LAST LOCAL GLACIAL MAXIMUM AND DEGLACIATION OF THE ANDEAN CENTRAL VOLCANIC ZONE: THE CASE OF HUALCAHUALCA VOLCANO AND PATAPAMPA ALTIPLANO (SOUTHERN PERU)

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ABSTRACT. The aim of this study is to constrain the timing of the deglaciation process since the Last Local Glacial Maximum in HualcaHualca volcano and Patapampa Altiplano, located in the Andean Central Volcanic Zone. Nine ³⁶Cl cosmogenic surface exposure dating of moraine boulders as well as polished and striated bedrock surfaces are presented. The ³⁶Cl cosmogenic exposure ages indicate that the glaciers reached their maximum extent at ~17-16 ka on the HualcaHualca volcano during the Heinrich 1 event and the Tauca paleolake cycle. Since then glaciers began to retreat until ~12 ka, when they went through a phase of readvance or stillstand. The deglaciation of HualcaHualca was constant since ~11.5 ka, coinciding with the disappearance of the ice cap from the Patapampa Altiplano. These glacial ages do not corroborate a Last Local Glacial Maximum prior to the global Last Glacial Maximum but they indicate a sensitive reaction of the glacier system to precipitation fluctuations. According to the analysis of cosmogenic exposure ages reported from HualcaHualca, Sajama and Tunupa volcanoes, the onset of deglaciation since Last Local Glacial Maximum occurred at the end of the Heinrich 1 event and the Tauca paleolake cycle in the Andean Central Volcanic Zone. However, the glacier retreat was not continuous because at least one significant readvance or stillstand phase has been reported in most of the volcanoes studied in this region although the ages cannot be clearly related to the Younger Dryas and/or the Antarctic Cold Reversal cold events. After this readvance or stillstand, the glaciers of the Central Volcanic Zone retreated, but at least three clear minor readvances evidence a not homogeneous warm and/or dry climate during the Holocene. Even though in situ cosmogenic exposure provides important glacial chronological data, it is difficult to establish a consistent regional glacial reconstruction and clear connections with the main Late Pleistocene cold episodes due to limitations associated with in situ cosmogenic production rates and the use of different scaling schemes. To reduce the uncertainty and compare...
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RESUMEN. El objetivo de este trabajo es conocer cuándo comenzó el retroceso de los glaciares desde el Último Máximo Glaciar Local tanto en el volcán HualcaHualca y el Altiplano de Patapampa, ambos localizados al sur de Perú, como en la Zona Volcánica Centroandina. Para ello se presentan 9 edades de exposición a la radiación cósmica procedentes de morrenas y umbrales rocosos pulidos y estriados. Dichas edades indican que los glaciares del HualcaHualca alcanzaron su máxima extensión hace ~17-16 ka en sincronía con el Heinrich 1 y la formación del paleolago Tauca. Desde entonces los glaciares empezaron a retroceder hasta ~12 ka cuando experimentaron un reavance o una fase de estabilización. La deglaciación fue constante desde hace ~11.5 ka en el HualcaHualca, coincidiendo con la desaparición del casquete de Hielo de Patapampa. Esta evolución de los glaciares del área de estudio no corrobora un Último Máximo Glaciar Local más antiguo que el Último Máximo Glaciar global pero sí indica su elevada sensibilidad a los cambios en la precipitación. De acuerdo con el análisis de las edades de exposición a la radiación cósmica de los volcanes HualcaHualca, Sajama y Tunupa, se infiere que el inicio de la deglaciación en la Zona Volcánica Centroandina tuvo lugar al finalizar el evento Heinrich 1 y la fase lacustre Tauca. Sin embargo, el retroceso no fue continuo ya que se registra al menos un reavance o fase de estabilización en la mayoría de los volcanes estudiados aunque la inconsistencia entre sus edades no permite relacionar con claridad dicha fase glaciar con los eventos climáticos fríos Younger Dryas y Antarctic Cold Reversal. Después, las masas de hielo de la Zona Volcánica Centroandina experimentaron un marcado retroceso, interrumpido temporalmente por al menos tres reavances o periodos de estabilización de menor entidad que evidencian que el clima durante el Holoceno no fue continuamente cálido y/o seco. Por último cabe destacar que a pesar de que las edades de exposición a la radiación cósmica proporcionan una información cronológica valiosa, de momento no permiten reconstruir de forma sólida la historia de los glaciares y establecer conexiones claras con los principales eventos fríos debido a las limitaciones que presentan las tasas de producción y el uso de diferentes modelos de escala. Con el objeto de reducir la incertidumbre derivada de estas limitaciones, sería necesario determinar una tasa de producción precisa de cada isótopo cosmogénico en los Andes Centrales, un modelo de escala de referencia y recalcular las edades publicadas.

Key words: Cosmogenic Surface Exposure dating, Tropical Glaciation, Last Glacial Maximum, Younger Dryas, Antarctic Cold Reversal, Holocene, Andean Central Volcanic Zone, Southern Peru.
1. Introduction

The current knowledge about the glacial history in the Central Andes is focused on dating the Last Local Glacial Maximum (LLGM) and the subsequent Late Glacial readvances (Rodbell, 1993; Clapperton, 1993; Clayton and Clapperton, 1997; Mark et al., 2002; Seltzer et al., 2000, 2002; Farber et al., 2005; Smith et al., 2005, 2009, 2011; Zech et al., 2007a, 2007b; Kull et al., 2008; Bromley et al., 2009, 2011; Glasser et al., 2009; Hall et al., 2009; May et al., 2011; Ubeda et al., 2012; Kelly et al., 2012; Blard et al., 2013, 2014; Stansell et al., 2015). However, our understanding of the LLGM, the readvance or stillstand phases and the beginning of deglaciation is limited in regions such as the Andean Central Volcanic Zone. This scenario does not allow verification of traditional hypotheses of glacier behaviour in the Central Andes as: (i) the LLGM took place several thousand years before the global Last Glacial Maximum (Smith et al., 2005, 2008); (ii) ice masses located in arid regions react to changes in precipitation while glaciers in areas with high levels of precipitation are more sensitive to temperature fluctuations, as proposed by Hastenrath (1971), Klein et al. (1999), Amman et al. (2001) and Sagredo and Lowell (2012).

The most detailed glacial chronology since the LLGM in the Andean Central Volcanic Zone has been obtained at Coropuna volcano. Based on the application of in-situ cosmogenic $^3$He surface exposure dating and Lm scaling (Lal, 1991; Stone, 2000; Nishiizumi et al., 1989) on the outer moraines of Coropuna, Bromley et al. (2009) suggest that the LLGM took place between ~15 and 25 ka. This chronology is consistent with the global Last Glacial Maximum (19-26 ka; Clark et al., 2009), but the mean age of LLGM moraines show significant changes when the different scaling schemes available are used: 20.95 ± 0.2 ka (Lm), 17.46 ± 0.2 ka (Du; Dunai, 2000); 17.18 ± 0.2 ka (Li; Lifton et al., 2005) and 16.85 ± 0.2 ka (De; Desilets et al., 2006). Deglaciation since LLGM on Coropuna began at ~19 ka until ~15 ka (Lm scaling) according to the $^3$He exposure ages from a valley-floor transect in Quebrada Sigue Chico, located on the western flank of the volcano. This post-LLGM recession was followed by a strong readvance dated at ~13 ka, although after that the retreat continued until it reached its modern extent. Similar glacier behaviour occurred on the Quebrada Santiago, located on the northeastern slope of Coropuna (Bromley et al., 2011). Nevertheless, deglaciation and readvance chronologies are ~3 ka younger if the other three scaling models (Du, Li and De) are used (Bromley et al., 2009).
By contrast, Úbeda et al. (2012) used in-situ cosmogenic $^{36}$Cl isotope to date the outer moraines on the northeastern flank of Coropuna volcano. The $^{36}$Cl exposure ages, following the Lal (1991) scaling model, suggest that glaciers reached their maximum extent between ~20 and 16 ka. Then the ice mass began to retreat but it experience a readvance at ~12-11 ka. However, the LLGM on the southern slope was dated at ~14 ka as well as one readvance or stillstand pulse at ~10-9 ka. Other minor readvances at ~9 ka (northern slope) and ~6 ka (southern slope) have been reported. These minor glacier expansions interrupted the deglaciation process that occurred during the Holocene.

In Bolivia, at Nevado Sajama, the outermost moraines were also dated with in-situ cosmogenic $^{36}$Cl isotope although the Stone (2000) scaling method was used (Smith et al., 2009). $^{36}$Cl exposure ages indicates that the LLGM occurred between ~16.9-11.8 ka and ~14.0-10.2 ka. The glacier began to retreat after this maximum advance, but glacial deposits have been dated at ~7.0 and ~3.3 ka, suggesting the lack of a homogeneous warm or/and dry climate during the Holocene.

The moraine record of Tunupa volcano (Bolivia) is another site where in-situ cosmogenic $^3$He surface exposure dating was used (Blard et al., 2013). Here the exposure ages were calculated though a new and local production rate obtained by Blard et al. (2013) from a fluvio-glacial outwash deposit located on the southern flank of the volcano. Moreover, they apply the scaling models of Lal (1991) and Stone (2000). The cosmogenic $^3$He ages indicates that the LLGM took place between ~17-15.5 Ka, in synchronicity with the Lake Tauca paleocycle (17-15 ka). A recessional moraine has also been dated at ~14.7 ka but this chronology is not consistent with the ages from polished and striated bedrock (~400 years older). Thus, Blard et al. (2013) proposed that the onset of glacial recession occurred between ~15.5-14.5 ka, approximately 500 years before the end of Tauca paleolake regression. Both deglaciation and paleolake regression have been related with the Bølling Allerød warm climate event.

Another glacial chronology has also been established on Uturuncu volcano, approximately 270 km to the south of Tunupa volcano, based on $^3$He surface exposure dating. These ages were constrained using the cosmogenic production rate calibrated in the Tunupa volcano as well as the scaling schemes of Lal (1991) and Stone (2000). Thus, the outer moraine was built between 65 and 37 ka, revealing that the LLGM is significantly older than the global Last Glacial Maximum although glaciers remained close to their maximum position until ~18 ka. At this point the ice mass began to retreat but a readvance or stillstand occurred around 16-14 ka, being in phase with the highest level of paleolake Tauca. After 14 ka the glacier retreated, a process that coincides with the Bølling Allerød interstadial (Blard et al., 2014).

To improve the knowledge of the evolution of glaciers since the LLGM in the Central Volcanic Zone, this study present nine $^{36}$Cl surface exposure ages of moraine boulders as well as polished and striated bedrock outcrops from HualcaHualca volcano and the Patapampa Altiplano. This glacial chronology also can be useful to infer if the ice masses of the Central Volcanic Zone are sensitive to climatic parameters such as temperature or precipitation or both.
2. Geographical setting

The Andes is the longest continental mountain range extending over 7500 km from the Caribbean coast of Venezuela to Cabo de Hornos in Chile. According to Clapperton (1993) there are three main units: the northern Andes between the Caribbean Sea and the Amotape suture zone; the central Andes, from the Amotape suture zone to the south of Chile; and the Southern Andes, from the south of Chile to the Shackleton-Seotra region. The distinctive characteristic of these regions is the alternation of active and inactive volcanic zones across the whole orogen.

The Central Andean Volcanic Zone (Fig. 1), associated to the subduction of the oceanic Nazca Plate beneath the South American Plate, with a seafloor dip of >25°, presents at least 6 caldera systems and 44 active volcanic centers (Isacks, 1988; Stern, 2004). One of them is the Ampato volcanic complex (15°24′ - 15°51´S / 71°51´- 73°W; 6288 m a.s.l.), located 70 km to the northwest of Arequipa city. HualcaHualca (6025 m a.s.l.) is the northernmost stratovolcano of the Ampato complex and it is considered extinct (Thouret et al., 2005). The formation of the HualcaHualca volcano began during the late Miocene or the early Pleistocene and it is compound of andesitic materials. As a consequence of the collapse of the northern flank, the HualcaHualca exhibits a horseshoe-shaped caldera.

HualcaHualca is characterised by a well-preserved moraine record, especially in the four valleys selected for this study: Huayuray (northern slope), Pujro Huayjo (southwestern slope), Mollebaya (eastern slope) and Mucurca (western slope). Modern glaciers cover the summit but they are experiencing a marked retreat in recent decades (Alcalá et al., 2010, Alcalá, 2015). The other study area, Patapampa

Figure 1. Location of the Andean Central Volcanic Zone and the volcanoes where the glacial record has been dated through in situ cosmogenic Surface isotopes.
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(15°43´40´´-15°46´ S / 71°43´30´´-71°36´ W; 4940 m a.s.l.), is an Altiplano situated 17 km southeast of HualcaHualca volcano that presents geomorphological evidences of glacial activity (Fig. 2).

Figure 2. Location of the study area: HualcaHualca volcano and Patapampa Altiplano.

The climate of the study area is mainly determined by seasonal changes in the Intertropical Convergence Zone that generate two marked seasons: the wet season (December to March), when humid air masses arriving the Central Andean Altiplano from the Atlantic ocean produce 70-90 % of the annual precipitation (800-1000 mm) (Dornbusch, 1998; Herreros et al., 2009), and the dry season (April to November) due to the strong influence of high pressure conditions. However, temperatures do not change significantly throughout the year.

3. Methods

To determine the glacier evolution since the LLGM and the pattern of deglaciation on HualcaHualca volcano, the in situ cosmogenic nuclide $^{36}\text{Cl}$ was measured. This cosmogenic isotope was selected because it permits dating quartz-free rocks from the
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glacial deposits and glacially-abraded bedrock of the study area. Moreover, the use of the cosmogenic $^{36}$Cl allows comparison with the other glacial areas of the Central Volcanic Zone where in situ cosmogenic nuclides have been applied.

3.1. Cosmogenic $^{36}$Cl surface exposure dating

3.1.1. Sampling strategy and lab protocol: measure of $^{36}$Cl concentration

A detailed geomorphological map at 1:20,000 scale was elaborated to represent glacial landforms of HualcaHualca volcano (Alcalá-Reygosa et al., 2016). The map was used to select the sites containing well-preserved lateral and frontal moraines corresponding to the maximum glacier extent, readvance phases and glacially-abraded bedrock for cosmogenic $^{36}$Cl surface exposure dating.

To constrain the cosmogenic $^{36}$Cl exposure age of the moraines and glacially abraded bedrock, nine samples were collected with hammer and chisel from solid rock surfaces (~5 cm). Seven of them (Hualca 1, 2, 3, 4; Pujro Huayjo 1; Mucurca 1; Patapampa 3) were taken from the surface of andesite boulders >1 m high, situated on the crests of moraines. The other two samples (Pujro Huayjo 2 and Patapampa 4) were collected from the highest surfaces of polished and striated bedrock outcrops. This method succeeded in reducing the potential uncertainty in $^{36}$Cl cosmogenic production caused by post geomorphological processes (erosion, weathering, toppling) as well as shielding by snow, tephra layers associated with volcanic activity of the active Sabancaya volcano and soils.

3.1.2. Lab protocol

Lichens, mosses and other organic material were removed from the samples with a brush. Then, the samples were crushed using a roller grinder and sieved to retrieve the sand size fraction (500-850 µm) in the Complutense University of Madrid (Spain). Whole-rock protocol (Zreda et al., 1999; Phillips, 2003) was applied for in situ $^{36}$Cl dating at the PRIME Laboratory (Purdue University). There, the sand fraction size was leached in deionized water and HNO$_3$ to remove atmospheric Cl and then dissolved in HNO$_3$ and HF acids. A spike of isotopically enriched $^{35}$Cl was added during the dissolution process. The isotope dilution method allowed the $^{36}$Cl and total Cl to be measured simultaneously (Desilets et al., 2006). The ratios $^{36}$Cl/Cl and $^{37}$Cl/$^{35}$Cl were determined by Accelerator Mass Spectrometry (AMS) analysis at PRIME Laboratory.

Aliquots of bulk rock (pre-treated) and target fraction (post-treated) were analysed at Activation Laboratories (Ancaster, Canada) to measure: (i) major elements, by fusion inductively coupled plasma optical emission spectrometry (ICP-OES); (ii) trace elements, by inductively coupled plasma mass spectrometry (ICP-MS); and (iii) boron, by prompt-gamma neutron activation analysis (PGNAA). The field and analytical data are presented in Table 1.

3.1.3. In-situ cosmogenic $^{36}$Cl exposure age calculations

In-situ cosmogenic $^{36}$Cl exposure ages were calculated using the spreadsheet developed by Schimmelpfennig (2009) and Schimmelpfennig et al. (2009). Several production rates of cosmogenic $^{36}$Cl from the spallation of Ca were used: 48.8 ± 3.4 atoms
Table 1. Field and analytical data for 36Cl samples from Hualca volcano and Patapampa Altiplano.

| Sample ID | Hualca 1 | Hualca 2 | Hualca 3 | Hualca 4 | Pujro Huayjo 1 | Pujro Huayjo 2 | Patapampa 3 | Patapampa 4 | Manoera 1 |
|-----------|----------|----------|----------|----------|---------------|---------------|-------------|-------------|-----------|
| Latitude  | (°S)     | -15.74   | -15.67   | -15.68   | -15.64        | -15.82        | -15.82      | -15.76      | -15.74    | -15.75    |
| Longitude | (°W)     | -71.76   | -71.85   | -71.85   | -71.85        | -71.95        | -71.96      | -71.63      | -71.64    | -71.98    |
| Elevation | (masl)   | 4444     | 4408     | 4512     | 4144          | 4521          | 4450        | 4671        | 4886      | 4460      |
| Sample thickness | (cm) | 3.0      | 0.8      | 1.5      | 2.0           | 3.0           | 3.0         | 2.0         | 1.0       | 3.0       |
| Shielding factor incorporating all effects | (unitless) | 0.97     | 0.99     | 0.99     | 0.97          | 0.99          | 0.99        | 0.998       | 1.0       | 0.99      |
| Snow shielding factor | (unitless) | 1.0      | 1.0      | 1.0      | 1.0           | 1.0           | 1.0         | 1.0         | 1.0       | 1.0       |
| Scaling factor for nucleonic production | (unitless) | 10.63    | 10.45    | 10.93    | 9.24          | 11.0          | 10.66       | 11.76       | 129.0     | 10.7      |
| Scaling factor for muonic production | (unitless) | 3.84     | 3.8      | 3.92     | 3.48          | 3.93          | 3.85        | 4.13        | 4.4       | 3.86      |
| Effective fast neutron attenuation length | (g cm−2) | 160.0    | 160.0    | 160.0    | 160.0         | 160.0         | 160.0       | 160.0       | 160.0     | 160.0     |
| Na2O      | (wt.%)   | 3.76     | 3.72     | 3.92     | 4.37          | 4.22          | 3.49        | 4.01        | 3.87      | 4.26      |
| MgO       | (wt.%)   | 2.45     | 2.46     | 1.79     | 1.45          | 2.11          | 5.76        | 1.14        | 1.15      | 2.07      |
| Al2O3     | (wt.%)   | 15.52    | 15.33    | 15.38    | 16.16         | 15.67         | 16.49       | 18.09       | 17.03     | 15.55     |
| SiO2      | (wt.%)   | 61.50    | 62.32    | 62.62    | 63.90         | 62.45         | 53.49       | 62.94       | 62.37     | 62.68     |
| P2O5      | (wt.%)   | 0.39     | 0.56     | 0.39     | 0.40          | 0.10          | 0.07        | 0.15        | 0.24      | 0.04      |
| K2O       | (wt.%)   | 3.19     | 2.84     | 3.67     | 3.99          | 3.24          | 1.59        | 3.30        | 3.49      | 3.20      |
| CaO       | (wt.%)   | 4.66     | 4.57     | 3.85     | 3.39          | 4.66          | 6.37        | 4.51        | 4.66      | 4.66      |
| TiO2      | (wt.%)   | 0.96     | 0.99     | 0.83     | 0.72          | 0.96          | 1.18        | 0.83        | 0.96      | 0.96      |
| MnO       | (wt.%)   | 0.08     | 0.08     | 0.06     | 0.05          | 0.07          | 0.11        | 0.05        | 0.06      | 0.07      |
| Fe2O3     | (wt.%)   | 5.86     | 6.31     | 5.12     | 4.38          | 5.86          | 9.88        | 4.25        | 5.86      | 5.86      |
| Cl        | (ppm)    | 525.1    | 81.1     | 228.2    | 129.7         | 59.4          | 101.1       | 273.34      | 42.2      | 27.8      |
| B         | (ppm)    | 11.9     | 9.8      | 21.0     | 19.8          | 13.0          | 4.3         | 28.0        | 22.1      | 17.1      |
| Sm        | (ppm)    | 6.2      | 5.8      | 5.4      | 5.2           | 6.0           | 4.8         | 4.7         | 6.4       | 6.4       |
| Gd        | (ppm)    | 4.5      | 3.8      | 3.9      | 3.4           | 3.8           | 3.6         | 3.3         | 4.6       | 3.9       |
| U         | (ppm)    | 1.5      | 1.5      | 2.4      | 2.9           | 2.2           | 0.7         | 3.3         | 3.2       | 2.1       |
| Th        | (ppm)    | 8.8      | 8.0      | 13.1     | 16.5          | 10.6          | 3.0         | 14.2        | 16.8      | 10.4      |
| Sample mass | (g) | 30.15    | 30.19    | 30.18    | 30.10         | 30.10         | 30.57       | 30.15       | 30.16     | 30.23     |
| Mass of 35Cl spike solution | (mg) | 1.01     | 1.01     | 1.01     | 0.988         | 1.04          | 1.04        | 1.14        | 1.03      | 1.03      |
| Concentration Spike solution | (g g−1) | 1.0      | 1.0      | 1.0      | 1.0           | 1.0           | 1.0         | 1.0         | 1.0       | 1.00      |
| Analytical stable isotope ratio | (36Cl/(35Cl + 37Cl)) | 3.39 ± 0.01 | 4.78 ± 0.03 | 3.729 ± 0.02 | 4.153 ± 0.02 | 5.476 ± 0.38 | 4.49 ± 0.3 | 3.68 ± 0.04 | 6.21 ± 0.04 | 7.88 ± 0.47 |
| Analytical 36Cl/Cl ratio | (36Cl/1015 Cl) | 301.2 ± 7.4 | 726.7 ± 16.11 | 357.6 ± 6.28 | 543.6 ± 11.58 | 504.4 ± 18.57 | 425.8 ± 18.25 | 310.53 ± 15.80 | 906.2 ± 27.72 | 554.2 ± 24.58 |
| Corrected 36Cl concentration | atoms per gram of rock | 2856217.0 | 1429192.0 | 1589940.0 | 1507711.00 | 806626.00 | 975356.00 | 1662912.91 | 1182168.0 | 581522.00 |
$^{36}$Cl (g Ca)$^{-1}$ a$^{-1}$ (Stone et al. 1996), 42.2 ± 4.8 atoms $^{36}$Cl (g Ca)$^{-1}$ a$^{-1}$ (Schimmelpfennig et al., 2011), 56.27 $^{36}$Cl (g Ca)$^{-1}$ a$^{-1}$ (Borchers et al., 2016) and 56.0 ± 4.1 (g Ca)$^{-1}$ a$^{-1}$ (Marrero et al., 2016).

$^{36}$Cl production rates from the spallation of K are as follows: 148.1 ± 7.8 atoms $^{36}$Cl (g K)$^{-1}$ a$^{-1}$ obtained by Schimmelpfennig et al. (2014), 156.09 atoms $^{36}$Cl (g K)$^{-1}$ by Borchers et al. (2016) and 155 ± 11 atoms $^{36}$Cl (g K)$^{-1}$ by Marrero et al. (2016) were used. Moreover production rates from spallation of Ti of 13 ± 3 atoms $^{36}$Cl (g Ti)$^{-1}$ a$^{-1}$ by Fink et al. (2000) and from the spallation of Fe of 1.9 atoms $^{36}$Cl (g Fe)$^{-1}$ a$^{-1}$ by Stone et al. (2005) were introduced. The production rate of epithermal neutrons from fast neutrons in the atmosphere at the land/atmosphere interface (626 ± 46 neutrons (g air)$^{-1}$ a$^{-1}$; 696 ± 185 neutrons (g air)$^{-1}$ a$^{-1}$; 759 ± 180) proposed by Phillips et al. (2001) and Marrero et al. (2016) were used. The ages derived from the production rates proposed by Marrero et al. (2016) were selected because they were obtained at similar latitude and elevation (Huancané, Peru) than HualcaHualca volcano and Patapampa Altiplano. However, results for the available production rates are presented here to show differences (Table 2). The elevation/latitude scaling factors for nucleonic and muonic production were established using CosmoCalc (Vermeesch, 2007), which is based on the scaling model of Stone (2000). The shielding factor was estimated using the Topographic Shielding Calculator v1.0 provided by the CRONUS-Earth Project (2014).

Table 2. Cosmogenic $^{36}$Cl surface exposure ages from HualcaHualca volcano and Patapampa Altiplano. Results for the available $^{36}$Cl production rates. (1) Spallation of Ca: 56.27 $^{36}$Cl (g Ca)$^{-1}$ a$^{-1}$ (Borchers et al., 2016); Spallation of K: 156.09 atoms $^{36}$Cl (g K)$^{-1}$; Production rate of epithermal neutrons from fast neutrons in the atmosphere at the land/atmosphere interface: 626 ± 46 neutrons (g air)$^{-1}$ a$^{-1}$ (Phillips et al., 2001). (2) Spallation of Ca: 56.0 ± 4.1 (g Ca)$^{-1}$ a$^{-1}$ (Marrero et al., 2016); 155 ± 11 atoms $^{36}$Cl (g K)$^{-1}$ (Marrero et al., 2016); Production rate of epithermal neutrons from fast neutrons in the atmosphere at the land/atmosphere interface: 759 ± 180 neutrons (g air)$^{-1}$ a$^{-1}$ and 696 ± 185 neutrons (g air)$^{-1}$ a$^{-1}$ (Marrero et al., 2016). (3). Spallation of Ca: 42.2 ± 4.8 atoms $^{36}$Cl (g Ca)$^{-1}$ a$^{-1}$ (Schimmelpfennig et al., 2011); Spallation of K: 148.1 ± 7.8 atoms $^{36}$Cl (g K)$^{-1}$ a$^{-1}$ (Schimmelpfennig et al., 2014); Production rate of epithermal neutrons from fast neutrons in the atmosphere at the land/atmosphere interface: 626 ± 46 neutrons (g air)$^{-1}$ a$^{-1}$ (Phillips et al., 2001). (4) Spallation of Ca: 48.8 ± 3.4 atoms $^{36}$Cl (g Ca)$^{-1}$ a$^{-1}$ (Stone et al., 1996); Spallation of K: 148.1 ± 7.8 atoms $^{36}$Cl (g K)$^{-1}$ a$^{-1}$ (Schimmelpfennig et al., 2014); Production rate of epithermal neutrons from fast neutrons in the atmosphere at the land/atmosphere interface: 626 ± 46 neutrons (g air)$^{-1}$ a$^{-1}$ (Phillips et al., 2001).

| Sample         | Exposure Age (1) | Exposure Age (2) | Exposure Age (3) | Exposure Age (4) |
|----------------|------------------|------------------|------------------|------------------|
| Hualca 1       | 13.4 ± 2.3       | 11.7 ± 2.1 / 12.5 ± 2.2 | 13.9 ± 2.5       | 13.7 ± 2.4       |
| Hualca 2       | 17.8 ± 1.6       | 16.9 ± 1.7 / 17.3 ± 1.7 | 19.4 ± 1.9       | 18.8 ± 1.8       |
| Hualca 3       | 12.0 ± 1.5       | 11.0 ± 1.5 / 11.5 ± 1.5 | 12.7 ± 1.6       | 12.5 ± 1.6       |
| Hualca 4       | 16.6 ± 1.6       | 15.5 ± 1.8 / 16.0 ± 1.8 | 17.6 ± 1.9       | 17.4 ± 1.8       |
| Pujro Huayjo 1 | 9.3 ± 1.3        | 9.0 ± 1.3 / 9.2 ± 1.3 | 10.2 ± 1.5       | 9.9 ± 1.4        |
| Pujro Huayjo 2 | 12.0             | 11.2 ± 2.3 / 11.5 ± 2.4 | 13.3 ± 2.7       | 12.7 ± 2.6       |
| Patapampa 3    | 10.5 ± 1.3       | 9.5 ± 1.3 / 10.0 ± 1.3 | 11.1 ± 1.5       | 10.9 ± 1.45      |
| Patapampa 4    | 12.0 ± 0.9       | 11.8 ± 1.05 / 11.9 ± 1.0 | 13.3 ± 1.1      | 12.9 ± 1.1       |
| Mucurca 1      | 8.0 ± 0.8        | 7.8 ± 0.8 / 7.9 ± 0.8 | 8.8 ± 0.9        | 8.5 ± 0.9        |
4. Results

The cosmogenic $^{36}\text{Cl}$ exposure ages are shown in Table 2. In Huayuray valley (northern slope; Fig. 3A), the sample (Hualca 4) collected in a moraine that corresponds to the maximum glacier extent yields an age of 16.0 ± 1.8 ka. The other two samples (Hualca 2 and 3) come from intermediate moraines and yield an age of 17.3 ± 1.7 and 11.5 ± 1.5 ka, respectively. In Pujro Huayjo valley (Fig. 3B), the age of the sample (Pujro

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**Figure 3 A.** Moraine record and location of the samples collected on Huayuray valley (northern flank of Hualca-Hualca volcano).

**Figure 3 B.** Moraine record and location of the samples collected on Pujro Huayjo valley (southwestern flank of Hualca-Hualca volcano).
Huayjo 1) from the outer and more voluminous moraine is $9.2 \pm 1.3$ ka, while the sample (Pujro Huayjo 2) collected on a polished and striated bedrock located at the inner base of the moraine reveals an age ~2 ka older ($11.5 \pm 2.4$ ka).

In Mollebaya valley (Fig. 3C), the sample collected from the crest of the outer moraine (Hualca 1) yields an exposure age of $12.5 \pm 2.2$ ka. Another sample (Mucurca 1) was collected from the outermost moraine, related to the maximum glacier extent in Mucurca valley (Fig. 3D) and yields an exposure age of $7.9 \pm 0.8$ ka. By contrast,
two samples were taken in Patapampa (Fig. 3E). One sample (Patapampa 4) comes from glacially polished and striated bedrock located on the top of the southern edge of the Altiplano at 4886 m a.s.l., which shows an exposure age of 11.9 ± 1.0 ka. The other sample (Patapampa 3) was collected from a boulder in the outermost moraine located on the southern edge of Patapampa at 4671 m a.s.l., yielding an exposure age of 10.0 ± 1.3 ka.

Figure 3 E. Moraine record and location of the samples collected on Patapampa Altiplano.

5. Discussion

The $^{36}$Cl exposure ages from the outer moraines on the HualcaHualca volcano, based on the production rate of Marrero et al. (2016), indicates that the Last Local Glacial Maximum took place at ~17-16 ka during the Heinrich 1 event and Tauca paleolake cycle. Later, the retreat process began, but a phase of readvance or stillstand formed moraines at ~12 ka. According to the $^{36}$Cl exposure ages from glacially polished and striated bedrock, deglaciation would be the main process since ~11.5 ka and the ice cap of the Patapampa Altiplano had already disappeared in phase with the termination of the Younger Dryas, the Coipasa paleolake cycle and the onset of the Holocene. But the scarce number of exposure ages and the presence of outliers in the moraines represent limitations that only allows their use for a preliminary glacial reconstruction. Thus, this glacial chronology suggests that the LLGM in the study area does not coincide with the global Last Glacial Maximum, and a LLGM prior to the global Last Glacial Maximum, as proposed by Smith et al. (2005, 2008), is not confirmed. Moreover, the synchronicity between the evolution of glaciers in the study area and the Tauca and Coipasa paleolake cycles corroborates the assumption postulated by Hastenrath (1971), Klein et al. (1999), Amman et al. (2001) and Sagredo and Lowell (2012) of a sensitive reaction of glaciers located in the Central Andean arid regions to precipitation changes.
However, the $^{36}\text{Cl}$ cosmogenic exposures ages related to the LLGM on HualcaHualca are significantly younger than the chronology established on the nearby Coropuna volcano by Bromley et al. (2009) and Úbeda et al. (2012) as well as on Uturuncu volcano (Blard et al., 2014), where the LLGM is in phase or much older than the global Last Glacial Maximum. Conversely, the LLGM on HualcaHualca is consistent with the chronological data from Sajama and Tunupa volcanoes (Smith et al., 2009; Blard et al., 2013), indicating a connection with the Tauca paleolake cycle and the Heinrich 1 cold event. An explanation of this discrepancy is associated with the scaling scheme used in each volcano. For example, Bromley et al. (2009) show on Coropuna an LLGM chronology similar to HualcaHualca when the scaling schemes devised by Dunai. (2000), Lifton et al. (2005) and Desilets et al. (2006) are introduced. On the other hand, the exposure ages from HualcaHualca are ~2 ka older if the production rates for $^{36}\text{Cl}$ proposed by Stone et al. (1996) and Schimmelpfennig et al. (2011) are applied. These changes confirm the uncertainties linked to imprecise production rates in the high tropics as suggested by Balco et al. (2008), which do not allow for a consistent glacial history. Moreover, new production rates for $^{36}\text{Cl}$ have been proposed, such as Marrero et al. (2016) and Schimmelpfennig et al. (2011), and, therefore, published data associated with this cosmogenic isotope should be recalculated.

A complementary interpretation involves a glacier advance that could be less extensive during the global Last Glacial Maximum than the Late Glacial advances in the Central Volcanic Zone, as proposed by Smith et al. (2009). According to several proxy data from the Central Andes, the global Last Glacial Maximum was cold (a temperature decrease of 8-12°) and relatively humid (Minchin or Sajsi paleolake cycles) (Thompson et al., 1995, 1998; Baker et al., 2001; Fornace et al., 2014) whereas the Heinrich 1 and Younger Dryas events were characterised by a substantial increase in precipitation as the Tauca and Coipasa paleolake cycles (Blard et al., 2011; Placzek et al., 2013) reveal. This scenario would indicate that the glaciers of this region are sensitive to temperature and precipitation fluctuations, and would also explain the chronological differences across the Central Andes.

Due to the discrepancies between the LLGM chronologies, the onset of deglaciation since LLGM presents a heterogeneous age in the Andean Central Volcanic Zone. The cosmogenic exposure ages indicate that glacier retreat began before at Coropuna (Bromley et al., 2009; Úbeda et al., 2012) and Uturuncu volcanoes (Blard et al., 2014) than in HualcaHualca, Sajama and Tunupa volcanoes. However, based on the consistency of ages obtained from these three volcanoes, a preliminary interpretation of the beginning of deglaciation in the Andean Central Volcanic Zone would be related to the end of the Heinrich 1 event and the Tauca paleolake cycle.

Glacier retreat was interrupted by a readvance or stillstand, reported in all the volcanoes except Sajama, although chronologies do not coincide. The exposure ages from HualcaHualca and Coropuna are similar (12-13 ka), but they are older on Tunupa and Uturuncu volcanoes (14-16 ka). These data do not permit to infer clearly if the glaciers in the Andean Central Volcanic Zone reacted in synchronicity with the Younger Dryas and/or the Antarctic Cold Reversal cold events as proposed by Jomelli et al. (2014). After this marked phase of readvance or stillstand, the volcanoes of the Central Volcanic Zone experienced a marked retreat during the Holocene. Although on HualcaHualca volcano
there are no cosmogenic exposure dates of minor readvances or stillstands phases, the dates reported on Coropuna (~9 and 6 ka; Úbeda et al., 2012) and Sajama (~7 and 3.3 ka; Smith et al., 2009) indicate a non-homogeneous warm or/and dry climate during the Holocene.

6. Conclusions

The LLGM took place at ~17-16 ka on HualcaHualca volcano during the Heinrich 1 event and Tauca paleolake cycle. Deglaciation began after these cold and wet episodes, although it was interrupted by at least one readvance or stillstand at ~12 ka. Deglaciation was the main process since ~11.5 ka coinciding with the termination of the Younger Dryas, the Coipasa cycle and the onset of the Holocene. These glacial ages do not corroborate an LLGM prior to the global Last Glacial Maximum but they indicate a sensitive reaction of the glacier system to precipitation changes.

Based on the relative consistency of cosmogenic exposure ages from HualcaHualca, Sajama and Tunupa volcanoes, a preliminary interpretation of the beginning of deglaciation since the LLGM in the Andean Central Volcanic Zone would be related to the end of the Heinrich 1 event and Tauca paleolake cycle. At least one significant phase of readvance or stillstand has been reported in most of the volcanoes studied, although the chronologies do not coincide. Thus, establishing a clear connection with the Younger Dryas and/or the Antarctic Cold Reversal cold events is a difficult task. After this positive pulse, the ice masses of the Central Volcanic Zone experienced a marked retreat during the Holocene, only interrupted by three clear but minor readvances that suggest a heterogeneous warm or/and dry climate in this period.

Despite in situ cosmogenic exposure provides new relevant chronological information, it is not possible to provide a consistent regional glacier evolution or a paleoclimatic reconstruction since the LLGM, due to limitations associated with in situ cosmogenic production rates and the different scaling schemes used. To reduce the uncertainty and compare the available cosmogenic ages, it would be necessary to determine a precise in situ cosmogenic production rate of each isotope in the Central Andes as well as a standard scaling scheme.

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