**Late Pleistocene glaciations of the arid subtropical Andes and new results from the
Chajnantor Plateau, northern Chile**

Dylan J. Ward

Jason M. Cesta

University of Cincinnati, Dept. of Geology, Cincinnati, OH 45221

Joseph Galewsky

*University of New Mexico, Dept. of Earth and Planetary Sciences, Albuquerque, NM, USA
87131*

Esteban Sagredo

*Pontificia Universidad Católica de Chile, Santiago, Chile
Millennium Nucleus PaleoClimate*

Correspondence to:

Dr. Dylan J. Ward

Dept. of Geology

University of Cincinnati

ML0013

Cincinnati, OH, USA 45221

Ph.: 1-513-556-2174

Dylan.Ward@uc.edu

Key words: Glacial deposits, paleoclimatology, cosmogenic dating, arid Andes, Altiplano, Last
Glacial Maximum

1. Abstract

The spatiotemporal pattern of glaciation along the Andes Mountains is an important proxy record reflecting the varying influence of global and regional circulation features on South American climate. However, the timing and extent of glaciation in key parts of the orogen, particularly the deglaciated arid Andes, are poorly constrained. We present new cosmogenic ^{10}Be and ^{36}Cl exposure ages for glacial features on and near the Chajnantor Plateau (23°S). The new dates, although scattered due to cosmogenic inheritance, imply that the most recent extensive glacial occupation ended before or during the global Last Glacial Maximum (LGM). We discuss this new record in the context of published glacial chronologies from glacial features in Peru, Bolivia, and northern Chile rescaled using the latest cosmogenic ^{10}Be production rate calibration for the tropical Andes. The results imply regionally synchronous moraine stabilization ca. 25-40 ka, 15-17 ka, and 12-14 ka, with the youngest of these moraines absent in records south of ~20°S, including in our new Chajnantor area chronology. This spatial pattern implicates easterly moisture in generating sufficient snowfall to glacialize the driest parts of the Andes, while allowing a role for westerly moisture, possibly modulated by the migration of the Southern Westerly Wind belt, in the regions near and south of the Atacama Desert.

2. Introduction

The Andes Mountains are presently glaciated, or have been glaciated in the past, along their entire length, from the Northern Hemisphere tropics (e.g., Jomelli et al., 2014) to the upper southern midlatitudes (e.g., Kaplan et al., 2008). The timing and extent of glaciation varies along the Andes, reflecting the spatial and temporal changes to climate factors that influence glaciation, chiefly temperature and precipitation (e.g., Sagredo et al., 2014). Along the strike of

the range, the relative influence of the major global (Hadley) circulation features changes. These larger-scale patterns are modulated by regional systems (e.g., the South American Summer Monsoon; Baker and Fritz, 2015) and by apparent teleconnections with longer-range climatic drivers such as Northern hemisphere Heinrich events (Kanner et al., 2012) or Antarctic polar front migration (Moreno et al., 2009). Accordingly, the history of glaciation along the length of the Andes is germane to understanding linkages between modern climate systems as well as the global progression of regional climate changes during major climate transitions (e.g., periods of glaciation, deglaciation, and abrupt warming; e.g., Rodbell et al., 2009; Denton et al., 2010).

However, in many parts of the Andes, the timing and extent of glaciation is poorly constrained.

The arid subtropical Andes, between 18°S and 27°S, are presently deglaciated, even on mountains higher than 6 km. A few studies (e.g., Jenny et al., 1996) have documented moraines, glaciated bedrock, and other glacial features between 18°S and 23°S, but there is little age control on these features, particularly in the area of the Chilean Altiplano (Zech et al., 2008; Rodbell et al., 2009). It is therefore unknown how the timing of glaciation here relates to the increasingly well-documented timing of glaciation in Patagonia (e.g., Ackert et al., 2008; Kaplan et al., 2011), tropical Peru (e.g., Smith et al., 2008; Jomelli et al., 2014; Kelly et al., 2015), and parts of the Bolivian Altiplano (e.g., Smith et al., 2008; Blard et al., 2009). Given the high precipitation sensitivity of the nearest adjacent groups of modern glaciers (Sagredo et al., 2014), and the overall aridity of the region, it is a fair assumption that the former glaciation of the Chilean Altiplano was modulated at least in part by precipitation changes (e.g., Kull and Grosjean, 2000). However, the area is much farther from oceanic moisture sources than the western Peru cordilleras or the glaciated Chilean peaks south of ~30°S, and far from the eastern Andean foothills and Amazon Basin moisture as well. It is not clear whether past glaciers in the Chilean

Altiplano should have responded similarly to glaciers more directly affected by one or another of these moisture sources.

Here, we report unambiguous evidence for the prior presence of a $> 200 \text{ km}^2$ ice cap on the 5-km-high Chajnantor Plateau (23°S) and a new hybrid cosmogenic ^{10}Be and ^{36}Cl chronology for glacial features in this arid, presently unglaciated part of the Andes. The new dates and the mapped glacial features suggest that this region last deglaciated early during Marine Isotope Stage 2 (MIS 2: 29-14 ka; Lisiecki and Raymo, 2005) and before the global Last Glacial Maximum (here taken as 21 ka). The Chajnantor Plateau site appears to be located near a transition within regional-scale patterns of glacier response to climatic changes, which we document here by compiling published cosmogenic ^{10}Be and ^3He chronologies of latest Pleistocene glacial deposition in the Andes from 10°S to 30°S.

3. Field area geology and modern climate

3.1. Geology and geomorphology

The Chajnantor Plateau (23.00°S, 67.75°W) is a high (5000 m), arid volcanic plateau in the Andes of northern Chile (Fig. 1) that hosts several astronomical observatories, including the Atacama Large Millimeter/Submillimeter Array (ALMA) radio telescope. Its broad, domelike topography was constructed by late Cenozoic volcanism (Schmitt et al., 2001), and most of the plateau is capped by the 1.3 Ma Cajón ignimbrite (Ramírez and Gardeweg, 1982). Many 5600-5700 m stratovolcanoes and lava domes of the Purico Complex (pyroclastic shield) ornament the plateau (Fig. 2). The active stratovolcano Láscar (5592 m) lies 35 km south of the Chajnantor Plateau. Extending approximately 100 km to the north of the Chajnantor Plateau is the Cordillera del Tatio (Fig. 2), a range of previously glaciated, 5000-5500 m volcanic peaks along the Chile-

Bolivia border. The well-known Tatio geyser field lies at ~4300 m on the western flank of these mountains. South of the geyser field, the Cordillera del Tatio includes a pancake-like dacite dome extrusion which we refer to as Cerro Torta; below, we present dating results from moraines near this extrusion in addition to those from the Chajnantor Plateau.

The western flank of the Chajnantor Plateau is overlapped by salt and sediment deposits of the Salar de Atacama, a ~4,500 km² salt flat at 2300 m elevation that extends approximately 100 km to the south from the north end of the plateau. While much of the ignimbrite shield of the Chajnantor Plateau is a bare bedrock surface, locally thick (>50 m) deposits are preserved in patches below the volcanic peaks. These are commonly diamictic and likely include lahar and debris flow deposits (Ramirez and Gardeweg, 1982; Cesta, 2015), other volcanogenic sediment, and glacial and periglacial deposits. ~60 km farther south is Laguna Miscanti, a shallow (10 m maximum depth), saline lake at ~4500 m on the western flank of the Cordón de Puntas Negras (Grosjean et al., 2001). Regionally, small salt flats, alpine wetlands, and small lakes exist above ~4000 m, in addition to hot springs and geyser fields. These features are related to complex groundwater circulation in the mountain block, recharged by the meager precipitation (Risacher and Fritz, 2009).

3.2. Glacial features

Glacial features have been previously recognized in the subtropical arid Andes, mostly above 4000 m elevation, and many of these were mapped by Jenny et al. (1996). Their mapping identifies three main stages of glaciation, with younger, well-preserved “Stage I” and “Stage II” moraines commonly closely nested, and “Stage III” – older moraines, undivided – which are much less well-preserved and in some cases very distal to the Stage I and II moraines. While age

assignments for the moraines in this area suffer from a lack of published dates, it is clear that this area was extensively glaciated, with at least Stages I, II, and the latest Stage III moraines interpreted to have formed in the late Pleistocene (< 120 ka). While the Chajnantor Plateau deposits that are the subject of this study were not part of their mapping, some of these deposits near Cerro Toco, on the northeastern portion of the plateau, were noted by Jenny et al. (1996) and Amman et al. (2001).

Better-known moraines near the Chajnantor Plateau lie about 70 km to the north, among the volcanic peaks above the El Tatio Geysers. These moraines extend ~ 10 km to either side of the range crest and have been variously assumed to date to the time of the global LGM (Graf, 1991; Grenon, 2007), or to late glacial times (Kull and Grosjean, 2000; Amman et al., 2001). Absolute ages for these features have not been published.

3.3. Modern regional climate

The modern mean annual temperature on the Chajnantor Plateau is between -3° and -4°C , and a discontinuous 50 cm permafrost layer exists above 5350 m (Grenon, 2007). The modern annual precipitation is too low to support glaciers: the upper elevations of the plateau receive between 200 and 350 mm/yr of water-equivalent precipitation (Grosjean et al., 2001; Grenon, 2007), with high interannual variability (e.g., Garreaud and Aceituno, 2001; Houston, 2006). Much of this precipitation falls during austral summer and is derived from moist air transported from the east, arriving from the tropical Atlantic via the Amazon Basin. This moisture promotes summer rain and snow and occasional damaging floods (Mather and Hartley, 2005; Houston, 2006). Because of the very low humidity, high winds, and high insolation, most of the snow that

does fall is quickly swept into depressions, where it sublimates. In most years, any snow cover is gone by August.

The overall aridity of the Altiplano and subtropical Andes is maintained primarily by the large-scale subtropical downwelling and cold oceanic upwelling along the west coast of South America related to the Humboldt Current (e.g., Hartley, 2003). An important component of the regional climate is related to the South American Summer Monsoon (SASM), which promotes heavy rainfall in the central Amazon Basin (e.g. Cook and Vizy, 2006). Latent heating associated with the monsoon induces an upper-level high-pressure system called the “Bolivian High” over the subtropical Andes centered around 19°S, 60°W (Lenters and Cook, 1997; Fig. 1). In the modern climate, inter- and intra-annual wet periods in the Altiplano correspond to southward displacements of the Bolivian High (Vuille et al., 1998; Garreaud and Aceituno, 2001). This shifts upper-level easterly winds southward, enhancing moisture transport from the east into the Altiplano and arid Andes (Garreaud, 2000). The basic climatology of the region is also modulated by the El Niño/Southern Oscillation (ENSO), which induces circulation anomalies over South America and is associated with regionally heterogeneous precipitation anomalies (e.g., Vuille, 1999; Garreaud and Aceituno, 2001).

3.4. Nonglacial paleoclimate proxy records

The longest paleoclimate record in the immediate vicinity of our study area comes from a 106 ka salt core dated by U-series from the Salar de Atacama (Bobst et al., 2001). Generally, wet and dry phases of the Salar are synchronized with those interpreted from the Sajama ice cap drill core (18°S, 6542 m) and the Salar de Uyuni (20°S, 3700 m), both in adjacent Bolivia (Baker et al., 2001; Placzek et al., 2013). Wet phases appear in these records near 70, 45, and 20 ka, with

further enhancement ca. 17-14 ka (the Tauca highstand; Sylvestre et al., 1999). These major phases are also observed in speleothem records from Brazil (Cruz et al., 2005, 2006; Wang et al., 2006, 2007) and Peru (Kanner et al., 2012), wherein they are attributed to an orbitally-driven insolation control on precipitation related to modulation of the SASM.

Similarly, based on pollen in rodent middens at Quebrada del Chaco, Chile (25.5°S), Maldonado et al. (2005) infer increased winter (westerly) precipitation during the periods >52, 40-33 ka, 24-17 ka, more precipitation in both seasons 17-14 ka, and more summer (easterly) precipitation from 14-11 ka. Other regional proxy records include a ~26.5 ka sediment core from Laguna Miscanti (Grosjean et al., 2001) and diatomaceous deposits and rodent middens associated with river-margin wetlands in the Atacama Desert (Rech et al., 2002; Latorre et al., 2006; Quade et al., 2008; Gayo et al., 2012). These records also indicate wet periods during MIS 2 (~26.5-22 cal ka), the Tauca highstand (~17.5-15 ka), and in late glacial times (13-10 ka). Based on pollen records from the Laguna Miscanti sediments and altitudinal displacement of taxa, during the late glacial wet periods, precipitation is inferred to have been 2-3x higher than modern (up to ~600 mm/yr rather than ~200 mm/yr; Grosjean et al., 2001; Latorre et al., 2006).

Taken together, these records suggest that the orbital forcing on the SASM system is responsible for the major periods of moistening in the Altiplano and arid Andes. Overprinting this, the SASM system responds to North Atlantic region climate changes on annual to orbital timescales (e.g., Kanner et al., 2012; Baker and Fritz, 2015), potentially allowing these drivers to modulate Altiplanic moisture. The southern and western extent of the influence of these systems, i.e., where and when over the late Pleistocene they become less important than westerly sources of moisture such as midlatitude storms associated with the Southern Westerly Wind (SWW) belt,

is poorly resolved, largely because of the paucity of continuous, high resolution proxy records in the Atacama Desert and surrounding regions.

3.5. Glacial climatology

Sagredo and Lowell (2012) used cluster analysis to classify glaciated regions of the Andes into seven distinct climatic groups. Each group differs from the others in the relative influence of total precipitation, temperature, and seasonality of precipitation on glacial equilibrium line altitudes. According to this classification (Fig. 1B), the Chajnantor area lies directly in between the dry outer tropical glaciers (Group 3; western cordilleras of Peru and Bolivia) and subtropical glaciers (Group 4). By the analysis of Sagredo et al. (2014), the dry outer tropical glaciers receive most of their moisture during DJFM – “Bolivian Winter”, a similar seasonal pattern to southern outer tropical glaciers (Group 2.2; eastern cordilleras, Cordillera Real), but with less precipitation overall. On the other hand, subtropical glaciers to the south of the climatic region known as the “Arid Diagonal” (De Martonne, 1934) receive more moisture during JJA than during the remainder of the year. The Chajnantor area lies in the deglaciated region near the north edge of the Arid Diagonal, so Sagredo and Lowell (2012) and Sagredo et al. (2014) did not include it in their analysis.

Kull and Grosjean (2000) used a dynamical glacier model that employed mass-balance parameters based on local modern lapse rates to simulate a valley glacier matching moraines in the El Tatio area (70 km north of the Chajnantor Plateau). The simulation results suggested that 1000 mm/yr of additional water-equivalent accumulation, dominated by enhanced easterly-sourced austral summer precipitation, would be needed at high altitudes to generate this glacier (Kull and Grosjean, 2000). We note that local late glacial precipitation estimates (Grosjean et al.,

2001; Latorre et al., 2006) do not indicate sufficient precipitation at that time to force this glacier model.

4. Methods

4.1. Geomorphic mapping

Former glacier extents were reconstructed by mapping of moraines and drift limits over an area of 50,000 km² that includes the southern Altiplano and Chajnantor Plateau to the south end of the Salar de Atacama. Regional mapping was performed at a scale of 1:30,000 using the Landsat 8 ETM+ 15-m panchromatic imagery, high-resolution (40 cm) satellite imagery from Google Earth, and a 90 m digital elevation model from the Shuttle Radar Topography Mission product v.2 (Farr et al., 2007).

Detailed mapping of the Chajnantor Plateau was performed at a scale of 1:15,000 using the same LandSat 8 imagery, plus a combination of high-resolution satellite imagery from Google Earth, aerial photographs (1:50,000 from the Chilean Air Force), an 8-m digital elevation model (DigitalGlobe elevation dataset derived from autocorrelation of stereo satellite imagery), and field observations.

Moraines and other drift can resemble landslide and volcanic deposits (e.g. dacitic flank flows and lahars) also present in the field area. We distinguish moraines and drift based on appearance, which is consistent throughout the area as determined by field identification in several locations in the Chajnantor Plateau and Cordillera del Tatio. Moraine deposits in this area tend to be thin and have discrete ridges as compared to the more massive, hummocky landslide and volcanic deposits, and the most prominent set of moraine deposits in the region all fall within a relatively restricted elevation band, whereas many landslide and volcanogenic deposits

are found many hundreds of meters lower in elevation. Other glacial and periglacial features, such as felsenmeer, solifluction lobes, and roche moutonnée are present in the field area, but were not mapped because their extent and interpretation is ambiguous in much of the imagery.

4.2. Cosmogenic exposure dating

In order to constrain the age of glacial deposits identified the Chajnantor area, we collected rock samples for surface exposure (^{10}Be and ^{36}Cl) analysis. Samples were collected from prominent boulders at the crests of moraine ridges and from glacially sculpted bedrock within the identified former ice margins, using hammer and chisel. We selected our sampling sites carefully, exploiting prominent boulders, knobs and ridges to minimize the chance of snow or sediment shielding, and favoring bedrock sampling locations with preserved glacial striations and polish to demonstrate zero post-glacial erosion. Many moraine boulders were in poor condition and had to be rejected for sampling, limiting the density of boulder samples on each moraine. Moreover, the felsic ignimbrite of the area contains variable amounts of quartz, which we found in some samples to be difficult to separate from volcanic glass and other similar-density phases in sufficient quantities, limiting the number of successful ^{10}Be analyses. Here we report the analysis of twenty individual samples, eighteen from the Chajnantor Plateau and two from a small glaciated valley in the Cerro Torta area ~50 km to the north. Of these, eight were analyzed for ^{10}Be only, six for ^{36}Cl only, and six analyzed using both isotopes (Table 2). The twelve ^{10}Be samples were prepared at the University of Colorado and University of Cincinnati using community standard methods (derived from Nishiizumi et al. [1984] and Kohl and Nishiizumi [1992]). The twelve ^{36}Cl samples (from Chajnantor Plateau samples) were prepared at the University of Cincinnati and at the Purdue Rare Isotope Measurement (PRIME) Lab. All

AMS measurements were performed at PRIME Lab. Please see Appendix 1 for complete background details of sample localities and preparation, AMS results, ^{36}Cl sample geochemistry, and age calculations.

Given that glacial margins fluctuate even under a steady climate (Roe and O’Neal, 2014), it is difficult in all cases to interpret a moraine age as the timing of a glacial advance. A moraine may, for instance, indicate the end of a glacial advance and the beginning of retreat, or reflect a large excursion during a protracted stillstand during a glacial maximum or a smaller excursion during an overall retreat (Anderson et al., 2014). In the absence of additional information, we infer moraine ages to represent abandonment of the moraine (i.e., initiation of glacial retreat), thus placing a minimum limiting age on glacial advances. Bedrock ages are interpreted to represent withdrawal of glacial ice from that specific location during retreat (e.g., Ward et al., 2009). In both cases, significant scatter implies either differential erosion between sample locations or variable cosmogenic nuclide inventory inherited from a previous period of exposure.

Table 1. Cosmogenic sample locations

Sample	Type	Analysis	Location	Latitude (deg S)	Longitude (deg W)	Elevation (m)
Chaj-I2	Scoured bedrock	^{10}Be	Central Chajnantor Plateau	23.03424	67.76790	4993
Chaj-12-11	Scoured bedrock	^{36}Cl	Cerro Toco outlet valley, Chajnantor Plateau	22.96558	67.79036	5036
Chaj-12-14	Moraine boulder	^{36}Cl	Cerro Toco outlet valley, Chajnantor Plateau	22.95100	67.81497	4706
Chaj-12-15	Moraine boulder	^{36}Cl	Cerro Toco outlet valley, Chajnantor Plateau	22.95102	67.81488	4704
Chaj-P2	Scoured bedrock	^{10}Be	Cerro Toco outlet valley, Chajnantor Plateau	22.96393	67.80302	4931
Tat-3	Scoured bedrock	^{10}Be	Cerro Torta glacial valley	22.45008	67.99693	4405
Tat-4	Moraine boulder	^{10}Be	Cerro Torta glacial valley	22.45470	67.99914	4533
Chaj-12-06	Moraine boulder	^{36}Cl	Eastern Chajnantor Plateau	22.98889	67.64395	4686
Chaj-12-08	Moraine boulder	^{36}Cl , ^{10}Be	Eastern Chajnantor Plateau	22.96808	67.66564	4708
Chaj-M1	Moraine boulder	^{10}Be	Eastern Chajnantor Plateau	22.98725	67.65284	4660
Chaj-M2	Moraine boulder	^{36}Cl	Eastern Chajnantor Plateau	22.95012	67.68478	4796
Chaj-M3	Moraine boulder	^{10}Be	Eastern Chajnantor Plateau	22.94925	67.68583	4802
Chaj-12-03	Scoured bedrock	^{36}Cl , ^{10}Be	Northern Chajnantor Plateau	22.91785	67.76319	4791
Chaj-12-04	Scoured bedrock	^{36}Cl , ^{10}Be	Northern Chajnantor Plateau	22.91783	67.76243	4789
Chaj-12-18	Moraine boulder	^{36}Cl , ^{10}Be	Western Chajnantor Plateau	23.02044	67.83226	4576
Chaj-12-19	Scoured bedrock	^{36}Cl , ^{10}Be	Western Chajnantor Plateau	23.01647	67.81591	4730
Chaj-12-20	Moraine boulder	^{36}Cl	Western Chajnantor Plateau	23.01366	67.81439	4758
Chaj-12-21	Scoured bedrock	^{36}Cl , ^{10}Be	Western Chajnantor Plateau	23.01327	67.81017	4791
Chaj-I3	Scoured bedrock	^{10}Be	Western Chajnantor Plateau	23.02242	67.82928	4585
Chaj-M5	Moraine boulder	^{10}Be	Western Chajnantor Plateau	23.02380	67.83113	4569

4.3. Age calculations

Conversion of ^{10}Be concentrations to exposure ages was performed using the community-standard CRONUS-Earth online calculator (Balco et al., 2008; <http://hess.ess.washington.edu/math/>; v.2.2, accessed 17 December, 2014). Input tables are provided as supplemental digital files. Terrain shielding was not a significant factor in our field area; our most-shielded sample site was calculated to have a terrain correction factor of 0.2%. As this is an order of magnitude below AMS measurement error, no corrections for shielding were applied except where local shielding or self-shielding were measured as noted in the data tables (Appendix 1).

A significant challenge involving cosmogenic dating in our field area is that the various commonly-used production rate scaling schemes (e.g., Lal/Stone (Lal, 1991; Stone, 2000); Dunai [2001] (Dunai, 2001); Desilets [2003, 2006] (Desilets and Zreda, 2003; Desilets et al., 2006); and Lifton [2005] (Lifton et al., 2005)) agree fairly well at high latitudes and low altitudes, but the Lal/Stone scheme in particular returns older ages than the other schema for sites at low latitudes and high altitudes. Recent work by Lifton (2014) points out systematic biases in the neutron monitor data upon which all but the Lal/Stone method are based and highlights new radiation modeling (Sato et al., 2008) that brings the various scaling schemes into good agreement, even for calibration sites at low latitudes and high altitudes. Pending a community-standard calculation scheme for ^{10}Be exposure ages using the Lifton/Sato scaling method, we used a new calibration of the time-dependent Lal/Stone scaling scheme based on co-dated ^3He and ^{10}Be samples in the subtropical Andes (20°-22°S; Blard et al. [2013]). This new scaling agrees closely (Fig. 3) with new calibration datasets from Patagonia (Kaplan et al., 2011), Peru (Kelley et al., 2015), and New Zealand (Putnam et al., 2010b). All of these calibration data,

individually or compiled (Heyman, 2014), yield lower baseline production rates (~ 3.95 at/g/yr) than the older Balco (2008) and Balco et al. (2009) calibrations by $\sim 20\%$ (although, coincidentally, resulting ages are similar to those given by time-independent Lal/Stone scaling with the Balco (2008) calibration dataset; Fig. 3). Because of the proximity of our field area to the calibration site of Blard et al. (2013), we use this as the basis for our reported ages.

Interpreting ages from cosmogenic ^{36}Cl is more complicated than from ^{10}Be , because there are multiple production pathways. These include high-energy neutron spallation of Ca and K, lower-energy epithermal and thermal neutron capture by ^{35}Cl , and minor muonogenic production, along with nucleogenic production (Gosse and Phillips, 2001), as opposed to simply neutron spallation and minor muonogenic production of ^{10}Be . These production pathways depend on the major and trace elemental chemistry of the samples, including water content to a small degree, and uncertainty in their scaling with latitude and altitude contributes to greater uncertainty, particularly in whole-rock samples as we have used here.

Our ^{36}Cl ages were calculated using the Age Calculation Engine (Zweck et al., 2012; <http://ace.hwr.arizona.edu>) and a time-varying production rate scaling scheme based on latitude-altitude-pressure factors of Stone (2000), sea level curve of Fairbanks (1989) and Shackleton (2000), geomagnetic intensity model of Guyodo and Valet (1999), and paleolatitude record of Ohno and Hamano (1992) and Yang et al. (2000). These choices are consistent with those used by the CRONUS calculator for ^{10}Be and therefore with the other chronologies presented here. Among the available calibration datasets for ^{36}Cl production rates, the Phillips et al. (1996) and Licciardi et al. (2008) datasets produce ^{36}Cl ages most consistent with the ^{10}Be ages of paired samples and samples from the same landform (see Results, section 5, below). Because the low-energy production pathways for ^{36}Cl are generally scaled along with the high-energy pathways

for convenience, yet their scaling is not well-constrained to be the same (e.g., Desilets et al., 2006; Schimmelpfennig et al., 2009), a discussion of which combination of scaling model and calibration dataset is theoretically most appropriate for our study area is beyond the scope of this work. Here, we present the ages based on the scheme that independently gives results best matched overall to the ^{10}Be ages, but do not directly calibrate ^{36}Cl production rates based on our paired samples, because the paired sample set is too small to screen for other random error such as erosion and anomalous histories of sediment cover.

4.4. Rescaled regional records of past glacial maxima

To take a broad look at spatiotemporal trends in the timing of glaciation in this part of the Andes, we used the CRONUS online calculator to rescale the reported ^{10}Be ages from 16 published studies of pre-Holocene tropical and subtropical glacial deposits using a base production rate calibrated based on the same dataset as our Chajnantor chronology, the tropical Andes dataset of Blard et al. (2013) (Table 2). These records were selected by the following criteria: 1) locations between $\sim 10^\circ$ and $\sim 30^\circ\text{S}$; 2) sufficient published metadata to recalculate exposure ages; and 3) multiple boulder ages reported from one or more moraines. We also include three ^3He chronologies, scaled consistently with Blard et al. (2013) as originally published, for a total of 20 records. Each record was examined individually, and outliers that had been excluded by the original authors based on landform context and/or statistical analysis were removed from our dataset. With the exception of our Chajnantor record, we also excluded bedrock exposure ages (e.g., from roche moutonnée) from each of these studies that were not immediately adjacent to a moraine deposit; this ensured that intermediate ages associated with retreat between moraine positions were not confused with periods of stillstand or advance. These

measure were meant to ensure that clusters of ages within each record reflect a consistent measure of the timing of stabilization of a particular moraine deposit.

Any combination of multiple published glacial chronologies spanning a large geographical area and published by different workers over the course of a decade is prone to factors that reduce precision relative to the original studies taken in isolation. These may include underconstrained or underreported geological biases related to the specific study sites (e.g., sample thickness, local constraints on erosion rate, and cosmogenic inheritance), and selective sampling of particular moraines either by study design or because of inadequate sample availability on other moraines. For example, deposits older than LGM are reported but not sampled at many locations, and where they are sampled, older ages are much less precise than younger ones due to scaling uncertainties. Regional atmospheric pressure patterns may also cause variability in production rate scaling between records that is difficult to assess, but may be of order 7-10% (Farber et al., 2005; Blard et al., 2013). Additionally, prior to revision of the ^{10}Be half-life and publication of Nishiizumi et al. (2007), it was not common practice to report AMS $^{10}\text{Be}/^9\text{Be}$ ratio standardizations in publications, so some chronologies may be prone to a systematic error related to inferring these values in rescaling the data, which could generate artificial misalignment between records (we have noted these cases in the supplemental data tables).

Because of these impediments to precision of our merged chronology, we took a broad view and grouped samples from each study that fell within specific time windows: 70-100 ka; 40-70 ka; 25-40 ka; 20-25 ka (LGM); 15-20 ka (Tauca); and 10-15 ka (late glacial). We did not address the Holocene portions of any of these chronologies. For one of these periods to be counted as a time of moraine stabilization, the central value of at least 3 samples had to fall within in that

period. Periods containing ≤ 2 samples or samples with high uncertainty relative to others in the dataset were classified as “few/poor ages”.

Table 2. Sources for regional cosmogenic chronologies.

Map ID	Nuclide	Location	Mean Latitude (deg S)	Mean Longitude (deg W)	Mean Elevation (m)	Reference
1	^{10}Be	Cordillera Blanca, Peru	9.57	77.42	3909	Farber et al., 2005*
2	^{10}Be	Cordillera Blanca, Peru	10.01	77.27	4321	Smith and Rodbell, 2010
3	^{10}Be	Cordillera Blanca, Peru	10.01	77.25	4391	Glasser et al., 2009
4	^{10}Be	Cordillera Huayhuash, Peru	10.23	76.89	4184	Hall et al., 2009
5	^{10}Be	Eastern Cordillera, Peru	11.05	75.97	4301	Smith et al., 2005*
6	^{10}Be	Cordillera Vilcabamba, Peru	13.35	72.51	4330	Licciardi et al., 2009
7	^{10}Be	Quelccaya, Peru	13.94	70.89	4855	Kelly et al., 2015
8	^3He	Nevado Coropuna, Peru	15.51	72.58	5074	Bromley et al., 2009, 2011
9	^{10}Be	Cordillera Real, Bolivia	15.98	68.54	4625	Zech et al., 2007a
10	^{10}Be	Cordillera Real, Bolivia	16.13	68.18	3675	Smith et al., 2005*
11	^{10}Be	Cordillera Real, Bolivia	16.26	68.11	4605	Jomelli et al., 2011
12	^{10}Be	Cordillera Real, Bolivia	16.38	68.18	4604	Smith et al., 2005*
13	^{10}Be	Cordillera Real, Bolivia	16.59	67.80	4352	Smith et al., 2011
14	^{10}Be	Cochabamba, Bolivia	17.22	66.36	4085	Zech et al., 2007a
15	^3He	Cerro Tunupa, Bolivia	19.87	67.61	4074	Blard et al., 2009
16	^{10}Be	Tres Lagunas, Argentina	22.20	65.12	4426	Zech et al., 2009
17	^3He	Cerro Uturuncu, Bolivia	22.31	67.20	4878	Blard et al., 2014
18	$^{10}\text{Be}/^{36}\text{Cl}$	Chajnantor Plateau, Chile	22.91	67.80	4701	This study
19	^{10}Be	Encierro Valley, Chile	29.10	69.90	3902	Zech et al., 2006
20	^{10}Be	Cordón de Doña Rosa, Chile	30.71	70.40	3144	Zech et al., 2007b

* As recompiled by Smith et al., 2008

5. Results

5.1. Mapping of glacial features

Our mapping of glacial features at the regional scale (Fig. 3) reveals that approximately 4100 km² of the 50,000 km² map area of Fig. 2 was either ice-covered or was encompassed within an ice-covered area. If the ice cover was contemporaneous across the entire map area, then ~10-15% of the land area above 4000 m was ice-covered at that time. Where they overlap, our map generally agrees with that of Jenny et al., 1996, including the two closely-spaced younger stages and the more distal, poorly preserved older stage or stages of glaciation; we therefore adopt their nomenclature for the Stage I-III moraines.

Glacial features on the Chajnantor Plateau include boulder-studded moraines, reaching 10 m in height and stretching for many tens of km around the east side of the plateau (Fig. 4). Their counterparts on the western side of the plateau are more subdued, between 2 and 5 m high, but are also studded with meter-scale boulders. In satellite imagery, these deposits can be traced continuously around the plateau, outlining a former ice cap with a diameter of ~ 20 km (Fig. 5). Glacially polished and striated bedrock are common within the moraine boundaries (particularly associated with hard andesite), as are erratics and other large, isolated boulders (Fig. 4). Outlying, more poorly preserved deposits that indicate a prior, more extensive glaciation are also apparent in the field and in the satellite imagery. These are typically broad, low (1-3 m) gravelly deposits with scattered boulders that follow lobate shapes in map view and could be interpreted as degraded moraine fragments.

The most prominent (local Stage II) moraines on the Chajnantor Plateau extend to approximately 4600 m elevation on the eastern side and ~4400 m on the western side of the plateau. The lowest, most distal Stage III moraines mapped reach as low as ~4200 m, on the

western side of the plateau. On the eastern side, the Stage I moraines appear to be closely nested within the Stage II moraines, separated by < 1 km (typically by 100-200 m). On the western side, the Stage I moraines are typically farther inboard than the Stage II moraines by 1.5-2 km, up to 5 km in one outlet valley (Fig. 5).

5.2. ^{10}Be and ^{36}Cl exposure ages

The 14 ^{10}Be analyses yielded ages that range from 18 ka to 141 ka (Table 3). Seven of the ^{10}Be samples yielded ages between 18 and 30 ka. Our dataset includes six samples with paired ^{10}Be and ^{36}Cl analyses. Using the ACE calculator and scaling scheme choices described above, four of the ^{36}Cl ages on double-dated samples are consistent with the ^{10}Be ages within error and three are nearly identical (Table 3; Fig. 7). Those with the most identically matched ^{10}Be and ^{36}Cl ages (Chaj-12-03: 37.6 ± 2.7 ka Be, 38.4 ± 5.2 ka Cl; Chaj-12-04: 24.9 ± 1.7 ka Be, 26.4 ± 4.4 ka Cl) are sampled from striated, polished andesitic bedrock, so erosion is known to have been zero over the period of exposure. The two others with well-matched exposure ages are from ignimbritic moraine boulders. The two remaining sets of paired samples yield irreconcilably different ^{36}Cl and ^{10}Be ages; one (Chaj-12-21) has an unreliable ^{10}Be result (~50% analytical uncertainty, and unreasonably high exposure age relative to nearby samples), and the other yields a ^{36}Cl age consistent with other ages from the same moraine, whereas the ^{10}Be age is approximately 10% as old, again suggesting an unreliable ^{10}Be measurement. We take the ^{36}Cl age as preferred for these two samples.

5.3. Description of age results and landform associations

Other ^{36}Cl ages in this dataset generally corroborate the picture painted by the ^{10}Be ages, with bedrock ages falling in proper relative order with ^{10}Be counterparts, and similar moraine

ages on the eastern Chajnantor Plateau (Fig. 5; Table 2). Bedrock ages tend to be clustered in the period 18-30 ka, whereas moraine boulder ages are spread throughout the last 180 ka (Fig. 8). The resulting overall age distribution (both methods) is broad and scattered, with a strong peak at 25 ka.

Stage I (Fig. 8 A, B): Four Stage I moraine exposure ages range from 19.1-30.1 ka, with one outlying 122 ka boulder. Five of the six exposure ages on bedrock within the Stage I moraines fall between 30 ka and 25 ka, in a general progression from oldest near the moraines and younger toward the center of the plateau near the high topography of Cerro Toco and Cerro Chajnantor. Similarly, the bedrock exposure age within the Stage I moraines near Cerro Torta is 30 ka. The 24.6 ka boulder from the moraine near Cerro Torta (sample Tat-4) was somewhat degraded and only ~50 cm tall, and so we regard this as a minimum age; given the consistently older bedrock exposure ages upvalley of the Stage I moraines, we take the 19.1 ka age of sample Chaj-12-20 as also affected by erosion or boulder rotation, and estimate the Stage I moraine abandonment age at ~30 ka.

Stage II (Fig. 8 C, D): Stage II moraine samples yield four overlapping ages that range from 43.1 to 58.6 ka, with two outliers of 89.1 ka and 141 ka. Two bedrock samples (Chaj-12-19 and Chaj-13) between the moraines of Stages I and II on the western side of the plateau yield ages of 20.8 and 18.2 ka, younger than the more proximal bedrock ages within the Stage I moraines and generally younger than the Stage I moraine samples. These samples were taken from striated bedrock, so erosion of the bedrock itself is negligible here, so we interpret these samples as having been affected by a transient cover of sediment. Our interpreted timing of the Stage II glacial maximum is therefore between ~40-60 ka, based on the overlapping moraine boulder ages.

Stage III: A single boulder sampled on a degraded Stage III moraine on the eastern side of the Chajnantor Plateau yielded an age of 177 ± 38 ka.

Table 3: Cosmogenic exposure ages by landform stage, in increasing age order

Sample	^{10}Be Age (ka)	^{10}Be Uncertainty (ka)	^{36}Cl Age (ka)	^{36}Cl Uncertainty (ka)	(M)oraine or (B)edrock	Local stage	Preferred Age (ka)	Pref. Age Uncertainty (ka)
Chaj-12-21	94.6	47.9	14.7	1.1	B	I	14.7	1.1
Chaj-12-04	24.9	1.7	26.4	4.4	B	I	24.9	1.7
Chaj-12-11			25.7	1.7	B	I	25.7	1.7
Chaj-P2	25.9	2.2			B	I	25.9	2.2
Chaj-I2	30.2	2.6			B	I	30.2	2.6
Tat-3	30.7	2.3			B	I	30.7	2.3
Chaj-12-03	37.6	2.7	38.4	5.2	B	I	37.6	2.7
Chaj-12-19	18.2	2.8	11.2	1.3	B	II	18.2	2.8
Chaj-I3	20.8	1.5			B	II	20.8	1.5
Chaj-12-20			19.1	1.6	M	I	19.1	1.6
Tat-4	24.6	1.8			M	I	24.6	1.8
Chaj-12-15			30.1	2.8	M	I	30.1	2.8
Chaj-12-14			122	14.3	M	I	122	14.3
Chaj-12-18	43.1	7	40.8	3.5	M	II	43.1	7
Chaj-12-08	8.2	2.6	49	13.1	M	II	49.0	13.1
Chaj-M1	50.2	3.7			M	II	50.2	3.7
Chaj-M2			58.6	16.4	M	II	58.6	16.4
Chaj-M3	89.1	6.2			M	II	89.1	6.2
Chaj-M5	141	9.3			M	II	141	9.3
Chaj-12-06			177	38	M	III	177	38

1. All uncertainties include 1- σ analytical uncertainty and propagated production rate scaling uncertainty.
2. ^{10}Be scaling: Time-varying Lal/Stone using Blard et al., 2013 calibration set. ^{36}Cl scaling: Time-varying Lal/Stone using Phillips, 1996 calibration set.
3. On paired samples (shaded rows), the ^{10}Be age is preferred except where inconsistent with landform context and/or exhibits large measurement uncertainty (struck through records).

5.4. Results of rescaled regional ^{10}Be glacial chronologies

To visualize the results of rescaling regional cosmogenic chronologies, we generated probability density (“camel”) plots using all rescaled ages in each record (Fig. 9; high-resolution version provided as supplemental figure S1). These plots are constructed by assuming Gaussian

uncertainty distributions around the central value, and that these distributions represent probability density functions of age; the probability density is binned along the time axis to generate the plot, such that many overlapping samples with slightly different central values generate one merged peak that effectively represents a weighted average (Lowell, 1995). We interpret the timing of peaks in each histogram as indicating moraine formation and stabilization (i.e., glacial maxima and beginning of retreat). Three main periods of moraine formation appear overall (Fig. 9): early LGM (25-40 ka), late LGM/deglacial (~16-17 ka), and late glacial (10-14 ka). In some records, particularly in the tropics, continuous glaciation and multiple stages of moraine formation between the late glacial and LGM are indicated. One important result that emerges from the older ages given by the new cosmogenic scaling is that some moraines previously assigned to the late glacial (<15 ka) are now perhaps better thought of as dating to the Tauca stage (~16-17 ka), while a separate late glacial (10-13 ka) peak, previously assigned to the early Holocene, appears in many records, especially those in the tropics.

To examine spatiotemporal trends, we generated a regional map that includes the locations of the rescaled cosmogenic records (Fig. 10). Each panel shows this map during one of the time windows of interest (70-100 ka; 40-70 ka; 25-40 ka; 20-25 ka; 15-20 ka; and 10-15 ka), and the points representing study locations are colored according to the presence or absence of clustered exposure ages during that period.

The maps show that few published records begin before or during the 70-100 ka period. Clustered exposure ages appear across the entire latitude range during the 40-70 ka, 25-40 ka, and 15-20 ka periods, with few ages falling during the 20-25 ka period. A strong 10-15 ka clustering appears as well, but only in areas north of ~20°S or on the eastern side of the

Altiplano. Southwest of the Salar de Uyuni area, this younger late glacial period of moraine development is not reported, and it is not in evidence at Chajnantor.

6. Discussion

6.1. Geological sources of uncertainty in Chajnantor exposure ages

Cosmogenic exposure ages are prone to two competing sources of geological uncertainty: 1) inheritance of inventory from a prior period of exposure, i.e., incomplete resetting by erosion, which increases apparent ages, and 2) postdepositional erosion or transient cover of sediment, tephra, or snow that reduces apparent ages. The scatter among the cosmogenic ages from our study area implies that one or both of these are affecting our results. For example, as described above, the exposure age of the Stage I moraine boulder from Co. Torta is 24.6 ± 1.8 ka, while that of the bedrock sample immediately upvalley is 30.7 ± 2.3 ka (Fig. 6), which suggests either incomplete resetting of the bedrock sample or postdepositional erosion of the moraine boulder. Because this moraine boulder was only ~50 cm high and slightly degraded, we take its exposure age to be minimum-limiting and the 30.7 ka bedrock exposure age to more accurately reflect the timing of deglaciation of this valley. Similarly, samples Chaj-12-03 (37.6 ka) and Chaj-12-04 (24.9 ka) are within 5 m of each other on a striated andesitic surface, and so would be expected to yield identical ages. Given that the striations provide evidence for no postdepositional erosion at this location, the older age of Chaj-13-3 is likely to be unreset, again implying shallow and variable bedrock erosion. The difference between these can be accounted for by ~30 cm of additional erosion (assuming a period of prior exposure of ~70 ka and a rock density of 2.7 g/cm^3) at the site of Chaj-12-4. The scatter among moraine ages (Fig. 8) further supports the likely presence of cosmogenic inheritance in many samples.

6.2. Interpretation of Chajnantor area glacial history

Our exposure age chronology supports stabilization of Stage II moraines ca. 40-60 ka; continued glaciation until 30 ka or reglaciation at that time (Stage I moraines); and retreat and deglaciation by 25 ka (bedrock exposure in the central plateau). Older Stage III deposits were likely formed prior to 100 ka. It is not possible with these data to determine whether the area deglaciated more completely in between Stages II and I, or if the ice cap simply retreated to the Stage I position. However, the relatively small moraines and widespread cosmogenic inheritance suggest limited erosion, perhaps implying that glacial occupations of the plateau were short-lived. Our cosmogenic ages, taken in isolation, suggest deglaciation of the Chajnantor area early in MIS 2 and prior to the global LGM (Fig. 11). We found no evidence of an extensive local glacial advance or stillstand during either the Tauca phase ca. 17 ka or later in the late glacial, implying either that the estimated 3x increase in precipitation in the area at that time (Grosjean et al., 2001) was insufficient or that temperatures were insufficiently cold to generate large glaciers.

6.3. Chajnantor chronology in the context of regional records

The Chajnantor area appears to have been glaciated during periods of elevated precipitation, as would be expected based on the precipitation sensitivity of modern glaciers in the outer tropics and subtropics (Sagredo et al., 2014). Stage II moraine ages at Chajnantor reasonably align with a period of lake expansion documented in the Salar de Uyuni, on the Bolivian Altiplano, ca. 40-50 ka during Marine Isotope Stage (MIS) 3 or late in MIS 4 (Fig. 11; Baker et al., 2001; Placzek et al., 2013). Similarly, the early LGM glaciation at Chajnantor coincided with moistening periods in the Salar de Atacama and Salar de Uyuni. Curiously, it appears to have deglaciated prior to the wettest (Tauca) phase, 17-14 ka, but this occurred after

global temperatures had increased $\sim 4\text{--}6\text{ }^{\circ}\text{C}$ following the LGM (Denton et al., 2010). The combination of precipitation and temperature conditions needed to glacialize the area may not have been met at that time, despite higher precipitation.

The only dated glacial record in our regional synthesis within $\sim 250\text{ km}$ of Chajnantor is the cosmogenic ^3He record at Uturuncu Volcano, Bolivia (Fig. 9; Blard et al., 2014). At both of these sites, moraine ages appear in the 40–70 ka period and near 30 ka. A paucity of moraines at Uturuncu during the $\sim 25\text{ ka}$ period is compatible with the inferred retreat and deglaciation at Chajnantor during that time. However, while there is no strong evidence for a 15–20 ka (Tauca) moraine at Chajnantor, there is one at Uturuncu. This might reflect the more proximal position of Uturuncu to Lake Tauca itself, which has been interpreted as a local moisture source supporting glaciation elsewhere in Bolivia (Blard et al., 2009). It may also be a function of altitude; the average elevation of the moraines dated at Uturuncu is nearly 200 m higher than that at Chajnantor (Table 2), with a different hypsometry ($\sim$ conical volcanic peak vs. the broad Chajnantor Plateau).

Along with our new chronology at Chajnantor, the regional ^{10}Be and ^3He glacial chronologies we compiled (Fig. 10) suggest that glaciers between 10°S and 30°S behaved more or less synchronously before, during, and following the global LGM. Most of these records imply moraine formation during the period 25–40 ka, deglaciation or still-stand during the classic Last Glacial Maximum (20–25 ka), and widespread moraine formation again during 15–20 ka. Glaciers south of $\sim 20^{\circ}\text{S}$ on the western flank of the Andes do not appear to have formed moraines during the late glacial (10–15 ka), while the more tropical glaciers did. The spatial pattern of the late glacial in particular mimics the climate zones of Sagredo and Lowell (2012) (Fig. 10), with the dry outer tropical and southern outer tropical glaciers (Groups 3 and 2.2,

respectively; dominated by easterly, austral summer precipitation) showing a strong late glacial response, but with late glacial deposits not present or unreported in the region of the subtropical glaciers (Group 4; dominated by westerly, austral winter precipitation).

6.4. Climate drivers of glaciation in the arid Andes

For the purposes of the following discussion, we interpret periods of moraine formation as minimum-limiting ages on glacial advances. Compared with global climate events, our merged record outlines an early LGM glacial advance, a Heinrich Stadial 1 (HS1) readvance (17-14 ka, coinciding with the Tauca highstand of Altiplanic lakes), and a late glacial readvance corresponding either to the Northern Hemisphere Younger Dryas stadial event (YD; 12.9-11.7 ka) or the Southern Hemisphere Antarctic Cold Reversal (ACR; 14.5-12.8 ka).

Our results support the hypothesis that subtropical South American glaciers are sensitive to Northern Hemisphere drivers. As we described above (section 3.4), recent speleothem work from Peru strongly implies that tropical climates respond to northern hemisphere climate forcing via modulation of the South American Summer Monsoon (SASM; Kanner et al., 2012). Speleothem records from eastern Brazil suggest that the SASM system (including the related Bolivian High) shifts southward during periods of high southern hemisphere insolation (Cruz et al., 2005), due to changes in the position and intensity of latent heating associated with the SASM (Rodwell and Hoskins, 2001; Gandu and Dias, 1998). A corresponding southerly shift in the associated Bolivian High would bring easterly-sourced precipitation to the Altiplano (Fig. 1B; Lenters and Cook, 1997; Vuille, 1999). Northern hemisphere glacial fluctuations, including ice sheet expansion and millennial-scale Heinrich events, appear to enhance southerly shifts of

the SASM and South Atlantic Convergence Zone (Cruz et al., 2006) and weaken the southern Hadley cell and southern subtropical jet (Chiang et al., 2014).

The influence of Atlantic drivers may wane to the southwest across the Andes, with the Southern Westerly Wind belt (SWW) becoming progressively more important as a moisture source. Some enhancement in precipitation to the arid Andes may have been caused by a more northerly position for the SWW at LGM (Maldonado et al., 2005; Rojas et al., 2009); while this was likely not the primary moisture source as far north as Chajnantor, it may have enhanced overall conditions in favor of glaciation. By ~17 ka, the SWW was migrating back to the south, returning to approximately its current position by 15 ka (e.g., McCullough et al., 2000; Rojas et al., 2009; Boex et al., 2013). This is consistent with the shift from winter, westerly precipitation sources to summer, easterly sources documented at Quebrada del Chaco between 17 ka and 11 ka by Maldonado et al. (2005). The corresponding loss of the westerly moisture may partly explain the lack of glacial response during the late glacial south of 20°S (Fig. 8).

Jomelli et al. (2014) argue based on rescaled cosmogenic chronologies from tropical glacier deposits (some of which were included in this study), and a new ¹⁰Be record in Colombia, that the largest late glacial advance throughout tropical South America took place not during the YD but during the ACR, a period during which southern mid-high latitude air temperatures (e.g., Monnin et al., 2001; Williams et al., 2005) and sea-surface temperatures in the Southern Ocean and Pacific Ocean near South America ceased warming or declined slightly (e.g., Lamy et al., 2004). Their results suggest a more limited glacial response during the Younger Dryas. Insofar as tropical glaciers are more sensitive than subtropical glaciers to temperature (Sagredo et al., 2014), it is conceivable that suppressed sea-surface temperatures during the ACR resulted in a readvance of tropical glaciers but was not combined with sufficient moistening to trigger

reglaciation of the arid Andes. Other Southern Hemisphere settings that are more sensitive to temperature than precipitation support this mechanism. For example, moraines in New Zealand dated by Putnam et al. (2010a) imply a glacial advance there at the end of the ACR. Most of the South American records do not attain the age precision of either the Jomelli et al. (2014) or the Putnam et al. (2010a) studies, however, and so the end of the ACR and the beginning of the YD would be difficult to distinguish with confidence. Likewise, our merged, rescaled record is insufficiently precise to distinguish between the YD and the ACR. We do note, however, that in eight of the 14 records in our study that show a strong late glacial signal (Fig. 8, 9 – records 1-4, 6-14, 16), the late glacial samples cluster more closely within the YD (12.9-11.7 ka) rather than in the ACR (14.5-12.8 ka). Similarly, precise radiocarbon and ^{10}Be ages indicate that the Quelccaya Ice Cap (Peru, 13.9°S) retracted during the ACR and advanced during the peak of the YD (12.5-12.35 ka; Kelly et al., 2012, 2015).

A potential resolution to these conflicts could be that glaciers in temperature-sensitive locations (e.g., the tropics and Patagonia) readvanced more prominently during the ACR, due to its effect on regional temperatures, whereas those in more precipitation-limited settings readvanced during the YD, which, like HS1, was accompanied by increased easterly moisture supplied to the arid Andes due to Northern Hemisphere modulation of South American circulation (Placzek et al., 2013). Glaciation in the more southern and western arid Andes (between 20°S and ~35°S; e.g., Chajnantor) required a combination of wetter and colder not found after HS1.

7. Conclusions

We presented the first cosmogenic ^{10}Be and ^{36}Cl exposure ages for glacial features in an arid, presently unglaciated part of the Chilean Andes. The new dates and the mapped glacial features suggest that this region was extensively glaciated prior to or during the global Last Glacial Maximum. Rescaled regional ^{10}Be glacial chronologies demonstrate that late Pleistocene wet periods and glaciations were generally synchronous across the dry Andes, with diminishing strength to the southwest across the Altiplano. This spatial pattern implicates easterly moisture in generating sufficient snowfall to glacialize the driest parts of the Andes, while allowing a role for westerly moisture, possibly modulated by the migration of the Southern Westerly Wind belt, in the regions near and south of the Arid Diagonal. Taken in context of the above discussion, our study suggests the hypothesis that the late glacial advance of Andean glaciers in temperature-sensitive locations occurred during the ACR, whereas those in more precipitation-limited settings readvanced during the YD, which like the pre-LGM and HS1, was accompanied by increased easterly moisture to the arid Andes. Glaciation in the arid Andes between 20°S and $\sim 35^{\circ}\text{S}$ required a combination of wetter and colder not found after HS1. To evaluate this hypothesis will require more precise glacial chronologies to resolve the small time differences between the YD and the ACR, along with detailed examination and precise dating of sparse arid Andes glacial deposits south of $\sim 20^{\circ}$ to confirm whether and to what extent glaciation occurred during the late glacial.

8. Acknowledgments

The authors would like to thank Marc Caffee, J Radler, and Susan Ma at PRIME Lab for AMS support; Bob Anderson, Miriam Dühnforth, and Kurt Refsnider for logistical support at the University of Colorado Cosmogenic Isotope Lab and Sarah Hammer for support at the University of Cincinnati Cosmogenic Isotope Labs; as well as Kimberly Samuels, Alex Lechler, Chris Sheehan, and the staffs of the Atacama Large Millimeter/Submillimeter Array and the Caltech Chajnantor Test Facility for assistance in the field. We thank Tom Lowell and Andrew Malone for discussion that helped clarify some of the ideas presented here. Funding was provided by the University of New Mexico Research Allocations Committee, the University of New Mexico Latin American and Iberian Institute, and NSF grants EPS-0918635 and EPS-0814449 (New Mexico EPSCoR) and EAR-1226611 (Geomorphology and Landuse Dynamics). We thank two anonymous referees for unusually thorough and constructive reviews.

9. References

- Ackert Jr., R.P., Becker, R.A., Singer, B.S., Kurz, M.D., Caffee, M.W., Mickelson, D.M., 2008. Patagonian glacier response during the Late Glacial-Holocene transition. *Science* 321, 392–395.
- Amman, C., Jenny, B., Kammer, K., Messerli, B., 2001. Late Quaternary glacier response to humidity changes in the arid Andes of Chile. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 172, 313–326.
- Anderson, L.S., Roe, G.H., Anderson, R.S., 2014. The effects of interannual climate variability on the moraine record. *Geology* 42, 55–58.
- Baker, P.A., Rigsby, C.A., Seltzer, G.O., Fritz, S.C., Lowenstein, T.K., Bacher, N.P., Veliz, C., 2001. Tropical climate changes at millennial and orbital timescales on the Bolivian Altiplano. *Nature* 409, 698–701.
- Baker, P., Fritz, S., 2015. Nature and causes of Quaternary climate variation of tropical South America. *Quaternary Science Reviews* 124, 31–47.
- Balco, G., Stone, J.O., Lifton, N.A., Dunai, T.J., 2008. A complete and easily accessible means of calculating surface exposure ages or erosion rates from ^{10}Be and ^{26}Al measurements. *Quaternary Geochronology* 3, 174–195.
- Balco, G., Briner, J., Finkel, R.C., Rayburn, J.A., Ridge, J.C., Schaefer, J.M., 2009. Regional beryllium-10 production rate calibration for northeastern North America. *Quaternary Geochronology* 4, 93–107.

667 Blard, P.-H., Lavé, J., Farley, K.A., Fornari, M., Jiménez, N., Ramirez, V., 2009. Late local
 668 glacial maximum in the Central Altiplano triggered by cold and locally-wet conditions during
 669 the paleolake Tauca episode (17-15ka, Heinrich 1). *Quaternary Science Reviews* 28, 3414-
 670 3427.

671 Blard, P.H., Braucher, R., Lavé, J., Bourlès, D., 2013. Cosmogenic Be-10 production rate
 672 calibrated against He-3 in the high Tropical Andes (3800–4900 m, 20–22° S). *Earth and*
 673 *Planetary Science Letters* 382, 140-149.

674 Blard, P.-H., Lave, J., Farley, K.A., Ramirez, V., Jimenez, N., Martin, L.C.P., Charreau, J.,
 675 Tibari, B., Fornari, M., 2014. Progressive glacial retreat in the Southern Altiplano (Uturuncu
 676 volcano, 22°S) between 65 and 14ka constrained by cosmogenic ^3He dating. *Quaternary*
 677 *Research* 82, 209-221.

678 Bobst, A.L., Lowenstein, T.K., Jordan, T.E., Godfrey, L.V., Ku, T.L., Luo, S., 2001. A 106 ka
 679 paleoclimate record from drill core of the Salar de Atacama, northern Chile.
 680 *Palaeogeography, Palaeoclimatology, Palaeoecology* 173, 21–42.

681 Boex, J., Fogwill, C., Harrison, S., Glasser, N. F., Hein, A., Schnabel, C., Xu, S., 2013. Rapid
 682 thinning of the late Pleistocene Patagonian Ice Sheet followed migration of the Southern
 683 Westerlies. *Scientific reports*, 3, doi:10.1038/srep02118.

684 Bromley, G.R.M., Schaefer, J.M., Winckler, G., Hall, B.L., Todd, C.E., Rademaker, K.M., 2009.
 685 Relative timing of last glacial maximum and late-glacial events in the central tropical Andes.
 686 *Quaternary Science Reviews* 28, 2514-2526.

- 687 Bromley, G.R.M., Hall, B.L., Schaefer, J.M., Winckler, G., Todd, C.E., Rademaker, K.M., 2011.
688 Glacier fluctuations in the southern Peruvian Andes during the late-glacial period,
689 constrained with cosmogenic ^3He . *Journal of Quaternary Science* 26, 37-43.
- 690 Cesta J., 2015. Timing of alluvial fan development along the Chajnantor Plateau, Atacama
691 Desert, northern Chile: Insights from cosmogenic ^{36}Cl . MS thesis. University of Cincinnati,
692 Cincinnati, OH, USA.
- 693 Chiang, J. C., Lee, S. Y., Putnam, A. E., Wang, X., 2014. South Pacific Split Jet, ITCZ shifts,
694 and atmospheric North–South linkages during abrupt climate changes of the last glacial
695 period. *Earth and Planetary Science Letters* 406, 233-246.
- 696 Cook, K.H., Vizy, E.K., 2006. South American climate during the Last Glacial Maximum:
697 delayed onset of the South American monsoon. *J Geophys Res* 3, 1–21.
- 698 Cook, K. H., 2009. South American climate variability and change: remote and regional forcing
699 processes. In *Past Climate Variability in South America and Surrounding Regions*, 193-212.
700 Springer Netherlands.
- 701 Cruz, F.W., Burns, S.J., Karmann, I., Sharp, W.D., Vuille, M., Cardoso, A.O., Ferrari, J.A., Dias,
702 P.L.S., Viana, O., 2005. Insolation-driven changes in atmospheric circulation over the past
703 116,000 years in subtropical Brazil. *Nature* 434, 63–66.
- 704 Cruz, F.W., Burns, S.J., Karmann, I., Sharp, W.D., Vuille, M., 2006. Reconstruction of regional
705 atmospheric circulation features during the late Pleistocene in subtropical Brazil from oxygen
706 isotope composition of speleothems. *Earth and Planetary Science Letters* 248, 495–507.
- 707 De Martonne, E., 1934. The Andes of the North-West Argentine. *Geographical Journal*, 1-14.

708 Denton, G.H., Anderson, R.F., Toggweiler, J.R., Edwards, R.L., Schaefer, J.M., Putnam, A.E.,
709 2010. The last glacial termination. *Science* 328, 1652-1656.

710 Desilets, D., Zreda, M., 2003. Spatial and temporal distribution of secondary cosmic-ray nucleon
711 intensities and applications to in situ cosmogenic dating. *Earth and Planetary Science Letters*
712 206, 21–42.

713 Desilets, D., Zreda, M., Prabu, T., 2006. Extended scaling factors for in situ cosmogenic
714 nuclides: New measurements at low latitude. *Earth and Planetary Science Letters* 246, 265–
715 276.

716 Dunai, T.J., 2001. Influence of secular variation of the geomagnetic field on production rates of
717 in situ produced cosmogenic nuclides. *Earth and Planetary Science Letters* 193, 197–212.

718 Fairbanks, R.G., 1989. A 17,000 year glacio-eustatic sea level record: influence of glacial
719 melting rates on the Younger Dryas event and deep ocean circulation. *Nature* 342, 637–642.

720 Farber, D., Hancock, G., Finkel, R., Rodbell, D., 2005. The age and extent of tropical alpine
721 glaciation in the Cordillera Blanca, Peru. *Journal of Quaternary Science* 20, 7- 8, 759–776,
722 doi: 10.1002/jqs.994.

723 Farr, T. G., Rosen, P. A., Caro, E., Crippen, R., Duren, R., Hensley, S., Kobrick, M., Paller, M.
724 Rodrigez, E., Roth, L., Seal, D., Shaffer, S., Shimada, J., Umlnd, J., Werner, M., Oskin., M.,
725 Burbank, D., Alsdorf, D., 2007. The shuttle radar topography mission. *Reviews of*
726 *geophysics* 45.

727 Gandu, A.W., Dias, P.L.S., 1998. Impact of tropical heat sources on the South American
728 tropospheric upper circulation and subsidence. *Journal of Geophysical Research* 103, 6001–
729 6015.

730 Garreaud, R.D., 2000. Intraseasonal variability of moisture and rainfall over the South American
 731 Altiplano. *Monthly Weather Review* 128, 3337–3346.

732 Garreaud, R.D., Aceituno, P., 2001. Interannual rainfall variability over the South American
 733 Altiplano. *Journal of Climate* 14, 2779–2789.

734 Gayo, E. M., Latorre, C., Jordan, T. E., Nester, P. L., Estay, S. A., Ojeda, K. F., Santoro, C. M.,
 735 2012. Late Quaternary hydrological and ecological changes in the hyperarid core of the
 736 northern Atacama Desert (~ 21 S). *Earth-Science Reviews* 113, 120-140.

737 Glasser, N. F., Clemmens, S., Schnabel, C., Fenton, C. R., McHargue, L., 2009. Tropical glacier
 738 fluctuations in the Cordillera Blanca, Peru between 12.5 and 7.6 ka from cosmogenic ¹⁰ Be
 739 dating. *Quaternary Science Reviews* 28, 3448-3458.

740 Gosse, J.C., Phillips, F.M., 2001. Terrestrial in situ cosmogenic nuclides: theory and application.
 741 *Quaternary Science Reviews* 20, 1475-1560.

742 Graf, K., 1991. Ein Modell zur eiszeitlichen und heutigen Vergletscherung in der bolivianischen
 743 Westkordillere. *Bamberger Geographische Schriften* 11, 139–154.

744 Grenon, M., 2007. Nature Around the ALMA Site-Part 1. *ESO: The Messenger* 59–63.

745 Grosjean, M., Van, L., JFN, Van, D.K., WO, Geyh, M.A., Ammann, B., Tanner, W., Messerli,
 746 B., Valero-Garcés, B.L., Veit, H., 2001. A 22,000 ¹⁴C year BP sediment and pollen record of
 747 climate change from Laguna Miscanti (23 S), northern Chile. *Global and Planetary Change*
 748 28, 35–51.

749 Guyodo, Y., Valet, J.-P., 1999. Global changes in intensity of the earth's magnetic field during
 750 the past 800 kyr. *Nature* 399, 249–252.

751 Hall, S. R., Farber, D. L., Ramage, J. M., Rodbell, D. T., Finkel, R. C., Smith, J. A., Mark, B.G.
 752 Kassel, C., 2009. Geochronology of Quaternary glaciations from the tropical Cordillera
 753 Huayhuash, Peru. *Quaternary Science Reviews* 28, 2991-3009.

754 Hartley, A.J., 2003. Andean uplift and climate change. *Journal of the Geological Society* 160, 7-
 755 10.

756 Heyman, J., 2014. Paleoglaciation of the Tibetan Plateau and surrounding mountains based on
 757 exposure ages and ELA depression estimates. *Quaternary Science Reviews* 91, 30-41.

758 Houston, J., 2006, Variability of precipitation in the Atacama Desert: its causes and hydrological
 759 impact. *International Journal of Climatology* 26, 2181–2198.

760 Jenny, B., Kammer, K., Ammann, C., 1996. Climate change in den trockenen Anden. Verlag des
 761 Geographischen Institutes der Universität Bern.

762 Jomelli, V., Favier, V., Vuille, M., Braucher, R., Martin, L., Blard, P.-H., Colose, C., Brunstein,
 763 D., He, F., Khodri, M., 2014. A major advance of tropical Andean glaciers during the
 764 Antarctic cold reversal. *Nature* 513, 224-228.

765 Jomelli, V., Khodri, M., Favier, V., Brunstein, D., Ledru, M.-P., Wagnon, P., Blard, P.-H.,
 766 Sicart, J.-E., Braucher, R., Grancher, D., 2011. Irregular tropical glacier retreat over the
 767 Holocene epoch driven by progressive warming. *Nature* 474, 196-199.

768 Kanner, L. C., Burns, S. J., Cheng, H., Edwards, R. L., 2012. High-latitude forcing of the South
 769 American summer monsoon during the last glacial. *Science* 335, 570-573.

770 Kaplan, M.R., Strelin, J.A., Schaefer, J.M., Denton, G.H., Finkel, R.C., Schwartz, R., Putnam,
 771 A.E., Vandergoes, M.J., Goehring, B.M., Travis, S.G., 2011. In-situ cosmogenic ^{10}Be
 772 production rate at Lago Argentino, Patagonia: Implications for late-glacial climate
 773 chronology. *Earth and Planetary Science Letters* 309, 21-32.

774 Kaplan, M., Moreno, P., Rojas, M., 2008. Glacial dynamics in southernmost South America
 775 during Marine Isotope Stage 5e to the Younger Dryas chron: a brief review with a focus on
 776 cosmogenic nuclide measurements, *J. Quaternary Sci.* 23, 649-658, doi:10.1002/jqs.1209.

777 Kelly, M., Lowell, T., Applegate, P., Smith, C., Phillips, F., Hudson, A., 2012. Late glacial
 778 fluctuations of Quelccaya Ice Cap, southeastern Peru. *Geology* 40, 991-994, doi:
 779 10.1130/G33430.1.

780 Kelly, M.A., Lowell, T.V., Applegate, P.J., Phillips, F.M., Schaefer, J.M., Smith, C.A., Kim, H.,
 781 Leonard, K.C., Hudson, A.M., 2015. A locally calibrated, late glacial Be-10 production rate
 782 from a low-latitude, high-altitude site in the Peruvian Andes. *Quaternary Geochronology*, 26,
 783 70-85.

784 Kohl, C. P., Nishiizumi, K., 1992. Chemical isolation of quartz for measurement of in-situ-
 785 produced cosmogenic nuclides. *Geochimica et Cosmochimica Acta* 56, 3583-3587.

786 Kull, C., Grosjean, M., 2000. Late Pleistocene climate conditions in the north Chilean Andes
 787 drawn from a climate--glacier model. *Journal of Glaciology* 46, 622–632.

788 Lal, D., 1991. Cosmic ray labeling of erosion surfaces: in situ nuclide production rates and
 789 erosion models. *Earth and Planetary Science Letters* 104, 424–439.

790 Lamy, F., Kaiser, J., Ninnemann, U., Hebbeln, D., Arz, H.W., Stoner, J., 2004. Antarctic timing
 791 of surface water changes off Chile and Patagonian ice sheet response. *Science* 304, 1959-
 792 1962.

793 Latorre, C., Betancourt, J.L., Arroyo, M.T.K., 2006. Late Quaternary vegetation and climate
 794 history of a perennial river canyon in the Rio Salado basin (22 S) of Northern Chile.
 795 *Quaternary Research* 65, 450-466.

796 Lenters, J.D., Cook, K.H., 1997. On the origin of the Bolivian high and related circulation
 797 features of the South American climate. *Journal of the Atmospheric Sciences* 54, 656–678.

798 Licciardi, J.M., Denoncourt, C.L., Finkel, R.C., 2008. Cosmogenic ^{36}Cl production rates from
 799 Ca spallation in Iceland. *Earth and Planetary Science Letters* 267, 365-377.

800 Licciardi, J. M., Schaefer, J. M., Taggart, J. R., Lund, D. C., 2009. Holocene glacier fluctuations
 801 in the Peruvian Andes indicate northern climate linkages. *Science*, 325, 1677-1679.

802 Lifton, N.A., Bieber, J.W., Clem, J.M., Duldig, M.L., Evenson, P., Humble, J.E., Pyle, R., 2005.
 803 Addressing solar modulation and long-term uncertainties in scaling secondary cosmic rays
 804 for in situ cosmogenic nuclide applications. *Earth and Planetary Science Letters* 239, 140–
 805 161.

806 Lifton, N., Sato, T., Dunai, T.J., 2014. Scaling in situ cosmogenic nuclide production rates using
 807 analytical approximations to atmospheric cosmic-ray fluxes. *Earth and Planetary Science*
 808 *Letters* 386, 149-160.

809 Lisiecki, L. E., Raymo, M. E., 2005. A Pliocene- Pleistocene stack of 57 globally distributed
 810 benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20.

811 Lowell, T.V., 1995. The application of radiocarbon age estimates to the dating of glacial
812 sequences: an example from the Miami sublobe, Ohio, USA. *Quaternary Science Reviews*
813 14, 85-99.

814 Maldonado, A., Betancourt, J. L., Latorre, C., Villagran, C., 2005. Pollen analyses from a
815 50000- yr rodent midden series in the southern Atacama Desert (25° 30' S). *Journal of*
816 *Quaternary Science*, 20, 493-507.

817 Mather, A.E., Hartley, A., 2005. Flow events on a hyper-arid alluvial fan: Quebrada Tambores,
818 Salar de Atacama, northern Chile. *Special Publication-Geological Society of London* 251, 9.

819 McCulloch, R. D., Bentley, M. J., Purves, R. S., Hulton, N. R. J., Sugden, D. E., Clapperton, C.
820 M., 2000. Climatic inferences from glacial and palaeoecological evidence at the last glacial
821 termination, southern South America. *Journal of Quaternary Science* 15, 409-417.

822 Monnin, E., Indermühle, A., Dällenbach, A., Flückiger, J., Stauffer, B., Stocker, T.F., Raynaud,
823 D., Barnola, J.-M., 2001. Atmospheric CO₂ concentrations over the last glacial termination.
824 *Science* 291, 112-114.

825 Moreno, P. I., M. R. Kaplan, J. P. François, Villa-Martínez, C. M. Moy, C. R. Stern, P. W.
826 Kubik, 2009. Renewed glacial activity during the Antarctic cold reversal and persistence of
827 cold conditions until 11.5 ka in southwestern Patagonia. *Geology* 37, 375-378,
828 doi:10.1130/G25399A.1.

829 Nishiizumi, K., Elmore, D., Ma, X. Z., Arnold, J. R., 1984. 10 Be and 36 Cl depth profiles in an
830 Apollo 15 drill core. *Earth and Planetary Science Letters* 70, 157-163.

- 831 Nishiizumi, K., Imamura, M., Caffee, M.W., Southon, J.R., Finkel, R.C., McAninch, J., 2007.
832 Absolute calibration of 10 Be AMS standards. Nuclear Instruments and Methods in Physics
833 Research Section B: Beam Interactions with Materials and Atoms 258, 403-413.
- 834 Ohno, M., Hamano, Y., 1992. Geomagnetic poles over the past 10,000 years. Geophysical
835 Research Letters 19, 1715–1718.
- 836 Phillips, F.M., Zreda, M.G., Flinsch, M.R., Elmore, D., Sharma, P., 1996. A reevaluation of
837 cosmogenic ³⁶Cl production rates in terrestrial rocks. Geophysical Research Letters 23, 949–
838 952.
- 839 Placzek, C. J., Quade, J., Patchett, P. J., 2013. A 130ka reconstruction of rainfall on the Bolivian
840 Altiplano. Earth and Planetary Science Letters 363, 97-108.
- 841 Putnam, A.E., Denton, G.H., Schaefer, J.M., Barrell, D.J.A., Andersen, B.G., Finkel, R.C.,
842 Schwartz, R., Doughty, A.M., Kaplan, M.R., Schlüchter, C., 2010a. Glacier advance in
843 southern middle-latitudes during the Antarctic Cold Reversal. Nature Geoscience 3, 700-704.
- 844 Putnam, A.E., Schaefer, J.M., Barrell, D.J.A., Vandergoes, M., Denton, G.H., Kaplan, M.R.,
845 Finkel, R.C., Schwartz, R., Goehring, B.M., Kelley, S.E., 2010b. In situ cosmogenic 10 Be
846 production-rate calibration from the Southern Alps, New Zealand. Quaternary
847 Geochronology 5, 392-409.
- 848 Quade, J., Rech, J.A., Betancourt, J.L., Latorre, C., Quade, B., Rylander, K.A., Fisher, T., 2008.
849 Paleowetlands and regional climate change in the central Atacama Desert, northern Chile.
850 Quaternary Research 69, 343–360.
- 851 Ramírez, C.F., Gardeweg, M., 1982. Hoja Toconau, Región de Antofagasta: Carta Geológica de
852 Chile, 1:250,000. Servicio Nacional de Geología y Minería de Chile, Boletín 54

- 853 Raymo, M. E., Oppo, D. W., Curry, W., 1997. The mid- Pleistocene climate transition: A deep
854 sea carbon isotopic perspective. *Paleoceanography* 12, 546-559.
- 855 Rech, J.A., Quade, J., Betancourt, J.L., 2002. Late Quaternary paleohydrology of the central
856 Atacama Desert (lat 22--24 S), Chile. *Geological Society of America Bulletin* 114, 334.
- 857 Risacher, F., Fritz, B., 2009. Origin of salts and brine evolution of Bolivian and Chilean salars.
858 *Aquatic Geochemistry* 15, 123-157.
- 859 Rodbell, D., Smith, J., Mark, B., 2009. Glaciation in the Andes during the Lateglacial and
860 Holocene. *Quaternary Science Reviews* 28, 2165-2212.
- 861 Rodwell, M.R., Hoskins, B.J., 2001. Subtropical anticyclones and monsoons. *J. Climate* 14,
862 3192–3211.
- 863 Roe, G.H., O’Neal, M.A., 2009. The response of glaciers to intrinsic climate variability:
864 observations and models of late-Holocene variations in the Pacific Northwest. *Journal of*
865 *Glaciology* 55, 839-854.
- 866 Rojas, M., Moreno, P., Kageyama, M., Crucifix, M., Hewitt, C., Abe-Ouchi, A., Ohgaito, R.,
867 Brady, E.C., Hope, P., 2009. The Southern Westerlies during the last glacial maximum in
868 PMIP2 simulations. *Climate Dynamics* 32, 525-548.
- 869 Sagredo, E.A., Lowell, T.V., 2012. Climatology of Andean glaciers: A framework to understand
870 glacier response to climate change. *Global and Planetary Change* 86, 101-109.
- 871 Sagredo, E.A., Rupper, S., Lowell, T.V., 2014. Sensitivities of the equilibrium line altitude to
872 temperature and precipitation changes along the Andes. *Quaternary Research* 81, 355-366.

873 Sato, T., Yasuda, H., Niita, K., Endo, A., Sihver, L., Lifton, N.A., 2008. Development of
874 PARMA: PHITS-based analytical radiation model in the atmosphere. *Radiation research* 170,
875 244–259-03.

876 Schmitt, A., de, S., S., Trumbull, R., Emmermann, R., 2001. Magma evolution in the Purico
877 ignimbrite complex, northern Chile: evidence for zoning of a dacitic magma by injection of
878 rhyolitic melts following mafic recharge. *Contributions to Mineralogy and Petrology* 140,
879 680–700.

880 Schimmelpfennig, I., Benedetti, L., Finkel, R., Pik, R., Blard, P. H., Bourles, D., Burnard, P.,
881 Williams, A., 2009. Sources of in-situ ^{36}Cl in basaltic rocks. Implications for calibration of
882 production rates. *Quaternary Geochronology* 4, 441-461.

883 Shackleton, N.J., 2000. The 100,000-year ice-age cycle identified and found to lag temperature,
884 carbon dioxide, and orbital eccentricity. *Science* 289, 1897–1902.

885 Smith, J. A., Seltzer, G. O., Farber, D. L., Rodbell, D. T., Finkel, R. C., 2005. Early local last
886 glacial maximum in the tropical Andes. *Science* 308, 678-681.

887 Smith, J. A., Mark, B. G., Rodbell, D. T., 2008. The timing and magnitude of mountain
888 glaciation in the tropical Andes. *Journal of Quaternary Science* 23, 609-634.

889 Smith, J. A., Rodbell, D. T., 2010. Cross-cutting moraines reveal evidence for North Atlantic
890 influence on glaciers in the tropical Andes. *Journal of Quaternary Science* 25, 243-248.

891 Smith, C. A., Lowell, T. V., Owen, L. A., Caffee, M. W., 2011. Late Quaternary glacial
892 chronology on Nevado Illimani, Bolivia, and the implications for paleoclimatic
893 reconstructions across the Andes. *Quaternary Research* 75, 1-10.

894 Stone, J. O., 2000. Air pressure and cosmogenic isotope production. *Journal of Geophysical*
895 *Research: Solid Earth*, 105(B10), 23753-23759.

896 Stone, J. O., Allan, G. L., Fifield, L. K., Cresswell, R. G., 1996. Cosmogenic chlorine-36 from
897 calcium spallation. *Geochimica et Cosmochimica Acta*, 60, 679-692.

898 Svensson, A., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Davies, S.M.,
899 Johnsen, S.J., Muscheler, R., Parrenin, F., Rasmussen, S.O., 2008. A 60 000 year Greenland
900 stratigraphic ice core chronology. *Climate of the Past* 4, 47-57.

901 Sylvestre, F., Servant, M., Servant-Vildary, S., Causse, C., Fournier, M., Ybert, J.-P., 1999.
902 Lake-level chronology on the Southern Bolivian Altiplano (18-23 S) during late-glacial time
903 and the early Holocene. *Quaternary Research* 51, 54-66.

904 Vuille, M., 1999. Atmospheric circulation over the Bolivian Altiplano during dry and wet
905 periods and extreme phases of the Southern Oscillation. *International Journal of Climatology*
906 19, 1579–1600.

907 Vuille, M., Hardy, D.R., Braun, C., Keimig, F., Bradley, R.S., 1998. Atmospheric circulation
908 anomalies associated with 1996/1997 summer precipitation events on Sajama Ice Cap,
909 Bolivia. *Journal of Geophysical Research* 103, 11191.

910 Ward, D.J., Anderson, R.S., Guido, Z.S., Briner, J.P., 2009. Numerical modeling of cosmogenic
911 deglaciation records, Front Range and San Juan mountains, Colorado. *Journal of Geophysical*
912 *Research* 114, F01026.

913 Wang, X., Auler, A.S., Edwards, R.L., Cheng, H., Ito, E., Solheid, M., 2006. Interhemispheric
914 anti-phasing of rainfall during the last glacial period. *Quaternary Science Reviews* 25, 3391-
915 3403.

916 Wang, X., Auler, A.S., Edwards, R.L., Cheng, H., Ito, E., Wang, Y., Kong, X., Solheid, M.,
 917 2007. Millennial-scale precipitation changes in southern Brazil over the past 90,000 years.
 918 *Geophysical Research Letters* 34.

919 Williams, P.W., King, D.N.T., Zhao, J.-X., Collerson, K.D., 2005. Late Pleistocene to Holocene
 920 composite speleothem $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ chronologies from South Island, New Zealand—Did a
 921 global Younger Dryas really exist? *Earth and Planetary Science Letters* 230, 301–317, doi:
 922 10.1016/j.epsl.2004.10.024.

923 Yang, S., Odah, H., Shaw, J., 2000. Variations in the geomagnetic dipole moment over the last
 924 12000 years. *Geophysical Journal International* 140, 158–162.

925 Zech, R., Kull, C., Veit, H., 2006. Late Quaternary glacial history in the Encierro Valley,
 926 northern Chile (29°S), deduced from ^{10}Be surface exposure dating. *Palaeogeography,*
 927 *Palaeoclimatology, Palaeoecology* 234, 277–286, doi: 10.1016/j.palaeo.2005.10.011.

928 Zech, R., Kull, C., Kubik, P. W., Veit, H., 2007a. LGM and Late Glacial glacier advances in the
 929 Cordillera Real and Cochabamba (Bolivia) deduced from ^{10}Be surface exposure dating.
 930 *Climate of the Past Discussions* 3, 839–869.

931 Zech, R., Kull, C., Kubik, P.W., Veit, H., 2007b. Exposure dating of Late Glacial and pre-LGM
 932 moraines in the Cordon de Doña Rosa, Northern/Central Chile (~ 31° S). *Climate of the Past*
 933 3, 1–14, doi: 10.5194/cp-3-1-2007.

934 Zech, R., May, J.H., Kull, C., Ilgner, J., Kubik, P.W., Veit, H., 2008. Timing of the late
 935 Quaternary glaciation in the Andes from ~15 to 40°S. *Journal of Quaternary Science* 23,
 936 635–647.

937 Zech, J., Zech, R., Kubik, P. W., Veit, H., 2009. Glacier and climate reconstruction at Tres
938 Lagunas, NW Argentina, based on ¹⁰Be surface exposure dating and lake sediment
939 analyses. *Palaeogeography, Palaeoclimatology, Palaeoecology* 284, 180-190.

940 Zweck, C., Zreda, M., Anderson, K., Bradley, E., 2012. The theoretical basis of ACE, an Age
941 Calculation Engine for cosmogenic nuclides. *Chemical Geology* 291, 199-205, doi:
942 10.1016/j.chemgeo.2011.10.005.

943

FIGURE CAPTIONS

Fig. 1. A: Location map of the study area in the subtropical Andes. Red dashed box indicates extent of Fig. 2; CP: Chajnantor Plateau. **B:** Modern climate features of South America. BH: Bolivian High; SASM: South American Summer Monsoon; ITCZ – Intertropical Convergence Zone (mean northern and southernmost annual positions after Cook et al., 2009 and Baker and Fritz, 2015). Chajnantor area is marked with a star. Glacial climate zones (colored areas) numbered as in Sagredo and Lowell (2012) and dominant sensitivity generalized from Sagredo et al. (2014): 1.1: Inner tropics (temperature-sensitive glaciers); 2.1, 2.2, and 3: Northern, southern, and dry outer tropics (precipitation- and temperature-sensitive); 4 and 5: Subtropics and northern mid-latitudes (precipitation-sensitive); 6.1, 6.2, and 7: Patagonia (temperature-sensitive); 1.2: Tierra del Fuego (precipitation- and temperature-sensitive).

Fig. 2. Regional distribution of moraines and drift limits (red lines). Chajnantor area shown in detail in Fig. 4; Co. Torta area is shown in Fig. 5.

Fig. 3. Age differences for a typical sample due to different production rate calibration datasets relevant to the central Andes. The new calibrations differ by $< 7\%$ across the relevant time range, similar to the $\sim 5\text{-}10\%$ analytical error on most ^{10}Be measurements. The Blard et al. (2013) calibration used in this study falls in the middle of the range of new calibrations.

Fig. 4. Field evidence for glaciation of the Chajnantor Plateau area. **A:** Polished, striated andesite near Co. Toco; dark erratic boulder is ~ 30 cm in diameter. **B:** Stage II moraine (low ridge in middle ground, ~ 100 m behind the large boulder) on the west edge of the plateau below Co. Negro. Backpack circled for scale. **C:** Nested 10-m-high Stage II and III moraines on the east side of the plateau. **D:** Stage I terminal and lateral moraines of the valley glacier near Co. Torta, 10-15 m high.

Fig. 5. Geomorphic and topographic map of the Chajnantor Plateau area. Topographic contours and hillshade are based on the 8-m DigitalGlobe DEM. Exposure ages (ka) as described below (black: ^{10}Be , gray: ^{36}Cl).

Fig. 6. Co. Torta area Landsat image with sketch map of moraine ridges and locations of ^{10}Be exposure ages. The moraine boulder was relatively small (< 50 cm) and weathered, so is taken as a minimum-limiting age.

Fig. 7. A: Correspondence between ^{10}Be and ^{36}Cl exposure ages of paired samples using a time-varying Lal/Stone scaling and the Phillips (1996) calibration dataset. Hollow symbols indicate samples $> 2\text{-}\sigma$ from the 1:1 line. **B:** Exposure ages for the four most reliable paired samples using different production rate calibration datasets for ^{36}Cl . Error bars omitted for clarity. For our field area, the calibrations of Phillips et al. (1996) and Licciardi et al. (2008) yield the same ages within a few hundred years, with differences $\ll$ uncertainty; the Swanson and Caffee (2001) calibration dataset gives ages lower by $\sim 12\%$, but still within $2\text{-}\sigma$ uncertainty. We use the Phillips (1996) calibration for ages reported in this work.

Fig. 8. Relative probability of cosmogenic ages given their respective uncertainties, separated into moraine boulder and bedrock samples, using the preferred exposure age for each sample as in Table 3. Note the time scale change between the moraine and bedrock panels. Shaded bands are our interpreted age ranges for each feature. See text for discussion.

Figure 9. Probability density plots from each rescaled study (numbered as in Table 2) with periods from Fig. 9 highlighted. For a period to be counted as a time of moraine stabilization, at least 3 samples had to form a peak or sub-peak in that period; periods containing ≤ 2 samples or samples with high uncertainty relative to others in the dataset were classified as “few/poor ages” in Fig. 10. A high-resolution version of this figure with expanded time axes is provided as Supplemental Fig. S1.

Fig. 11: Bedrock and moraine boulder ^{10}Be and ^{36}Cl ages from the Chajnantor Plateau. Hollow circles represent ages that may be excluded based on landform context. Sedimentary hydrologic proxy records from Salar de Atacama (Bobst et al., 2001) and Salar de Uyuni (Baker et al., 2001). NGRIP ice core $\delta^{18}\text{O}$ record from Svensson et al., 2008. Globally distributed benthic $\delta^{18}\text{O}$ records of Lisiecki and Raymo (2005), black, and regional benthic $\delta^{18}\text{O}$ from IODP core 677 off the Peruvian coast as described in Raymo et al. (1997), red.

Appendix 1. Laboratory methods and full sample data tables

A1.1. Sample preparation – ^{10}Be

The procedures to purify quartz from rock samples were modified after that of Kohl and Nishiizumi (1992), and the chemical preparation of AMS targets was based on the procedures of Nishiizumi et al. (1984) as modified for the University of Colorado and University of Cincinnati cosmogenic isotope labs. In general, approximately 1 kg of each sample was crushed and sieved to the 350-850 μ fraction, 100 grams of which were processed by heavy liquids separation and leaching by dilute HCl and HF/HNO₃ to yield pure quartz. 0.6 – 10 g aliquots of quartz were spiked with ~0.25 mg of SPEX CLBE2-2Y ^9Be carrier, then dissolved in concentrated HF. Dissolved samples were dried and treated with HClO₄ to remove excess fluorine, then eluted through low-pressure anion and cation exchange columns to separate Be from B, Ti, Fe, and other interfering ions. Ti was also selectively removed by HCl-HNO₃ titration to pH 4-5. Be hydroxide was precipitated using HCL-HNO₃ titration to pH 8, then transferred to quartz crucibles and oxidized in a muffle furnace. The Be oxide was packed with Nb powder into cathode sample holders for the AMS at Purdue Rare Isotope Measurement (PRIME) Lab, where $^{10}\text{Be}/^9\text{Be}$ ratios were determined. From these ratios concentrations of ^{10}Be were calculated for the original quartz samples.

A1.2. Sample preparation – ^{36}Cl

Extraction, chemical preparation, and sample dissolution for chlorine followed the procedures of Stone et. al. (1996b). Procedures were slightly modified to accelerate the

dissolution process (Radler, pers. comm.) and are outlined here. Samples were pulverized in a jaw crusher and sieved to 125 - 250 μ m. Densities were determined on representative sample pieces, ~2 cm diameter, using Archimedes principle. Two ~10g aliquots of the 125 - 250 μ m sieved whole rock fraction were retained for bulk rock geochemistry. Major and trace element compositions were analyzed by ICP-MS at Activation Laboratories Ltd in Ontario, Canada. Approximately 80g of the sieved fraction for each sample was leached overnight (12-16 hr.) in 70% HNO₃ (2 M) to remove contaminants and eliminate the meteoric ³⁶Cl component. The remaining solids were rinsed five times with 18 M Ω H₂O to remove the fines and dried overnight (12-16 hr.) at 85°C. After drying, roughly 30g of the leached material was dissolved in a 40% Low Chloride HF/70% HNO₃ (2 M) solution. During dissolution, the sample was spiked with ~1.0mg of enriched ³⁵Cl carrier. Following complete dissolution, solid fluoride compounds were isolated from the sample solution by centrifuging. AgCl was precipitated through the addition of HNO₃ (0.15 M) and allowed to accumulate overnight. The precipitate was collected by centrifuging and subsequently dissolved in 30% NH₄OH. Cl was extracted and impurities were removed through Anion Exchange Chromatography, and the final AgCl precipitate isolated through the addition of 0.1 M AgNO₃ and 70% HNO₃. Each sample yielded approximately 3-5mg of AgCl. The AgCl pellet was rinsed with 18 M Ω H₂O, dried overnight (12-16 hr.) at 70°C, and loaded into Cu targets packed with AgBr for accelerator mass spectrometry (AMS) analysis. ^{35/37}Cl and ^{36/37}Cl ratios were measured via AMS at the Purdue Rare Isotope Measurement Laboratory at Purdue University, Indiana.

1054

1055 *(Tables A1 and A2 included as Excel files.)*

1056

Table A1. 10Be sample data

Sample	Latitude deg	Longitude deg	Elevation m	Thickness cm	Density g/cm3	Shielding Factor	Erosion Rate cm/kyr	10Be at/g	10Be uncertainty at/g
Chaj-I2	-23.03424	-67.76790	4993	10.0	2.3	1	0	1.768E+06	1.061E+05
Chaj-I3	-23.02242	-67.82928	4585	5.0	2.3	1	0	9.915E+05	3.966E+04
Chaj-M1	-22.98725	-67.65284	4660	5.0	2.5	1	0	2.844E+06	1.138E+05
Chaj-M3	-22.94925	-67.68583	4802	10.0	2.5	1	0	5.024E+06	1.507E+05
Chaj-M5	-23.02380	-67.83113	4569	2.0	2.5	1	0	7.903E+06	1.581E+05
Chaj-P2	-22.96393	-67.80302	4931	2.0	2.5	1	0	1.513E+06	9.077E+04
TAT-3	-22.45008	-67.99693	4405	5.0	2.3	0.998	0	1.415E+06	5.662E+04
TAT-4	-22.45470	-67.99914	4533	5.0	2.3	0.998	0	1.157E+06	4.626E+04
Chaj-12-08	-22.96808	-67.66564	4702	5.0	2.5	1	0	3.9712E+05	1.2192E+05
Chaj-12-18	-23.02044	-67.83226	4564	5.0	2.3	1	0	2.3556E+06	3.5098E+05
Chaj-12-19	-23.01647	-67.81591	4719	5.0	2.3	1	0	9.1163E+05	1.3036E+05
Chaj-12-21	-23.01327	-67.81017	4787	5.0	2.3	1	0	5.3622E+06	2.7186E+06
Chaj-12-03	-22.91785	-67.76319	4784	5.0	2.5	1	0	2.1659E+06	8.4472E+04
Chaj-12-04	-22.91787	-67.76275	4782	5.0	2.5	1	0	1.3267E+06	4.2456E+04

Appendix Table A2

Table A2. 36Cl sample data																												
Sample	Latitude deg	Longitude deg	Elevation m	Thickness cm	Density g/cm3	Shielding Factor	Erosion Rate cm/kyr	Total Cl ppm	Cl uncertainty ppm	36Cl Ratio 10 ⁻¹⁵	Ratio Uncertainty 10 ⁻¹⁵	Al2O3 wt %	CaO wt %	Fe2O3 wt %	K2O wt %	MgO wt %	MnO wt %	Na2O wt %	P2O5 wt %	SiO2 wt %	TiO2 wt %	B ppm	Gd ppm	Sm ppm	Th ppm	U ppm	CO2 (as 0.5 * LOI) wt %	Volumetric Water (as 0.5 * LOI)
Chaj-12-03	-22.91785	-67.76319	4784	5	2.5	1	0.0	133.44	15.31	2216.380	91.856	16.08	5.11	6.97	2.33	3.59	0.088	3.16	0.07	61.9	0.874	19.6	2.3	3.4	11.3	3.5	0.215	0.005375
Chaj-12-04	-22.91787	-67.76275	4782	5	2.5	1	0.0	176.98	26.34	1336.530	49.867	16.39	5.14	6.01	2.36	2.66	0.075	3.24	0.06	62.76	0.725	24.6	2.5	3.3	12.9	4.1	0.29	0.00725
Chaj-12-06	-22.98889	-67.64395	4686	5	2.5	1	0.0	448.43	91.88	4933.290	145.779	14.25	4.91	6.67	2.49	3.02	0.104	2.71	0.09	62.37	0.617	38.2	3.1	3.6	17.1	7	1.145	0.028625
Chaj-12-08	-22.96808	-67.66564	4702	5	2.5	1	0.0	279.80	70.84	1899.670	69.866	15.17	4.38	5.58	2.47	2	0.071	3.07	0.08	63.12	0.653	23.7	3	4.1	15.3	5.9	0.7	0.0175
Chaj-12-11	-22.96558	-67.79036	5036	5	2.5	1	0.0	70.75	2.32	2592.930	64.599	16.84	5.71	6.46	2.16	2.84	0.077	3.31	0.03	59.53	0.772	14.4	2	3.1	11.5	3.6	0.265	0.006625
Chaj-12-12	-22.96809	-67.79295	5025	5	2.5	1	0.0	47.02	1.38	731.102	19.439	16.78	5.15	6.39	2.3	3.06	0.087	3.18	0.01	62.69	0.784	13.9	1.9	2.8	11.4	3.8	0.18	0.0045
Chaj-12-14	-22.951	-67.81497	4706	5	2.5	1	0.0	479.44	54.14	3647.480	91.950	15.26	4.64	5.84	2.51	2.52	0.079	3.07	0.1	61.78	0.751	27.3	2.8	4.1	13.6	4.2	0.835	0.020875
Chaj-12-15	-22.95102	-67.81488	4704	5	2.5	1	0.0	94.94	6.77	2073.030	43.473	15.25	4.64	6.03	2.49	3	0.089	2.83	0.06	63.17	0.759	27.3	2.5	3.5	12.9	5	0.545	0.013625
Chaj-12-18	-23.02044	-67.83226	4564	5	2.3	1	0.0	93.67	5.62	2625.590	60.105	15.47	4.32	4.86	2.67	1.22	0.055	3.2	0.06	64.59	0.587	32.1	2.3	2.8	14.5	6.1	0.435	0.010875
Chaj-12-19	-23.01647	-67.81591	4719	5	2.3	1	0.0	85.26	8.06	870.174	19.795	14.94	4.5	4.99	2.34	1.24	0.058	3.15	0.01	64.68	0.591	20.1	2	3	12.8	5.8	0.415	0.010375
Chaj-12-20	-23.01366	-67.81439	4758	5	2.3	1	0.0	117.77	6.59	1266.590	34.929	15.8	4.58	4.93	2.73	1.4	0.056	3.23	0.04	66.1	0.629	25.3	2.4	2.9	13.6	6.2	0.72	0.018
Chaj-12-21	-23.01327	-67.81017	4787	5	2.3	1	0.0	97.40	3.53	1038.020	26.026	15.56	4.54	4.97	2.33	1.23	0.061	3.1	0.03	64.71	0.624	27.8	2.3	3.1	12.8	5.7	0.36	0.009
Chaj-M2	-22.95012	-67.68478	4796	5	2.5	1	0.0	347.20	92.79	2145.360	73.829	16.77	5.92	6.8	2.08	3.2	0.101	3.04	0.09	60.78	0.791	21.8	2.9	3.4	10	4.2	0.285	0.007125

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11