2018; Sagredo et al., 2018; Mendelova et al., 2020). MSL (Fig. 1) is potentially more sensitive to both temperature and precipitation changes than glaciers of the main icefields because of its smaller size and location in the rain shadow to the east of the main Andean Cordillera. The valley floors here are located altitudinally above the regional proglacial lake systems that developed during deglaciation (Thorndycraft et al., 2019), allowing greater preservation potential for ice-marginal landforms. Recent geomorphological mapping and morpho-sedimentological analyses along the Río del Salto...
and Pedregoso valleys (Fig. 1) identified a total of 13 ice margins (Martin et al., 2019), extending from an ACR ice limit down valley (Glasser et al., 2012; Davies et al., 2018) back to the early 20th century position of MSL ice. The landform record here therefore provides the potential to constrain rates of glacier recession during a sustained phase of ablation.

The aim of this study is to present new cosmogenic nuclide surface exposure ages in the Pedregoso and Salto valleys (Patagonia), the pathway of Glaciar Calluqueo. These ages, alongside previously published ages from Glasser et al. (2012) and Davies et al. (2018), are used to date moraines formed during the recession of Glaciar Calluqueo, an important outlet glacier of the MSL ice cap (Fig. 1). We use Bayesian age modelling to robustly evaluate the timing of moraine formation and subsequently quantify rates of glacier recession over a period of ~1000 years, post-dating the ACR. Finally, we evaluate the landform record within the context of Patagonian palaeoclimate and compare the Glaciar Calluqueo record with modern rates of glacier recession.

**Study area**

MSL, centred on 47°35′S, 72°18′W (Fig. 1), is an isolated granodioritic to granitic massif (Ramos et al., 1982), located 70 km east of the main Andean Cordillera. It experiences a temperate climate with mean annual air temperature of 8.4°C and average annual precipitation of 750 mm w.e. (Dirección Meteorologica de Chile, 2001) recorded at the nearest meteorological station (47°14′S, 72°33′W; 182 m asl) in the town of Cochrane (Fig. 1). MSL presently supports a small ice cap (207 km²) with the largest outlet glacier, Calluqueo, descending to 520 m asl (Falaschi et al., 2013). There is asymmetry in the snowline attributed to regional precipitation gradients:

![Figure 1. Multidirectional hillshade digital terrain model showing the location of the Monte San Lorenzo (MSL) massif near Cochrane (Chile), Glaciar Calluqueo and Barrancos mountains. Contemporary glacier outlines from the Randolph Glacier Inventory (RGI Consortium, 2017) are shown in white. Also shown are the Pedregoso, Tranquilo and Salto valleys discussed in the text. Inset: the location of MSL in southernmost South America.](image-url)
1700–1750 m asl in the wetter western sectors and 1800 m asl in the drier eastern sectors (Falaschi et al., 2013).

At the local Last Glacial Maximum (LGM), ice from glaciars on the western and northern flanks of MSL discharges into the Salto and Tranquilo valleys, coalescing with the Cochrane-Pueryredón outlet lobe from the eastern flank of the terminal moraine region of Glaciar Colonia (NPI) (Wenzens, 2002). At the LGM, the Cochrane-Pueryredón outlet lobe reached the Argentinian lowlands forming large moraine sequences (Caldenius, 1932; Mercer, 1976; Hein et al., 2010; Bendle et al., 2017b). Recession of ice occurred from ~18 ka (Hein et al., 2010; Bendle et al., 2017a; Davies et al., 2020), with ice receding back to the study area shown in Fig. 1. During this time, large ice-dammed palaeolakes formed (Bell, 2008; Turner et al., 2005; Glasser et al., 2016), with drainage controlled by the recession of glacier termini past cols. The largest of these palaeolakes, Palaeolake Chelenko, was extinct from 14.2 to 12.6 cal ka BP (Thorndycraft et al., 2019). Following drainage of Palaeolake Chelenko and subsequent catastrophic flooding (Benito and Thorndycraft, 2020), the present-day extent of Lago Cochrane/Pueryredón was established (Thorndycraft et al., 2019).

During the ACR readvance, ice discharging from MSL and the eastern Barrancos Mountains (Fig. 1) terminated in Palaeolake Chelenko at the glaciolacustrine Lago Esmeralda moraine, dated to 13.4 ± 0.2 ka (Davies et al., 2018), which is numbered M1a (Figs. 2 and 3a) according to subsequent mapping in the Salto and Pedregoso valley (Martin et al., 2019). Post-ACR recession resulted in a sequence of 12 further moraines, M2–M13 (Martin et al., 2019), the numbering system used in this paper (Fig. 2). Davies et al. (2018) and Martin et al. (2019) recalculated the ‘Moraine Mounds’ detailed by Glasser et al. (2012) to 12.7 ± 0.4 ka and reinterpreted their origin as the subaerial M4 moraine (Fig. 3a) of the Salto valley sequence (Fig. 2). The minimum age for ice-free conditions in the Salto valley is provided by identification of the H1 tephra (~8.2 ka) identified at various sites (Fig. 2) in the Salto and Tranquilo valleys (Gardeweg and Sellés, 2013; Stern et al., 2016). Herein, we present new geochronological data from moraines M9 and M12 in the Salto and Pedregoso valleys respectively (Figs. 1 and 2), to date ice-margin positions during post-ACR recession.

Methods

Cosmogenic nuclide surface exposure dating

Cosmogenic nuclide surface exposure dating of glacially transported boulders provides an age at which the boulder was deposited at the ice front through the measurement of 10Be accumulated in the rock’s upper surface (Balco, 2011; Gosse and Phillips, 2001). Measuring the exposure to cosmic radiation of multiple boulder surfaces found on a moraine ridge crest provides an age of moraine formation, giving a chronologically constrained position of the frontal ice margin (e.g. Shulmeister et al., 2005; Putnam et al., 2013; Sagredo et al., 2018).

Two moraines, M9 and M12 (Fig. 2), from the Salto-Pedregoso sequence mapped by Martin et al. (2019), were targeted for cosmogenic surface exposure dating. M9 is a terminal moraine located on the Salto valley floor, downstream of the confluence of the Pedregoso and Tranquilo valleys, and 11.5 km up-valley from the M4 moraine dated by Glasser et al. (2012). M12 is a lateral moraine complex on the right valley margin of the Pedregoso valley. Its inferred valley floor terminus position is located ~15 km up-valley from M9 and 4.5 km down-valley from the M13 moraine (Fig. 2), the last in the sequence (Martin et al., 2019).

Five granitic boulders were sampled from each moraine, with one sample taken from each boulder. Large granitic boulders with a b-axis >1 m (Heyman et al., 2011) and abundant quartz grains were sampled (see Fig. 3 for examples). Boulders located on moraine ridge crests and showing signs of subglacial transport were selected, for example faceted, striated and sub-rounded boulders (cf. Cockburn and Summerfield, 2004; Gosse and Phillips, 2001; Darvill, 2013). Boulders showing significant weathering or flaking on the surface, or which may have rolled or slumped post-deposition, were avoided. Samples greater than 1 kg in weight (to ensure sufficient quartz content) were taken from the upper 5 cm of the flat top surface of quartz-rich boulders using a hammer and chisel. Boulder surface dip and strike and the elevation of the horizon were measured to calculate shielding correction factors.

Following sample collection, the whole-rock samples were mechanically crushed and sieved to 250–500 µm. Sample processing and analysis was done at the National Environment Research Council (NERC) Cosmogenic Isotope Analysis Facility (CIAF) at the Scottish Universities Environmental Research Centre (SUERC). Quartz was isolated by froth flotation and then magnetic mineral separation using a Frantz® iso-dynamic mineral separator, before being treated by hexafluorosilicic acid to remove non-quartz minerals and undergoing repeated etching in 2% hydrofluoric acid (HF) to remove at least 25% of the quartz mass. The purity of the obtained quartz cores was assayed by ICP-OES (inductively coupled plasma optical emission spectrometry). Between 11 and 19 g of purified quartz per sample was dissolved in 40% HF with between 0.29 and 0.30 g of the CIAF-PHF in-house 10Be carrier solution ([Be] = 849 ± 12 ppm) and Be was chemically isolated following the methods developed by Child et al. (2000). The solution was precipitated as Be(OH)2, baked to BeO in a quartz crucible before being mixed with Nb (BeO:Nb ratio of 1:6) to prepare the BeO–Nb targets. The 10Be/9Be ratios were measured by the 5-MV NEC Pelletron accelerator mass spectrometer (AMS) at SUERC (Xu et al., 2010), normalized to the NIST SRM4325 standard, 10Be/9Be ratio 2.79 × 10^-11 (Nishizumi et al., 2007). The processed blank ratio (5.6 ± 0.9 × 10^-15 10Be/9Be) ranged between 6 and 10% of the sample 10Be/9Be ratios (see Supporting Information Table S1). The carrier used (CIAF-PHF) contains a 10Be/9Be ratio of 1.5 × 10^-15, that was subtracted from all measured ratios. The remaining 10Be content in the blank was considered as contamination and subtracted from the total 10Be content in the unknown samples. The uncertainty of this correction is included in the stated standard uncertainties (Table 1).

Surface exposure ages were calculated using the CRONUS-Earth online calculator version 3 (hess.ess.washington.edu; Balco et al., 2008), using the regional Patagonian spallation production rate from the Kaplan et al. (2011) calibration data set (calibration.ice-d.org/pubs/28), and assuming an erosion rate of 0 mm ka^-1 to obtain a minimum age and using the LSDm scaling scheme (Lifton et al., 2014). Systematic production rate uncertainties are included in the external age uncertainties. Ages were also calculated using the St and Lm scaling schemes for comparison (Table 1). No corrections were made for snow cover or isostatic uplift. An age for a given moraine is derived from the surface exposure ages of multiple boulders sampled on the moraine, presented as an Uncertainty Weighted Mean (UWM) and weighted standard deviation, weighted by the inverse of the sample’s internal age uncertainty (cf. Jones et al., 2019). Further discussion of the use of UWM is presented in the
Results section as it is based on our assessment of analytical uncertainty and the Bayesian age modelling.

**Bayesian age modelling**

Applying a sequence of dated moraines to a Bayesian model can robustly constrain the chronology of ice recession (e.g. Chiverrell et al., 2013; Jones et al., 2015; Smedley et al., 2017). A uniform phase sequence model (Bronk Ramsey, 2008) was utilized in Oxcal v4.3 (c14.arch.ox.ac.uk) to produce probability density estimates for each surface exposure age sample, utilizing phase and boundary functions to identify and separate the samples by moraine site. The General outlier model (Bronk Ramsey, 2009), was applied enabling outliers to be identified and retained in the model, but their influence was proportionately scaled down. Samples were further assessed for their fit within the wider dataset based upon the agreement index, with a minimum recommended value of 60% (Bronk Ramsey, 2009).

Figure 2. Sampled boulder locations and ages (with internal uncertainties) for the M9 and M12 moraines (inset panels). An asterisk denotes an outlier boulder according to our analysis (see text). Note there was insufficient quartz from sample PMCL5. Main map shows locations of the M1–M13 moraines of the Salto and Pedregoso valleys, as well as the previously published ages from Glasser et al. (2012) and Davies et al. (2018) used in our Bayesian age model. Also shown are the sample locations and ages from Glacier Tranquilo (Sagredo et al., 2018). Note the M1c moraine ages are younger than M1a because the sampled boulders were shielded beneath Chelenko lake waters.
Figure 3. Photomontage of study area and selected boulders sampled for cosmogenic nuclide exposure dating. (a) View from the lateral M1a moraine towards the terminal Antarctic Cold Reversal moraine (Davies et al., 2018). The position of the M4 moraine, dated by Glasser et al. (2012), is indicated but is hidden from view. (b) Up-valley view, from the M1a moraine, of the Río del Salto showing the position of the M7 moraine. (c) View of Monte San Lorenzo (MSL) from the lateral M11 moraine complex. (d) View from the M11 moraine complex of the M13 ice-margin and Lago Calluqueo. (e, f) Two views of sampled boulder SVM3 located on the crest of moraine M9. (g) View of sampled boulder PMCL1 from the M12 lateral moraine complex. (h) View of sampled boulder PMCL2 from the M12 lateral moraine complex.
Here we present nine new ages for moraines M9 and M12 located in the Salto and Pedregoso valleys respectively (Table 1 and Fig. 3). M9 ages range from 14.9 ± 1.3 to 11.2 ± 1.0 ka. Normal kernel density estimate plots (Fig. 4), plotted with internal uncertainties, show a spread in ages from the M9 moraine. SVM2 dated to 14.9 ka is too old stratigraphically as it is older than the down-valley ACR moraines (Davies et al., 2018) and is therefore identified as an outlier. Furthermore, a reduced chi-squared test can be used to statistically indicate which ages form a consistent age population for the landform. A reduced chi-squared test value $\chi^2_r$ less than the 2σ error in the age internal uncertainty. The black curve represents the sum of the kernels. The dashed curves of M9 and M12 represent the sum of the kernels after outliers SVM2, SVM5 and PMCL3 have been removed. ACR, Antarctic Cold Reversal.

**Results**

**Exposure age chronology**

The M12 moraine complex marks the lateral ice margin position of Glaciar Calluqueo in the Pedregoso valley (Fig. 3), giving a terminal position, inferred by extrapolating the M12 lateral moraine along the valley side at a constant gradient, south of the Salto-Tranquilo river confluence. M12 ages range from 19.0 ± 1.6 to 11.7 ± 1.0 ka (Table 1) and the normal kernel density plots (Fig. 4) demonstrate that PMCL3 is an outlier, older than the ACR ice limits down valley. Applying the reduced chi-squared test to the remaining ages

![Figure 4. Normal kernel density estimates of published and new (this study) cosmogenic nuclide surface exposure ages from the Pedregoso and Salto valleys. Light grey Gaussian kernels represent a single sample age with the width equal to the 1σ error in the age internal uncertainty. The black curve represents the sum of the kernels. The dashed curves of M9 and M12 represent the sum of the kernels after outliers SVM2, SVM5 and PMCL3 have been removed. ACR, Antarctic Cold Reversal.](image-url)
indicates the remaining samples are probably representative of a single population, and subsequently a UWM of $12.1 \pm 0.4$ ka is calculated, stratigraphically consistent with the younger moraine. Ages calculated using the LSNn (Table 1) St and Lm scaling schemes (Supporting Information Table S2) overlap within their internal uncertainties, and therefore have a negligible impact on later calculations of glacier recession rate.

Bayesian age modelling

In this section we combine our new exposure ages with the previously published dates of Davies et al. (2018) and Glasser et al. (2012) for the M1 and M4 moraines respectively and apply Bayesian age modelling to provide a robust evaluation of the chronology. This is because the M4, M9 and M12 ages of $12.7 \pm 0.4$, $12.4 \pm 0.4$ and $12.1 \pm 0.4$ ka respectively overlap statistically, but there is a clear stratigraphic order of mapped moraines allowing the application of a sequence age model. Furthermore, the age modelling allows a statistical determination of the probable start and end dates for moraine formation phases (Table 1). Two Bayesian age model set-ups were run, with Model 1 using the General Outlier Model to further test outlier samples SVM2 and SVM5. PMLC3 was not included in the model but is presented in Fig. 5 for context. Agreement values of 27.6% and 57.6% for SVM2 and SVM5 respectively were below the recommended acceptable 60% threshold (Bronk Ramsey, 2009) confirming they can be considered outliers.

Model 2 was run with the SVM2 and SVM5 ages excluded. Table 2 shows the $2\sigma$ age range produced for each moraine phase, alongside the UWM moraines ages obtained from surface exposure dating, while Fig. 5 shows the output from Model 2. The posterior density estimate ($2\sigma$ age range) generated by the model is shown in dark grey on Fig. 5 and the relative probability of each sample age within the range of the external uncertainty is in light grey.

The $2\sigma$ age range for the end of the ACR M1 moraine phase was modelled to $13.2 \pm 0.4$ ka, which statistically constrains the start of glacier recession to the end of the ACR in Patagonia at $\sim 13.0$ ka (Pedro et al., 2016). Recession to moraine M4 occurred by $12.8 \pm 0.4$ ka, to M9 by $12.4 \pm 0.4$ ka and to M12 by $12.0 \pm 0.6$ ka (Table 1, Fig. 5). Assuming ice receded from the ACR M1 moraine at the end of the ACR at $\sim 13.0$ ka, then the period spanning the post-ACR moraine record M2–M12 in the Pedregoso/ Salto valley probably spans 400–1600 years, based on upper and lower $2\sigma$ ages of 12.6 and 11.4 ka respectively for the modelled M12 age. The modelled ages also confirm that the UWM ages are appropriate to define the timing of moraines.

Discussion

Glacier reconstruction

We present a reconstruction of the northern sector of the MSL in Fig. 6, starting from its ACR extent (Fig. 6A). At this time MSL ice was merged with glaciers from the Barrancos mountains (Fig. 1), with the smaller Glaciar Tranquilo ice limit altitudinally higher at the RT1 to RT4 latero-terminal moraines (Sagredo et al., 2018). We infer that Calluqueo ice was also advanced to the M1b limit at the ACR given the moraine’s position as the furthest advanced moraine set in the Tranquilo valley. The dates we use for the ACR limit are those of the M1a moraine ($13.8$–$13.2$ ka) because these were located above Palaeolake Chelenko (Thorndycraft et al., 2019), while ages from the M1c moraine are younger because the sampled boulders were located below lake water and date palaeolake drainage (Davies et al., 2018). Ice receded from the M1a moraine at a modelled age of $13.2 \pm 0.3$ ka to the M4 position by $12.9 \pm 0.3$ ka (Fig. 6B). An ice-marginal fan on the western side of Lago Esmeralda and palaeoshorelines (Martin et al., 2019) indicate that ice was discharging into Palaeolake Chelenko at this time. By $12.5$ ka $\pm 0.3$ ka, ice had receded up the Salto valley, back to the M9 moraine (Fig. 6C), by which time Palaeolake Chelenko had drained, an event that released $\sim 300$ km$^3$ of freshwater to the Pacific Ocean (Thorndycraft et al., 2019). We infer that ice in the Tranquilo valley had receded back towards the Salto confluence at this time, forming the ice-dammed palaeolake Lago Tranquilo at $520$ masl, the lake waters draining eastwards into Lago Brown. During subsequent recession and thinning of Glaciar Calluqueo, palaeolake Tranquilo drained to a stand at $425$ masl, with outflow down the Salto valley (Fig. 6C). Further recession to the M12 moraines occurred by $12.1 \pm 0.4$ ka (Fig. 6D, E), providing a minimum age for a glacial lake outburst flood from palaeolake Lago Tranquilo (Martin et al., 2019). The flood formed boulder bars in a reach of the lower Salto valley that had previously been submerged by Lago Chelenko (Martin et al., 2019), an event constrained to $12.4$–$11.8$ ka by age modelling (Thorndycraft et al., 2019), providing a minimum age for drainage of palaeolake Lago Tranquilo. The drainage of palaeolake Lago Tranquilo established the present-day Río Tranquilo drainage into the Salto Valley (Fig. 6E). Glaciar Tranquilo at $12.2 \pm 0.4$ ka was located at the RT5 moraine (Sagredo et al., 2018).

We have no landform or geochronological data for the Early Holocene but we infer ice was stabilized, or readvanced to, the M13 moraine complex in the mid-Holocene ($\sim 5.6$ ka, Fig. 6F), based on the dating evidence from the RT6 moraine in the Tranquilo valley (Sagredo et al., 2017, 2018, 2021). Glaciar Calluqueo remained stable at M13 during the Late Holocene or may have undergone recession/readvance during this time, based on multiple moraine crests. The glacier again stabilized at the M13 position during the latest Holocene neoglacial, dated to 0.3–0.15 ka (Davies et al., 2020), as evidenced by well-defined trimlines. Aerial photography shows that Glaciar Calluqueo began receding to its present-day margin after 1945 CE, leading to the formation of Lago Calluqueo, dammed by the M13 moraine (Fig. 1).

Recession rates and palaeoclimate significance

Our glacier reconstruction allows us to quantify post-ACR ice recession rates for Glaciar Calluqueo. In Table 3 we present a range of data on average recession rates based on the minimum and maximum modelled ages for a sequence model for the M1 to M12 moraine record (Fig. 5). The rates are considered an average for the distances between dated moraines because time was needed for intermediary (undated) moraine formation during years of glacier stillstand/advance, perhaps due to decadal-scale climate variability during the overall pattern of recession. The maximum modelled time frame for the moraine record is 1900 years (Table 3), the time between the oldest modelled age for the end of the M1a moraine phase and the youngest age for M12 formation. This time span results in an average recession rate of $16.6$ m a$^{-1}$ for recession from the M1 to M12 moraines. The minimum modelled time span is 100 years resulting in an average recession rate of $317$ m a$^{-1}$ but this can be considered unrealistic given time is needed to build the M2 to M12 moraines.

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Figure 5. Bayesian age model for the timing of moraine formation in the Salto and Pedregoso valleys based on ages and external errors. SVM2, SVM5 and PMCL3 are not included in the model but are plotted here for context. Light grey shows the relative probability of each age estimate and with the probability of each age estimate and with the posterior density estimate (2σ age range) generated by the model in dark grey.
Given the Bayesian age model (Fig. 5) provides robust statistical support for the UWM ages (Fig. 4) we use those ages to calculate average recession rates for each time step in the record (Table 3). Following the ACR, ice receded at an average rate of 12.75 m a\(^{-1}\) from M1a to M4 (Fig. 7), where landform evidence (Martin et al., 2019) suggests the frontal margin stabilized, herein dated to a UWM age of 12.9 ± 0.3 ka (Table 1). Ice then receded more rapidly up the Salto valley to M9 (29 m a\(^{-1}\)) before accelerating further to M12 at an average rate of 50.3 m a\(^{-1}\). The absence of spatial and temporal constraint of a frontal ice margin between M12 (~12.2 ka) and the M13 moraine complex, or within the M13 complex, means we cannot calculate rates of change through the Holocene. Calculating recession rates within the Holocene is challenging, because comparison with landform records elsewhere across Patagonia suggests that ice probably occupied the M13 position during one or more Holocene neoglacial, such as those at 4.0–6.0, 1.0–2.0 and 0.15–0.3 ka (cf. Davies et al., 2020).

We can use our data to compare the recession rate for Glaciar Calluqueo during post-ACR warming with its recent response to anthropogenic climate change. Regional dating of the large moraines surrounding glacier termini (Garibotti and Villalba, 2017; Davies et al., 2020) suggests that the M13 ice position was relatively stable from the Late Holocene neoglacial until the mid-20th century (Davies et al., 2020). From 1945 ice receded from the M13 moraine forming the progressively larger proglacial Lago Calluqueo (Fig. 3). From 1945 to its 2016 limit (Davies and Glasser, 2012; RGI Consortium, 2017) ice receded at an average rate of 55.5 m a\(^{-1}\) (Fig. 7). These rates therefore contextualize the average rates of 35.2 m a\(^{-1}\) calculated for recession from M1 to M12 during ~12.0–11.0 ka (Table 1). For Glaciar Calluqueo, we therefore conclude that rates of recession during post-ACR warming were lower than modern rates, though of a similar order of magnitude. In terms of ice dynamics, the post-ACR warming phase would probably contrast with the modern phase as the snout was at lower altitudes and there were additional accumulation areas, such as in the Barrancos mountains (Fig. 1). Separation of Calluqueo ice from other accumulation sources may have contributed to the faster rates of recession calculated for the M9 to M12 recession. Similar types of geomorphic setting may occur in modern glaciated terrains (Barr and Lovell, 2014).

To set the Calluqueo record into a spatial context with other glaciers in central Patagonia, we compare Glaciar Calluqueo recession during 13.0–12.0 ka with those from other ice-lakes in central Patagonia with well-constrained Late Quaternary ice margins. Glaciar Tranquilo on the northern flanks of MSL contrasts with Calluqueo in that it is a smaller glacier and its ACR ice extent was much less advanced (Fig. 3). Glaciar Tranquilo built four moraines (RT1–RT4), which are statistically indistinguishable in age, during the ACR (Sagredo et al., 2018). Ice then receded ~0.5 km to the RT5 moraine by 12.0 ka, at an average rate of ~0.5 m a\(^{-1}\), assuming a period of 1000 years from the end of the ACR. The altitudinally lower snout position of Calluqueo at the ACR, and the probable separation of multiple ice sources during post-ACR warming can account for snout recession rates an order of magnitude larger than Glaciar Tranquilo.

The Buenos Aires ice-lobe of the NPI receded ~150 km during a 3.5-kyr warming phase preceding the ACR. Average rates of surface temperature warming in Antarctica were ~0.27°C per 100 years during 18.0–14.5 ka, compared to ~0.2°C per 100 years over 13.0–12.0 ka (Fig. 8). Lake-terminating ice receded ~150 km from 18.0 ka (Bendle et al., 2017a) before stabilizing at a bedrock pinning point at Lago Bertrand moraines 15.1 ± 0.7 ka (Davies et al., 2018), constraining an average recession rate of 41–68 m a\(^{-1}\). A mid-Holocene terminus of Glaciar Colonía (NPI) was dated to 4.96 ± 0.21 ka (Nimick et al., 2016), with this lake-terminating glacier receding ~12 km to its present position at an average recession rate of 2.4 m a\(^{-1}\) over this 5-kyr interval. These data probably demonstrate, therefore, the spatial and temporal complexity of glacier response to past climate changes depending on the volume and elevation of source ice fields/ice caps, topographic controls, temperature and the strong west to east precipitation gradient.

To assess the influence of precipitation on the Calluqueo record we compare our data with Quade and Kaplan’s (2017) reconstruction of the shifting mean latitudinal position of the austral westerlies (Fig. 8), because these winds control the location of frontal systems and therefore regional precipitation (Garreau et al., 2013). Despite weaker chronological control for the Late Pleistocene, Quade and Kaplan’s data suggest that after a relatively stable position of the austral westerlies during the ACR, they shifted polewards during post-ACR warming, with MSL located at the northerly one-sigma latitudinal limit of the wind belt (Fig. 8). We therefore hypothesize that Glaciar Calluqueo was also affected by reduced precipitation in addition to increasing temperatures in the millennium following the ACR, although the shift of the austral westerlies also influences regional temperature (Garreau et al., 2013).

Our data from Monte San Lorenzo can also be compared with moraine records from the mid-latitudes in New Zealand. Following a phase of glacier growth during the ACR (Putnam et al., 2010), both Kaplan et al. (2010) and Koffman et al. (2017) reported glacier recession at 44°S in the Southern Alps from ~13.0 to 11.5 ka (i.e. during the Younger Dryas interval). Our data therefore support the evidence for a pan-hemispheric phase of ACR glacier expansion (e.g. Putnam et al., 2010; Davies et al., 2018) and subsequent recession during the Younger Dryas interval. This phasing extended from Antarctica

| Event | Model 1 | Model 2 |
|-------|---------|---------|
|       | Range (ka) | μ ± 2σ (ka) | UWM ± 1σ | Range (ka) | μ ± 2σ (ka) | UWM ± 1σ |
| M1a   | 14.5–12.5 | 13.4 ± 0.5 | 13.3 ± 0.3 | 14.6–12.4 | 13.5 ± 0.5 | 13.3 ± 0.3 |
| (start) | 14.1–12.4 | 13.2 ± 0.4 | | 14.1–12.4 | 13.2 ± 0.4 | |
| (end)  | 13.8–12.3 | 13.0 ± 0.4 | 12.9 ± 0.3 | 13.8–12.3 | 13.0 ± 0.4 | 12.9 ± 0.3 |
| M4    | 13.5–12.1 | 12.8 ± 0.3 | | 13.6–12.1 | 12.8 ± 0.4 | |
| M9    | 13.3–12.0 | 12.7 ± 0.3 | 12.6 ± 0.2 | 13.4–11.9 | 12.6 ± 0.4 | 12.5 ± 0.3 |
| (start) | 13.2–11.7 | 12.5 ± 0.4 | | 13.2–11.6 | 12.4 ± 0.4 | |
| (end)  | 13.1–11.4 | 12.3 ± 0.4 | 12.2 ± 0.3 | 13.1–11.3 | 12.2 ± 0.5 | 12.2 ± 0.4 |
| M12   | 13.1–11.0 | 12.0 ± 0.5 | | 13.1–10.9 | 12.0 ± 0.6 | |
to at least 36°S in the mid-latitudes (cf. Koffman et al., 2017), probably driven by latitudinal shifts in the austral westerlies (Quade and Kaplan, 2017).

Changes in the mid-latitude position of the austral westerlies, over sub-decadal to decadal timescales, are probably controlled by the Southern Annular Mode (SAM), considered the primary pattern of Southern Hemisphere climate variability influencing temperature and precipitation from the subtropics to Antarctica (Abram et al., 2014). A positive phase of the SAM occurs when there is anomalously low air pressure over the Antarctic (the Amundsen Sea Low), and high-pressure anomalies over the mid-latitudes (Abram et al., 2014; Lee et al.,...
This enhanced atmospheric pressure gradient results in a strengthening and poleward (southwards) contraction of the austral westerlies. It has been hypothesized, when comparing Antarctic Peninsula and Patagonian Holocene glacial chronologies, that persistent positive phases of the SAM in the past resulted in poleward migration of the austral westerlies causing reduced precipitation in the mid-latitudes, driving negative mass balance and glacier recession (Kaplan et al., 2020). Moreno et al. (2018) reconstructed strong mid-latitude westerlies during the ACR, subsequently weakening during the early Holocene. However, as Kaplan et al. (2020) note, while favouring a positive SAM hypothesis to explain their Holocene moraine records in southern Patagonia and the Antarctic Peninsula, El Niño Southern Oscillation (ENSO)
cannot be discounted as this too has been shown to influence high-latitude climate.

Unpicking past controls on glacier response is important because modern data show a persistently positive phase of the SAM over recent decades. Furthermore, this positive phase has been linked to greenhouse gas emissions and ozone depletion (Thompson et al., 2011; Abram et al., 2014), and is responsible for driving persistent glacier recession in Patagonia and in the eastern Antarctic Peninsula (Kaplan et al., 2020), and increased upwelling of warm ocean waters in the western Antarctic Peninsula (Cook et al., 2014). This is forecast to continue (Thompson et al., 2011; Lee et al., 2019; Pabón-Caicedo et al., 2020), with positive SAM conditions and increased contraction of the austral westlies poleward, forcing glacier recession on the Antarctic Peninsula (Siegert et al., 2019) and western Antarctica (Rintoul et al., 2018), and drying over Patagonia (Boisier et al., 2018). A strong ENSO in 2015/2016 caused drought conditions in central Patagonia, with decreased cloud cover and precipitation in the region (Garreaud, 2018). This therefore demonstrates the relevance of using glacial landform and sedimentary records during Late Pleistocene warming phases to provide empirical data on glacier response to palaeoclimate, providing datasets that can subsequently be used to underpin numerical model simulations of glacier behaviour. Our data from Glaciar Calluqueo demonstrate that targeting small ice caps, rather than large ice fields, offers potential for exploring spatiotemporal glacier response to temperature and precipitation drivers.

Conclusions

Herein, we present nine new cosmogenic surface exposure ages alongside previously published ages (Glasser et al., 2012; Davies et al., 2018) to date four ice limits of Glaciar Calluqueo (Monte San Lorenzo ice cap). The landform record documents active ice recession evidenced by 13 ice-marginal moraines. The oldest M1 moraine dates to the ACR (Davies et al., 2018), the youngest M12 moraine to 12.1 ± 0.4 ka. There was probably relative stability during the Holocene with ice occupying the M13 position until 1945.

We applied Bayesian age modelling to robustly evaluate the cosmogenic surface exposure ages. The resulting geochronology demonstrates that Glaciar Calluqueo receded at 12.75 m a⁻¹ from M1 (13.3 ± 0.3 ka) to M4 (12.9 ± 0.3 ka) before rates increased to a minimum of 29.0 m a⁻¹ back to M9 (12.5 ± 0.4 ka), and 50.3 m a⁻¹ to M12 (12.1 ± 0.4 ka). These rates of recession are of a similar order of magnitude to recession of the Lago Buenos Aires ice-lobe (46.5°S) during Southern Hemisphere warming prior to the ACR (~18.0–15.0 ka), and Calluqueo ice in response to anthropogenic climate heating.

The landform record and geochronology demonstrate rapid glacier recession during a period of Southern Hemisphere warming (~13.0–12.0 ka) following the ACR. Rapid recession during this warming phase was probably sustained during this period by decreases in precipitation and temperature as the austral westlies shifted polewards (Garreaud et al., 2013; Quade and Kaplan, 2017; Moreno et al., 2018).

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Supporting information

Additional supporting information can be found in the online version of this article.

Table S1. Analytical data for the cosmogenic nuclide exposure dating samples reported in the text.

Table S2. Cosmogenic nuclide exposure ages calculated using the St and Lm scaling methods, the regional Patagonian spallation production rate calibration data set (Kaplan et al., 2011) and presented at an erosion rate of 0 mm ka⁻¹.
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