Evidence of early Holocene glacial advances in southern South America from cosmogenic surface-exposure dating

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1GSA Data Repository item 2005##, Appendix DR1, Equilibrium line altitude reconstructions, cosmogenic surface-exposure methods, and data reduction, Table DR1 (boulder compositions), Table DR2 (10Be data), Table DR3 (36Cl data), and Figure DR1 (valley topography and hypsometry) is available online at www.geosociety.org/pubs/ft2005.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

ABSTRACT

Cosmogenic nuclide surface-exposure dating reveals that glaciers in southern South America (46°S) advanced at ca. 8.5 and 6.2 ka, likely as a result of a northward migration of the Southern Westerlies that caused an increase in precipitation and/or a decrease in temperature at this latitude. The older advance precedes the currently accepted initiation of Holocene glacial activity in southern South America by ~3000 yr. Both of these advances are temporally synchronous with Holocene climate oscillations that occurred in Greenland and the rest of the world. If there are causal links between these events, then rapid climate changes appear to be either externally forced (e.g., solar variability) or are rapidly propagated around the globe (e.g., atmospheric processes).
INTRODUCTION

Denton and Karlén (1973) identified three main episodes of Holocene alpine glacial expansion occurring at ca. 5300, 2800, and 200–300 calibrated yr B.P. (early, middle, and late Neoglacial or Little Ice Age, respectively; these and all radiocarbon ages have been calibrated with Calib 4.4 available at http://radiocarbon.pa.qub.ac.uk/calib/). Despite the prediction of glacial advances at ca. 8 ka based on the periodicity of the Neoglacial advances, confirming reports are generally not well accepted. In a review of Holocene glaciations (Davis and Osborn, 1988), six of seven papers discuss, but ultimately reject, evidence for glacial advances during the early Holocene. This skepticism stems from poor chronologic constraints for these deposits and the fact that warm conditions during the early Holocene are consistently interpreted from pollen and other paleoclimate records. One notable exception are the Cockburn moraines in northern Canada, which mark a readvance before the final collapse of the Laurentide Ice Sheet. However, hesitation to accept early Holocene glacial activity is at odds with the increasing evidence for global, quasi-periodic climate change that has occurred throughout the entire Holocene (e.g., Alley et al., 1997; Mayewski et al., 2004).

Radiocarbon dating of a few small moraines adjacent to the southern Andes (Röthlisberger, 1986; Wenzens, 1999) provides tantalizing evidence in support of these early Holocene advances; however, these results have been characterized as either local, anomalous events (Bennett et al., 2000) or requiring further confirmation (Heusser, 2003).

No clear consensus on the cause of Holocene climate variability has emerged. Some of the proposed mechanisms include (1) cyclic variations in solar output (Denton and Karlén, 1973; Bond et al., 2001); (2) modulation of thermohaline circulation (THC) (Teller and Leverington, 2004); (3) the stochastic resonance model, which combines elements of the first two mechanisms (Alley et al., 2001); and (4) changing concentrations of greenhouse gases such as water and methane in the atmosphere (Cane and Clement, 1999; Brook et al., 1999). However, rigorous testing of these hypotheses is hampered by a poor understanding of the synchronicity of climate-change events around the globe, as well as past changes in many of the weather systems of the world, especially the Southern Westerlies (e.g., McCulloch et al., 2000).

The Westerlies dominate the climate of southern South America, delivering on average 4000 mm of precipitation per year to the west side of the Andes south of 40°S; the core of the westerlies is focused at ~50°S, which receives almost 8000 mm/yr. Due to steep north-south and east-west precipitation gradients, changes in the position and/or intensity of the Westerlies could cause large changes in precipitation and temperature for a given location. Because glaciers are sensitive to these parameters, determination of the timing and magnitude of glacial advances is a powerful method of reconstructing past configurations of this important climate system and provides first-order data that can be used to test proposed mechanisms of climate change. We present equilibrium-line altitude reconstructions and $^{10}$Be and $^{36}$Cl cosmogenic surface-exposure ages from two moraines in southern Chile that indicate substantial glacier advances at ca. 8.5 and 6.2 ka.

GEOLOGIC SETTING

These moraines are located at Fachinal, Chile (Fig. 1A; 46.57°S 72.22°W). In the east, the moraines are two separate ridges, whereas the western part is a single complex
of hummocky deposits (Fig. 1B). They are sharp crested, vegetated by sparse grass and shrubs, and dotted with numerous large boulders. The moraines are deposited on a delta surface that is ~100 m above the modern level of Lago General Carrera (this lake spans the Chilean-Argentine border and is called Lago Buenos Aires in Argentina). The delta formed when drainage to the Pacific Ocean via the Rio Baker was blocked by glaciers advancing eastward out of the Northern Patagonian Icecap (Fig 1A). The moraines contain deformed lake sediment, and outwash channels flow across the delta surface, but do not incise the delta front. From this geomorphic evidence we infer that the moraines at Fachinal are coeval with advances of outlet glaciers of the Northern Patagonian Icecap. This regional synchronicity indicates that these glacier advances were responses to regional climate, rather than isolated glacial surges.

METHODS

The magnitude of climate changes responsible for these advances is estimated from the difference between the modern equilibrium-line altitude (ELA) and the paleo-ELA at the time of moraine deposition. The drainage currently contains several small (2–3 km) isolated cirque glaciers, from which we estimate a modern ELA of ~1400 m (Appendix DR1). The paleo-ELA is estimated by using an accumulation-area ratio of 0.65 ± 0.05, a value adopted by many other researchers (Brugger and Goldstein, 1999, and references therein). Paleo–glacier extents were digitally traced in a GIS over a digital elevation model (DEM) and a satellite image. The DEM was then clipped with the glacier outline and analyzed to determine equilibrium-line altitudes for a number of accumulation-area ratios.

Organic material suitable for radiocarbon dating was not found in the moraines; consequently the timing of deposition is determined by measuring the concentrations of in-situ cosmogenic $^{10}$Be and $^{36}$Cl in 16 erratic boulders. Samples were collected from 0.3–1.5 m diameter boulders at or near the moraine crests. The preferred boulder is one that has a wide, flat top that can be easily sampled, does not appear to have moved or to be weathering quickly (remnant glacial polish or sculpting is ideal), and has 20%–30% quartz if sampling for $^{10}$Be. Lithologies sampled were generally quartz-bearing rhyolite, but also included granites and metamorphic rocks; $^{36}$Cl was measured in two basalt samples. Some rocks showed signs of weathering, but many preserved glacial sculpting or polish. Chemical isolation of $^{10}$Be from pure quartz was performed at the University of Wisconsin–Madison following the methods of Bierman et al. (2003). Chlorine was separated from whole-rock samples following methods outlined in Stone et al. (1996) at the Cosmogenic Isotope Laboratory at the University of Washington. Accelerator mass spectrometry (AMS) analyses for both $^{10}$Be and $^{36}$Cl were performed at PRIME Lab, Purdue University.

Cosmogenic surface-exposure ages were calculated using production rates and scaling factors of Stone (2000). Corrections are applied to account for paleomagnetic field intensity, sample thicknesses, topographic shielding, and a slow erosion rate of 2 ± 2 mm/k.y. None of these corrections have significant impacts on the resulting ages. Uncertainties are reported at the 95% confidence level to represent the full uncertainty of the age determinations, and include all analytic errors (weighing of sample, weighing and concentration of spike, and AMS error), as well as erosion rate and attenuation-length uncertainty. Production-rate uncertainties are not explicitly treated; however, a ± 10%
systematic uncertainty would not fundamentally affect the conclusions of this paper. The full details of the methods and data reduction are presented in Appendix DR1.

RESULTS

At the time of moraine deposition, the glacial system was a 25-km-long, 15-km-wide ice field with a paleo-ELA of 1120 ± 65 m, estimated from an accumulation-area ratio of 0.65 ± 0.05 (Brugger and Goldstein, 1999). This ~300 m difference from the current ELA corresponds to conditions 2.4 °C cooler (if no change in precipitation has occurred), or 1000 mm/yr wetter (if no change in temperature has occurred) than present, on the basis of relationships between precipitation and temperature conditions in Patagonia and ELA (Hulton et al., 1994).

Five of six boulders from the inner moraine yield a weighted mean age of 6.2 ± 0.8 ka, and seven of 10 boulders from the outer moraine yield a weighted mean age of 8.5 ± 0.7 ka (Fig. 2; uncertainties at the 95% confidence level). Four outliers between 10.3 and 15.3 ka are identified on the basis of chi-squared statistics and bimodal probability-distribution curves; these outliers are excluded from the weighted means. We infer that they contain inherited cosmogenic $^{10}$Be from prior exposure. Complete results are available (Appendix DR1).

DISCUSSION

The high percentage of boulders with inheritance is a departure from prior interpretive guidelines. Putkonen and Swanson (2003) reviewed data from 638 moraine boulders presented in 22 papers. Following the original interpretations of the authors, they found that only 2% of the boulders were thought to have inherited isotopes and that subsequent exhumation was a much more prevalent problem. Nevertheless, we feel justified in excluding the four outliers (Fig. 2). First, the samples that remain in the population have a well-defined mean and show strong central tendency. Second, these glacial advances were much shorter than full glacial conditions (the focus of most of the chronology efforts reviewed in Putkonen and Swanson). Less material would have been eroded from the landscape, causing the moraine to contain a greater percentage of previously exposed material.

The most widely accepted ages of Neoglacial activity in southern South America are 5200–4500, 2800–1900 cal. yr B.P. as well as the Little Ice Age 300–200 yr BP (Mercer, 1982). However, very few moraines are bracketed by both minimum and maximum ages, thus the first period of Neoglacial activity is poorly constrained, but could have occurred between ca. 5400 and 4900 calibrated yr B.P. (Porter, 2000). The 8.5 ka surface-exposure age of the outer Fachinal moraine indicates that the most extensive Holocene glacial advance at this location was ~3000 yr earlier than the previously recognized onset of Neoglacial activity. This finding is supported by two other reports of glacier advances in the early Holocene: a maximum $^{14}$C age of ca. 9.4 cal. ka for a moraine at Venitsquero “José” (Röthlisberger, 1986; Fig. 1A and 3), as well as bracketing $^{14}$C ages of ca. 10.9 and 8.2 cal. ka, and 10.9 and 9.5 cal. ka for two moraines in the Rio Guanaco drainage (Wenzens, 1999; Fig. 1A and 3). Outside southern South America, there are few well-documented moraines correlative to the outer Fachinal moraine other than the Cockburn moraines in North America (Fig. 3). This lack of correlatable moraines is probably related to the limited preservation of early Holocene glacial
deposits, which may have been eroded or overrun by subsequent advances. The surface-
exposure age of the inner Fachinal moraine is potentially older than, but indistinguishable
from, Neoglacial activity in South America and on three other continents at ca. 5400–
4900 cal. yr B.P. (Fig. 3), given the analytical uncertainties and potential systematic shifts
in the production rate of the cosmogenic nuclides.

The closest nonglacial, Holocene climate records come from pollen in lakes and
bogs on the Taitao Peninsula (Lumley and Switsur, 1993; Bennett et al., 2000; Fig. 1A).
They show little evidence for climate change from deglaciation to the middle Holocene.
Cores from Lago Condorito, in the Chilean Lake District (Moreno, 2004; Fig. 1A and 3),
indicate dry and relatively warm conditions between 10 and 8 ka; precipitation started to
increase at 8 ka and reached a maximum at ca. 6–5 ka. However, the pollen and beetle
record at Puerto Eden (Fig. 1A), indicates wetter conditions in the early Holocene
(Ashworth et al., 1991). Reconstructions of the water level in Lago Cardiel (Fig. 1A), an
internally drained lake basin that responds to changes in effective precipitation, agrees
with a wet early Holocene (Stine and Stine, 1990; Markgraf et al., 2003). Shoreline
features and a variety of proxies from sediment cores indicate that lake levels were
highest at ca. 11 cal. ka, were falling but still high through the early Holocene, and
reached approximately modern levels at ca. 5 ka (Fig. 3).

Markgraf et al. (2003) hypothesized that this antiphased relationship between
Lago Cardiel and the Chilean Lake District is caused by a northward migration of the
Southern Westerlies from 50°S to 40°S between ca. 11 and 6 ka. The two glacier
advances at Fachinal suggest there may have been two increases in precipitation at 46°S,
first at ca. 8.5 ka and again at ca. 6.2 ka. While there are too few climate records to
adequately constrain the position of the Southern Westerlies through the Holocene, there
is little doubt that their position is the first order control on climate in this area. The
Fachinal moraines also appear to be synchronous with prominent excursions in sodium
and potassium ion concentrations in the GISP2 ice core (interpreted to track changes in
the Icelandic Low and Siberian High, respectively; Mayewski et al., 2004), as well as a
myriad of other paleoclimate proxies from Greenland and the rest of the world (Alley et
al., 1997). This is intriguing because the position of the Southern Westerlies is controlled
by the equator-to-pole thermal gradient, the position and strength of the southeast Pacific
high-pressure system, and the El Niño/Southern Oscillation (ENSO), and is thereby
related to global atmospheric processes in general (Cerveny, 1998). This apparent
synchronization of changes in the Southern Westerlies, Icelandic Low, and Siberian High
is most easily explained by either external forcing mechanisms such as variable solar
output (Denton and Karlén, 1973; Bond et al., 2001) or changes in atmospheric methane
or water-vapor contents (Brook et al., 1999; Cane and Clement, 1999). Oceanic heat
pumps, such as modulation of the thermohaline circulation (Teller and Leverington,
2004; Alley et al., 2001), are powerful climate-forcing mechanisms, but require a strong
coupling between the ocean and the atmosphere to propagate the climate signals rapidly
across the globe.

CONCLUSIONS

These substantial changes in glacier size and mass balance were caused by
significant changes in regional climate and the Southern Westerlies. The 8.5 ka surface-
exposure age of the outer Fachinal moraine is a clear indication that Neoglacial activity
started ~3000 years earlier than previously recognized in southern South America. Collectively, this and other paleoclimate records indicate that millennial-scale climate variability in mid-latitude South America, so prominent during the Last Glacial Maximum (e.g., Denton et al., 1999; Kaplan et al., 2004), continued throughout the entire Holocene. We hypothesize that the Fachinal moraines are synchronous with changes in the global climate system, and that the outer moraine may be correlative to the “8.2 ka event”. However, rigorous documentation of coeval glacier advances across the region and improvements in cosmogenic production rate uncertainties are needed before this hypothesis can be evaluated.

ACKNOWLEDGMENTS

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FIGURE CAPTIONS

Figure 1. A: Southern South America. Heavy black line indicates ice extent at Last Glacial Maximum. Arrow indicates current focus of southern westerlies at 50°S. See text for description of locations. B: Aerial photograph of Fachinal moraines. Circles and crosses indicate sample positions on outer and inner moraines, respectively. Heavy arrows (labeled MC) indicate positions of meltwater breaches of moraine crests onto delta surface. Abbreviations: CLD—Chilean Lake District; NPI—Northern Patagonian Icecap; RB—Rio Baker; TP—Taitao Peninsula; RG—Rio Guanaco; VJ—Ventisquero “Jose”; PE— Puerto Eden; LC—Lago Cardiel; LGCBA—Lago General Carrera–Buenos Aires; m asl—meters above sea level.

Figure 2. Cosmogenic surface-exposure ages and relative probability distributions for inner and outer Fachinal moraines. Each diamond or square represents ¹⁰Be or ³⁶Cl data from one boulder and may represent average of replicate analyses. Open symbols depict outliers. Error bars represent analytical and erosion-rate uncertainties at 95% confidence level. Vertical gray boxes represent weighted means of 5 of 6 and 7 of 10 samples for inner and outer moraines, respectively.

Figure 3. Selected climate records from Northern and Southern Hemisphere. Vertical gray bars represent globally expansive periods of rapid climate change (Mayewski et al., 2004). From top to bottom: Gaussian-smoothed (200 yr) K and Na concentrations from GISP2 ice core (Mayewski et al., 2004), paleovegetation index (positive values are warmer and drier) from Chilean Lake District (Moreno, 2004), lake-level record from Lago Cardiel (Stine and Stine, 1990; Markgraf et al., 2003), global glacial advances (Mayewski et al., 2004, and references therein), early Holocene glacial advances in southern South America at Fachinal, Ventisquero “Jose” (Röthlisberger, 1986), and Rio Guanaco valley (Wenzens, 1999).
Inner moraine
6.2 ± 0.8 ka
• $^{10}$Be

Outer moraine
8.5 ± 0.7 ka
• $^{10}$Be
• $^{36}$Cl

Relative probability
Exposure age (ka)

Douglass et al. Figure 2
Equilibrium Line Altitude Reconstructions

The difference between the modern and paleo equilibrium line altitude (ELA) is used to estimate the severity of cooling that caused these glacial advances. The modern ELA is estimated from snow lines observed on the glaciers currently present in the basin (Figure DR1A). Estimates range from 1300 to 1700 m; accordingly we conservatively estimate the modern ELA to be ~1400 m. The paleo-ELA is estimated by using the accumulation-area ratio method. Paleo–glacier extents were digitally traced in a geographic information system over a digital elevation model (DEM) and a satellite image, which was then analyzed to generate a valley hypsometry curve (Figure DR1B). Based on an accumulation area ratio of 0.65 ± 0.05, a value adopted by many other researchers (Brugger and Goldstein, 1999, and references therein), we estimate a paleo-ELA of ~1100 m. A ~300 m drop in regional snowline at the time of moraine deposition represents a significant cooling event. For comparison the ELA depression at the time of the Last Glacial Maximum is estimated to be ~1000 m (Rabassa and Clapperton, 1990).

Cosmogenic Surface-exposure Dating Methods

Samples were collected by hand with hammer and chisel from large (0.2 to 1.5 m diameter) boulders at or near the moraine crest. The preferred boulder is one that has a wide flat surface that can be easily sampled, does not appear to have moved, or be weathering quickly (remnant glacial polish or sculpting is ideal), and has 20-30% quartz if sampling for $^{10}$Be. Lithologies sampled were generally quartz-bearing rhyolite but also
included granites and metamorphic rocks; $^{36}$Cl was measured in two basalt samples. Some rocks were weathered and showed signs of spallation, but many preserved glacial sculpting or polish. The position and elevation of the samples was determined with a GPS and digital barometric altimeter respectively. There was good agreement between the altimeter and GPS elevations and we estimate a precision of ± 5 m for elevation measurements. The horizon shielding by the valley walls is characterized by several clinometer measurements and applied as a correction to production rates (see below).

Chemical isolation of $^{10}$Be was performed at the University of Wisconsin-Madison following the methods of Bierman and others (2003), with only minor variations for some of the samples. For most samples, up to 100 g of quartz was separated from 1 to 2 kg of crushed rock by mechanical (magnetic and heavy liquid separation) and chemical (acid dissolution of non-quartz minerals) procedures. The quartz was cleaned in repeated etchings in a dilute solution of nitric and hydrofluoric acid to remove the outer rim of the quartz crystals and any meteoric $^{10}$Be. After cleaning and a test to verify quartz purity, between 30 and 55 grams of quartz was dissolved in a mixture of concentrated hydrofluoric and nitric acids, along with a 500 µg spike of beryllium standard (Spex CertiPrep – Claritas PPT). A procedural blank for each batch of 11 samples is used to determine the amount of contaminant, non-cosmogenic $^{10}$Be contained in the spike and other chemicals used in sample processing. After dissolution, beryllium was isolated from other trace impurities in the quartz through a series of selective chemical precipitations, as well as cation and anion exchange column separations. The Be is oxidized, mixed with niobium metal, and packed into a target for accelerator mass spectrometry (AMS) analysis.
Chlorine was separated from whole rock samples following methods outlined in Stone and others (1996) at the Cosmogenic Isotope lab at the University of Washington. Ground samples were leached in hot, dilute nitric acid to remove meteoric $^{36}$Cl. A sub sample was retained for major and trace element analysis (Table DR1). After weighing and addition of a chloride carrier-spike, the sample was dissolved in a warm mixture of concentrated hydrofluoric and nitric acid. AgCl was precipitated by addition of AgNO$_3$. Sulfur was removed by dissolution of the AgCl in dilute ammonia and precipitation as BaSO$_4$ ($^{36}$S is an isobaric interference on $^{36}$Cl during AMS measurements). Pure AgCl was reprecipitated by acidification of the solution and loaded into targets for AMS analysis.

AMS analyses were conducted at PRIME Lab, Purdue University. $^{10}$Be samples were measured against standards derived from NIST SRM 4325, and we accordingly increase ratios by 14% (keeping percent error constant) to correct for activity differences between this standard and those used at other AMS facilities. As cosmogenic production rates are determined empirically, knowledge of the exact activity of $^{10}$Be is not necessary as long as there is robust intercalibration between standards (the basis of the 14% correction). $^{10}$Be concentration in the quartz is determined by standard isotope dilution calculations from the amount of spike added to the sample and the blank corrected isotopic ratio (the blank ratio is subtracted from the sample ratio). $^{36}$Cl concentration is determined by standard isotope dilution.

*Cosmogenic Surface-exposure Age Calculations (Table DR2, DR3)*
Be production rates are calculated according to Stone (2000) as the proximity to the Antarctic convergence leads to low air pressures and higher production rates in the region. Multi-decadal climate records in the region are sparse, but we estimate that mean annual sea level temperature is 285K and the mean annual sea level pressure is 1009.3 mb. A correction for modulation of production rate by changes in paleomagnetic intensity follow Nishiizumi et al., (1989) using the Sint-800 record of Guyodo and Valet (1999). The production rate is then scaled for sample thickness according to equations 3.78 and 3.81 of Gosse and Phillips (2001), and topographic shielding according to equation 6 of Dunne et al., (1999). We do not apply a correction for snow cover as the winter snow pack is generally thin and short lived.

Cl production rates are calculated using sea-level, high latitude values of 48.8 and 4.8 atoms gCa\(^{-1}\) yr\(^{-1}\) from spallation and muon capture reactions on Ca respectively (Stone et al., 1996, Stone et al., 1998), and 161 and 10.2 atoms gK\(^{-1}\) yr\(^{-1}\) from spallation and muon capture reactions on K respectively (Evans et al., 1997). Thermal and epithermal neutron capture rates are treated according to the method of Phillips et al. (2001). Sea-level, high latitude production rates are scaled to sample locations after Stone (2000) using atmospheric conditions described above. Cl decay constant used is 2.303 x 10\(^{-6}\) yr\(^{-1}\).

The exposure ages presented assume an erosion rate of 2 ± 2 mm/kyr. Kaplan et al. (2005) estimated erosion rate to be 1.5 ± 1.2 mm/kyr 100 km to the east of this area. We use a value that is slightly higher because this area is more humid, and we increase the uncertainty because of the extrapolation. Including erosion in the data reduction does
not significantly impact these ages; $^{10}$Be boulder ages and uncertainties generally increase by about 1.5% and 3.5% respectively.

Uncertainties are calculated according to the General Rule presented in Taylor (1997; equation 3.47; partial differential of the age equation with respect to each of the variables) and includes analytic errors (weighing of sample, weighing and concentration of spike, and AMS error), as well as erosion rate and attenuation length uncertainty. All errors are reported at the 95% confidence level as to represent the full uncertainty in the age determinations. Uncertainties related to production rate scaling were not explicitly treated in the data reduction. These uncertainties are poorly constrained but thought to be better than 10% (Gosse and Phillips, 2001). As samples integrate production rate variability over their exposure history, young samples are more sensitive to high-frequency changes in the magnetic field of the Earth and Sun. Accordingly, we estimate that systematic production rate error for these samples may be closer to 10%; resulting in systematic age uncertainties of about 600 and 900 years for the inner and outer moraines respectively.

Replicate analyses were performed for five samples. Splits were isolated after crushing to ensure independent measurements, and no replicates were processed in the same batch. Two of these replicates are not reproducible at the 95% confidence level (FAC-02-16A/B and LBA-02-07A/B). We expect is fundamentally related to the very low amounts of cosmogenic nuclides in these samples caused by their youth and low elevations. $^{10}$Be/Be ratios were on average only one order of magnitude higher than the procedural blank, and samples contain only about 1-5 million $^{10}$Be atoms. Low ratios make AMS measurements very difficult, and relatively small sample sizes exacerbate the
problem. Replicate measurements are averaged to produce one age for each boulder (which are shown in Fig. 2). Note that using only the older or younger age for these two problematic replications does not change the overall interpretation of the moraine ages. Using the same outlier identification protocol defined below, using the younger ages leads to moraine ages of $6.2 \pm 0.8$ ka (with FAC-02-16A and FAC-02-12 excluded) and $8.3 \pm 0.7$ ka (FAC-02-05, FAC-02-13, and LBA-02-04 excluded) for the inner and outer moraine, respectively. Using only the older ages leads to ages of $6.5 \pm 0.7$ ka (FAC-02-12 excluded), and $8.8 \pm 0.7$ ka (FAC-02-13 and LBA-02-04 excluded) for the inner and outer moraine, respectively.

**Cosmogenic Surface-exposure Data Interpretation**

Four boulders were excluded from the weighted mean calculations because of potential inheritance problems. We make the fundamental assumption that the majority of the boulders reflect the timing of deposition (with random errors), but that some boulders will have non-random, geologic errors related to inheritance or exhumation problems, making the boulders anomalously older or younger than the others. Geologic outliers are identified on the basis of Chi-Square statistics and frequency distribution curves. The Chi-Square test is used to test the null hypothesis that the error distribution of the population is statistically different from random noise. Samples are rejected until population errors appear random. The bi-modal nature of the frequency distribution curves agrees with the Chi-Square analysis; the samples rejected during the Chi-Square analysis are the samples which form the secondary frequency maxima. Including all of
the samples in the weighted means calculations does not affect produce significantly different ages: 6.8 ± 1.8 and 9.3 ± 1.4 ka for the inner and outer moraine, respectively.

Such a high percentage of boulders with inheritance is somewhat unusual and is a departure from prior interpretive guidelines. Putkonen and Swanson (2003) reviewed data from 638 moraine boulders presented in 22 papers. Following the original interpretations of the authors, it was found that only 2% of the boulders were interpreted to have inherited isotopes, and that subsequent exhumation was a much more prevalent problem. Nevertheless, we feel justified in excluding these outliers. First, the samples that remain in the population have a well defined mean and show strong central tendency. Second, ice thinning during glacier retreat started at about 15 ka (Kaplan et al., 2004) exposing bedrock and boulders at the surface to cosmic rays. These glacial advances were much shorter and less severe than full glacial conditions (the focus of most of the chronology efforts reviewed in Putkonen and Swanson), so less material would have been eroded from the landscape causing boulders with inherited isotopes to comprise a greater percentage of the total volume of sediment in the moraine.
Data Repository References:


### Table DR1: Composition data for Fachinal moraine boulders.

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### Table DR2: ¹⁰Be data from moraine boulders on Fachinal moraines

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<th>Sample ID</th>
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<th>Boulder height(m)</th>
<th>Boulder diam (m)</th>
<th>Thickness (cm)</th>
<th>Lat. (°S)</th>
<th>Long. (°W)</th>
<th>Altitude m asl</th>
<th>Mass Quartz</th>
<th>¹⁰Be Conc.</th>
<th>Prod. Rate at g⁻¹</th>
<th>Exposure Age</th>
<th>2σ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>FAC-02-04</td>
<td>Rhyolite</td>
<td>1.0</td>
<td>1.5</td>
<td>2.5</td>
<td>46.569</td>
<td>72.212</td>
<td>335</td>
<td>50.546</td>
<td>4.33 ± 0.90</td>
<td>7.01 ± 3.0</td>
<td>6.3 ± 1.3</td>
<td>2</td>
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<tr>
<td>FAC-02-10</td>
<td>Granite</td>
<td>1.0</td>
<td>1.1</td>
<td>2</td>
<td>46.571</td>
<td>72.238</td>
<td>343</td>
<td>31.849</td>
<td>3.99 ± 4.34</td>
<td>7.10 ± 6.3</td>
<td>5.7 ± 6.3</td>
<td>2</td>
</tr>
<tr>
<td>FAC-02-12</td>
<td>Vein Qtz</td>
<td>0.3</td>
<td>0.4</td>
<td>5</td>
<td>46.570</td>
<td>72.237</td>
<td>343</td>
<td>48.808</td>
<td>6.89 ± 1.20</td>
<td>6.85 ± 10.3</td>
<td>10.3 ± 1.8</td>
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<tr>
<td>FAC-02-16A</td>
<td>Rhyolite</td>
<td>0.4</td>
<td>0.5</td>
<td>2.5</td>
<td>46.570</td>
<td>72.237</td>
<td>348</td>
<td>41.220</td>
<td>4.70 ± 1.08</td>
<td>7.09 ± 7.6</td>
<td>7.7 ± 1.6</td>
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<tr>
<td>FAC-02-16B</td>
<td>Rhyolite</td>
<td>0.4</td>
<td>0.5</td>
<td>2.5</td>
<td>46.570</td>
<td>72.237</td>
<td>348</td>
<td>47.934</td>
<td>2.77 ± 0.86</td>
<td>7.09 ± 3.9</td>
<td>3.9 ± 1.3</td>
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</tr>
</tbody>
</table>

### Table DR3: ³⁶Cl exposure ages of Fachinal moraine boulders

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Lithology</th>
<th>Boulder height(m)</th>
<th>Boulder diam (m)</th>
<th>Thickness (cm)</th>
<th>Lat. (°S)</th>
<th>Long. (°W)</th>
<th>Altitude m asl</th>
<th>Ca %</th>
<th>K %</th>
<th>³⁶Cl conc.</th>
<th>Exp Age</th>
<th>2σ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>FAC-02-04</td>
<td>Rhyolite</td>
<td>1.0</td>
<td>1.8</td>
<td>2.5</td>
<td>46.568</td>
<td>72.211</td>
<td>323</td>
<td>20.907</td>
<td>5.27 ± 1.66</td>
<td>6.91 ± 7.8</td>
<td>2.5 ± 1.4</td>
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</tr>
<tr>
<td>LBA-02-04</td>
<td>Rhyolite</td>
<td>1.1</td>
<td>2.0</td>
<td>4.5</td>
<td>46.567</td>
<td>72.214</td>
<td>337</td>
<td>49.338</td>
<td>8.13 ± 16.2</td>
<td>6.88 ± 12.9</td>
<td>2.5 ± 1.2</td>
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</tr>
</tbody>
</table>

Table notes: Age equation used is \( t = (-1/(λ + E/L))*[\ln(1-N(λ + E/L)/P)]; \) where \( N \) is \(^{10}\)Be concentration, \( P \) is production rate, \( E \) is 2±2 mm/kyr expressed as g cm\(^{-2}\) y\(^{-1}\), \( λ \) is the decay constant (4.56±0.15 E⁻⁰⁷; Holden,1990), and \( L \) is the attenuation length (145±7 g cm\(^{-2}\); Brook et al., 1996). Topographic shielding for these samples is less than 1%. Procedural blanks contains 1.0 to 2.7 x 10\(^5\) atoms \(^{10}\)Be (\(^{10}\)Be/Be of 3-8 x10\(^{-15}\)); typically ~10% of total \(^{10}\)Be content.
Figure DR1: A: Landsat 7 ETM (Bands 7, 2, 3) and outline of glacier valley (in yellow). Elevation contours (100 m contour interval) from Shuttle Radar Topography Mission data (SRTM) are shown in red. B: Valley hypsometry curve with envelope for an accumulation area ratio of 0.65 ± 0.05. The upper part of the hypsometry curve is dashed because of “holes” in the DEM over steep snow covered slopes.