1 **Timing and nature of alluvial fan development along the Chajnantor Plateau, northern Chile**

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Abstract

Alluvial systems in the Atacama Desert provide a unique opportunity to elucidate the sedimentary response to climate variability, particularly changes in precipitation, in hyperarid environments. Alluvial fans along the eastern margin of the Salar de Atacama, adjacent to the Chajnantor Plateau in the Atacama Desert of northern Chile, provide an archive of climate-modulated sediment transfer and erosion at an extreme of Earth's climate. Three regional alluvial fan surfaces (Qf1 [oldest] to Qf3 [youngest]) were mapped along the western flank of the Chajnantor Plateau. The alluvial fans were examined with geomorphic and terrestrial cosmogenic ^{36}Cl surface exposure dating methods to define the timing of alluvial fan formation and to determine the role of climatic processes on fan development in a hyperarid environment. Alluvial fans in the study area are comprised of hyperconcentrated flow and boulder-rich debris flow deposits that reflect deposition transitioning between cohesive and noncohesive regimes. Alluvial fan surfaces yield exposure ages that range from 49.6 ± 4.4 to 194 ± 12 ka, while debris flow boulders yield exposure ages ranging from 12.4 ± 2.1 to 229 ± 53 ka. Cosmogenic ^{36}Cl exposure ages indicate that abandonment of alluvial fan surfaces Qf1, Qf2, and Qf3 date to 175 ± 22.6 ka (MIS 6), 134.5 ± 9.18 ka (MIS 6), and 20.07 ± 6.26 ka (MIS 2), respectively. A ^{36}Cl concentration-depth profile through alluvial fan Qf1 suggests a simple depositional history with minimal nuclide inheritance implying relatively rapid aggradation (6 m in ca. 25 ky) followed by surface abandonment ca. 180-200 ka. Our data support a strong climatic control on alluvial fan evolution in the region, and we propose that the alluvial fans along the margins of the Salar de Atacama form according to the humid model of fan formation.

Keywords: cosmogenic nuclides; alluvial fans; debris flows; arid Andes

1. Introduction

Alluvial fans are landforms associated with deposition along mountain fronts at transitions to lower gradients where valleys go from confined to unconfined. They serve a buffering role in mountain sediment systems, storing sediment derived from mountain catchments (Harvey, 1996, 1997; Harvey et al., 1999b, 2005). Environmental changes within the source area modulate sediment delivery to the fan and prompt the fan surface to respond by varying the mode and/or location of erosional or depositional regime. Thus, alluvial fans may preserve a record of environmental change within the source region (Harvey et al., 1999b, 2005). Alluvial fans form under an array of climatic and geomorphic settings and are controlled by autogenic depositional processes that vary spatially and temporally (Nicholas and Quine, 2007; Ventra et al., 2013; Ventra and Nichols, 2014), as well as external factors such as climate and tectonics (Bull and Schick, 1979; Harvey, 1990, 1997; Bull, 1991; Harvey and Wells, 1994; Ritter et al., 1995; Harvey et al., 1999a, 1999b; McDonald et al., 2003; Viseras et al., 2003; Pope and Wilkinson, 2005; Haug et al., 2010; Enzel et al., 2012; Hedrick et al., 2013; Owen et al., 2014). The processes that exert primary control on the timing and rate of formation of alluvial surfaces are often debated, and they vary from region to region. Tectonics, climate, paraglacial, and periglacial processes have been suggested as the primary drivers of alluvial surface development (e.g., Church and Ryder, 1972; Ritter et al., 1995; Hall et al., 2008). The coupling of several variables in time and space such as sediment supply, source area lithology, tectonics, climate, base-level change, etc., adds spatial and temporal complexity to the deposits, and often inhibits deciphering the contribution of a specific variable on the development of an alluvial system (e.g., Hartley et al., 2005; Pope and Wilkinson, 2005). In arid and hyperarid environments climatic change, specifically changes in precipitation, has been implicated as the primary driver on the evolution of alluvial fans, particularly regarding the transition from phases of aggradation to periods of stability and erosion (e.g., Wells et al., 1990; Harvey et al., 1999a, 1999b; Bull, 2000; Gunster and Skowronek, 2001; Enzel et al., 2012; McPhillips et al., 2013).

Inadequate chronological control on alluvial fan successions has restricted the extent to which these sedimentary systems can be utilized to understand tectonic and climatic controls on fan evolution and has hindered interpreting the environmental record preserved within alluvial fans (Owen et al., 2014). In arid and hyperarid environments, the paucity of organic matter preserved within alluvial fans restricts the application of radiocarbon dating. However, the development of terrestrial cosmogenic nuclide (TCN) surface exposure dating provides an additional method to determine the timing of alluvial fan formation and erosion (e.g., Dühnforth et al., 2007; Frankel et al., 2007a; Hall et al., 2008; Blisniuk et al., 2012; Hedrick et al., 2013; Owen et al., 2014) and expands the age range of landforms that can be dated well beyond the limit of radiocarbon dating. Terrestrial cosmogenic nuclide exposure ages on alluvial fan surfaces define periods of stability and represent the timing of surface abandonment and incision (Pope and Wilkinson, 2005; Owen et al., 2014).

The high preservation potential and sensitivity to climate shifts of arid and hyperarid alluvial systems provides an excellent opportunity to test the influence of climate on the timing of alluvial surface development and evaluate process-response models for fan aggradation and incision (e.g., Harvey, 1990; Bull, 1991; Harvey and Wells, 1994; Harvey et al., 1999b; McDonald et al., 2003; Enzel et al., 2012). The Atacama Desert of northern Chile, in particular, provides a unique opportunity to gain insight into the sedimentary response to precipitation at an extreme of Earth's climate. This study documents an extensive alluvial apron that has formed along the eastern margin of the Salar de Atacama, adjacent to the Chajnantor Plateau in the central Andes of northern Chile, to determine the nature and timing of alluvial fan formation in a region that lacks detailed chronologies. We report ^{36}Cl TCN exposure ages for several abandoned alluvial fan surfaces and discuss these ages in the context of local paleoclimate proxies and local records of glaciation to assess the role of climate on alluvial fan evolution in the region.

2. Regional setting

2.1. Geologic setting

The Atacama Desert of northern Chile is situated in the active Andean forearc between the Pacific Ocean and the central Andes. The Andean Cordillera comprises a number of distinct volcanic provinces along the length of the orogenic belt. Within the central Andes, the Central Volcanic Province (CVZ) extends from 14° to 28° S and hosts extensive Tertiary ignimbrite provinces such as the Altiplano-Puna Volcanic Complex (de Silva, 1989). The Chajnantor Plateau (23° S, 67° W) is a late Cenozoic, arid, volcanic plateau set within the framework of the subtropical Andes Mountains (Figs. 1, 2). The plateau reaches elevations of 5000 m above sea level (asl) and is capped by a number of more recent volcanic peaks that range in elevation from 5600 to 5900 m asl. A major succession of ignimbrites associated with the 4.5-4.1 Ma eruptions of the La Pacana caldera form the core of the plateau (Gardeweg and Ramírez, 1987; Lindsay et al., 2000). This older sequence of ignimbrites is overlain by the Purico Complex, which comprises an apron of three ignimbrite flow units. These ignimbrite sheets dip away from the vent at ~4°, generating the broad, dome-like structure of the plateau (Schmitt et al., 2001). The Lower Purico Ignimbrite, designated the Cajón Ignimbrite by Guest (1969) and later by Ramírez and Gardeweg (1982) and Gardeweg and Ramírez (1987), is the most extensive of the flow units and caps the majority of the western Chajnantor Plateau (Schmitt et al., 2001). The K-Ar eruptive ages of the Purico Ignimbrite units overlap and range between 1.38 ± 0.07 and 0.87 ± 0.52 Ma (Francis et al., 1984; de Silva, 1989b). A number of post-ignimbrite stratovolcanoes and domes overlie the Purico Complex; however, radiometric ages are generally lacking. The moderately pristine morphology and high degree of preservation of the post-ignimbrite stratovolcanoes and domes implies relatively young ages for them, <0.5 Ma as suggested by Francis et al. (1984). Schmitt et al. (2001) suggested post-glacial formation for two of the volcanic peaks, Cerro Aspero and Cerro Chascon.

Neighboring the western margin of the Chajnantor Plateau, the Salar de Atacama Basin is positioned between the Cordillera Domeyko and the western flank of the Andean Cordillera. The structural basin extends from 22.5° S to 24.5° S and is located in a region that has been subjected to subsidence since the Paleozoic (Ramírez and Gardeweg, 1982; Hartley et al., 1992a). The basin has been hydrologically closed since the Oligocene (Jordan et al., 2007) and is occupied by the modern Salar de

Atacama, a 3000 km² salt flat at 2300 m elevation that hosts an ~8-km-thick succession of nonmarine strata, providing a nearly complete record of Cenozoic sedimentation (Fig. 1; e.g., Bobst et al., 2001; Lowenstein et al., 2003; Arriagada et al., 2006; Jordan et al., 2007). Debate over the tectonic evolution of the Salar de Atacama Basin is considerable, and several conflicting models are described in the literature. Arriagada et al. (2002) and Mpodozis et al. (2005) suggested that the Salar de Atacama Basin developed during the Late Cretaceous following the inversion of the Tarapacá back arc basin. Uplift of the Cordillera Domeyko occurred as a consequence of the tectonic closure and inversion of the Tarapacá Basin (Jolley et al., 1990; Pananont et al., 2004). Other authors have argued that the basin evolved under extensional conditions for the entirety of the Mesozoic and Cenozoic (e.g., Flint et al., 1993) or initiated during an extensional regime that later transitioned to compression (e.g., Muñoz et al., 1997).

Along the western flank of the Chajnantor Plateau, a broad bajada extends from the base of the plateau (~2500 m asl) to the margin of the Salar de Atacama salt flat (~2300 m asl; Figs. 1, 2). The coalescing gravel fans and debris flows of this alluvial apron emanate from large bedrock canyons (*quebradas*) and smaller, fracture-controlled bedrock channels that have incised into the volcanic stratigraphy of the plateau. Hoke et al. (2004) proposed a groundwater-sapping mechanism for the large quebradas and argued that the smaller channels formed prior to the onset of aridity in the region ca. 5.8 Ma. Irwin et al. (2014) proposed flood erosion of strong-overlying-weak stratigraphy and flood erosion of vertically jointed rock as two other mechanisms of quebrada formation. The origin and mechanisms of formation of the smaller channels that are observed in the study area are enigmatic; however, they must postdate the 0.87 - 1.38 Ma Cajón Ignimbrite. Recent work has shown that wind abrasion plays a significant role in modifying channel morphology and extending bedrock canyons (Perkins et al., 2015).

The gravel fans and terraces of the Chajnantor bajada form a gently undulating surface with surface slopes that range from 0.6° to 3°, similar to those described by Hartley et al. (2005) on fans along the western Central Depression. Narrow, shallow channels (<1 m) are observed on some of the surfaces; however, none of these channels link back to the catchment. The gravels that constitute the alluvial fans and terraces are predominantly andesitic to dacitic and are sourced from the andesitic peaks and diamictic

volcaniclastic deposits (e.g., lahars) on the higher reaches of the plateau (Fig. 2). Clasts range in size from 1.5 to 19 cm, with average clast size ca. 4.5 cm. The alluvial surfaces are presently inactive and have been abandoned and incised. They are elevated 2-15 m above the modern channels and display both proximal and distal incision, implying that they are fully trenched. Numerous debris flow deposits are entrained within these incised channels, containing multiple lobes of poorly sorted boulders that suggest the amalgamation of discrete events into an expansive deposit (Fig. 2). The lobes consist of tens of meters long trains of ignimbrite and andesite boulders that range from 0.3 to 4 m. The matrix has been eroded from the majority of the deposits; however, where it is preserved the matrix consists of coarse sand with minor (<5%) proportions of silt. On the ignimbrite plateau, intermittent broad, andesitic gravel deposits remain perched, and stratigraphic sections of these gravels suggest multiple periods of aggradation with well-defined bed boundaries. These deposits may indicate volcaniclastic, glacial, or glaciofluvial (e.g., lahars, diamicts, etc.) generation and/or mobilization of sediment that has subsequently been reworked and scoured from regions of the ignimbrite surface (Fig. 2). Age constraints on the alluvial deposits along the western flank of the Chajnantor Plateau are sparse, but these deposits have been mapped as Quaternary alluvium or Quaternary gravels (Ramírez and Gardeweg, 1982; Marinović and Lahsen, 1984). The ^{10}Be concentrations converted to surface exposure ages for channel sediments and gravels that border the Salar de Atacama reported by Placzek et al. (2010, 2014) yield ages that span the mid-late Pleistocene.

2.2. Climate setting

The extreme paucity of precipitation in the Atacama Desert is the result of several factors. These include: (i) the Atacama lies in the Andean rain shadow which inhibits moisture from the Atlantic (Houston et al., 2006); (ii) its proximity to the descending limb of the Hadley circulation, resulting in the subtropical high pressure belt (Houston and Hartley, 2003); (iii) the Humboldt Current and upwelling cold water to the west limits the evaporation of ocean water (Caviedes, 1973; Houston et al., 2006; Amundson et al., 2012); and (iv) a regional circulation of subsiding air along the Altiplano (Houston et

al., 2006). The seasonal precipitation observed along the western Andean flank is a consequence of the South American Summer Monsoon (SASM) that controls the position of the Bolivian High and the incursion of extratropical precipitation events derived from the Atlantic (Lenters and Cook, 1997). These occur in the form of austral spring and summer storm events (Rech et al., 2003; Mather and Hartley, 2005; Houston, 2006; Placzek et al., 2009). Winter precipitation may also occur, especially during El Niño years (Vuille, 1999; Garreaud and Aceituno, 2001). Modern precipitation in the Altiplano region displays a steep precipitation gradient (Fig. 1). On the upper reaches of the Chajnantor Plateau (4500 m asl) precipitation averages $\sim 200 \text{ mm y}^{-1}$, which diminishes westward to $<20 \text{ mm y}^{-1}$ along the western slope (2500 to 3000 m asl) and $<10 \text{ mm yr}^{-1}$ in the Salar de Atacama Basin (2000 m asl; Latorre et al., 2003; Rech et al., 2003, Placzek et al., 2014). Three major physiognomic vegetation belts are present in the vicinity of the study area: the Prepuna (3000-3300 masl, $<30 \text{ mm y}^{-1}$) with sparse ($<5\%$) shrubs and succulents, the Andean talar zone or Puna belt (3200-4000 masl, $50\text{-}100 \text{ mm y}^{-1}$) with moderate (20-30%) shrubs and perennial grasses, and the high Andean steppe (4000-4500 masl, $100\text{-}200 \text{ mm y}^{-1}$) dominated by perennial bunch grasses (Latorre et al., 2002).

Considerable debate regarding the timing of the onset of hyperaridity in the Atacama Desert exists, and a number of postulations exist in the literature. Hoke et al. (2004), Clark (2005), Dunai et al. (2005), Nishiizumi et al. (2005), Rech et al. (2006), Nalpas et al. (2008), Placzek et al. (2014), and others argued for a Miocene onset of hyperarid conditions. Others proposed a Pliocene-Pleistocene onset of hyperaridity in the region (Hartley and Chong, 2002; Amundson et al., 2012; Jungers et al., 2013). Many recent studies have challenged the longstanding view that the central Atacama is a static, relict landscape with minimal to negligible Quaternary modification. Placzek et al. (2010, 2014), Amundson et al. (2012), and Jungers et al. (2013) offered evidence for a dynamic landscape in the central Atacama that has been subject to significant modification on Pleistocene and Holocene timescales. Indeed the hyperaridity of the central Atacama Desert, particularly in the region encompassing the Salar de Atacama and the adjacent Chajnantor Plateau along the Andean flank, has been punctuated by periods of increased precipitation throughout the Quaternary. Regional paleoclimate proxy records that show precipitation variability

include salt cores from the Salar de Atacama and the Salar de Uyuni in adjacent Bolivia (Baker et al., 2001; Bobst et al., 2001), sediment cores from high altitude lakes (Geyh et al., 1999; Fornari et al., 2001; Grosjean et al., 2001), shoreline deposits (Placzek et al., 2006; 2009; 2013), and rodent middens (Latorre et al., 2003, 2006; Maldonado et al., 2005; Díaz et al., 2012) and paleowetland deposits (Betancourt et al., 2000; Rech et al., 2002; Quade et al., 2008).

The majority of regional paleoclimate proxy records adjacent to the Salar de Atacama are restricted to the last ca. 45 ka, with a paucity of records that extend beyond ca. 100 ka. A 100 m salt core from the Salar de Atacama (Bobst et al., 2001) and a 121-m sediment core from the Salar de Uyuni (Fornari et al., 2001) are the longest paleoclimate records in the vicinity of the study area. The Salar de Atacama salt core (23° S), dated by U-series, records wet and dry phases of the salar over the past 106 ka (Bobst et al., 2001). Prominent wet phases appear near 70, 45, and 20 ka, with enhancement ca. 17-14 ka, which are substantiated by synchronous records interpreted from the Salar de Uyuni in adjacent Bolivia (Baker et al., 2001a; Fornari et al., 2001). The Salar de Uyuni sediment core (20° S), dated with radiocarbon and $^{40}\text{Ar}/^{39}\text{Ar}$, contains lacustrine layers and salt crusts that are interpreted as wet and dry phases over the past 191 ka (Fornari et al., 2001). Placzek et al. (2006a; 2009; 2013) proposed correlations between the lacustrine layers of the Salar de Uyuni sediment core with dated shoreline sediments from four large lake basins on the Altiplano (18-22° S): Titicaca, Poopó, Coipasa, and Uyuni. Several lake cycles are identified on the Bolivian Altiplano: Ouki (120-98 ka), Salinas (95-80 ka), Inca Huasi (47-46 ka), Sajsi (24-20.5 ka), Tauca (18.1-14.1 ka), and Coipasa (12.8-11 ka; Placzek et al., 2006a; 2013). The Tauca (ca. 140 m) and Ouki (ca. 80 m) lake cycles resulted in the largest and deepest lakes in the basin over the past 120 ka (Placzek et al., 2009).

Sediment records from Laguna Miscanti (Grosjean et al., 2001), rodent middens (Latorre et al., 2003; 2006; Maldonado et al., 2005; Díaz et al., 2012) and river-margin wetland deposits (Rech et al., 2002; 2003; Quade et al., 2008) support these wet phases at times when their ages overlap and suggest that the Salar de Atacama has experienced intermittent wet phases throughout the Holocene. These paleoclimate proxies indicate periods of increased precipitation during Marine Oxygen Isotope stage 2

(MIS 2, ca. 26.5-22 ka), the Tauca highstand (ca. 17.5-15 ka), and late glacial times (ca. 13-10 ka). Grosjean et al. (2001) and Latorre et al. (2006) inferred that precipitation increased up to 600 mm y⁻¹ during late glacial wet phases based on pollen assemblages.

Regional glacial chronologies allude to similar phases of increased moisture availability throughout the late Pleistocene. Glacial chronologies from the Chajnantor Plateau, northern Chile (Ward et al., 2015); Uturuncu Volcano, Bolivia (Blard et al., 2014); Encierro Valley, northern Chile (Zech et al., 2006); and Cordon de Doña Rosa, northern Chile (Zech et al., 2007) generally agree with the regional nonglacial paleoclimate proxies. Glacial chronologies between 10° S and 30° S are relatively synchronous and suggest moraine formation between 40 and 25 ka, deglaciation or stillstand between 25 and 20 ka, and moraine formation between 20 and 15 ka (Ward et al., 2015).

Evidence from the Chajnantor Plateau (23° S) suggests multiple stages of glacial occupation (Ward et al., 2015). Mapping and surface exposure dating of glacial landforms and scoured bedrock on the southern Altiplano and Chajnantor Plateau by Ward et al. (2015) suggests at least three periods of moraine stabilization. Deposits associated with Stage III likely formed prior to 100 ka. Stage II deposits likely formed around 50-60 ka, with continued glaciation or reglaciation ca. 30-25 ka establishing the stage I deposits. On the basis of this chronology, Ward et al. (2015) argued for glaciation of the Chajnantor Plateau ca. 50 ka, with reglaciation or continued glaciation until ca. 30 ka, followed by deglaciation by ca. 25 ka.

Surface exposure dating of glacial landforms on Uturuncu volcano (22° S) by Blard et al. (2014) suggested three main periods of glacial stillstand. Blard et al. (2014) argued that the Uturuncu glacier may have attained its maximum extent as early as 65 ka, remaining proximal to this position with two stillstands identified near 40 ka and between ca. 35 and 17 ka. The Uturuncu glacier underwent a limited retreat around 17 ka, followed by a final glacial stillstand between 16 and 14 ka synchronous with the Tauca highstand (Placzek et al., 2006; Blard et al., 2014). Exposure ages from scoured bedrock in the upper reaches of the valley suggest deglaciation was accomplished ca. 14 ka (Blard et al., 2014).

Zech et al. (2006, 2007) proposed similar glacial chronologies from Encierro Valley (29° S) and Cordon de Doña Rosa (31° S). Using the rescaled ¹⁰Be exposure ages of Ward et al. (2015), the chronology of Zech et al. (2006) from Encierro Valley suggests that extensive glaciation occurred ca. 35 ka, with the most prominent advance between 20.9 and 18.3 ka. Dating of younger recessional moraines within the upper reaches of the valley suggest minor glacial advances around 17.4 ka (Zech et al., 2006; Ward et al., 2015). The glacial chronology from Cordon de Doña Rosa suggests an extensive glaciation between 49 and 41 ka, with several advances occurring between 21.5 and 17.2 ka (Zech et al., 2007; Ward et al., 2015).

3. Methods

3.1. Geomorphic mapping

Detailed mapping of alluvial fans, gravel surfaces, and debris flow deposits utilized high resolution (40 cm) satellite imagery from Google Earth coupled with a 30-m digital elevation model from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) project alongside corroboration in the field (Fig. 3). Alluvial surfaces were delineated based on surface morphology, stratigraphic relationships, and weathering characteristics (Fig. 4; e.g., McFadden et al., 1988; Owen et al., 2014). Geologic maps of Ramírez and Gardeweg (1982) and Marinović and Lahsen (1984) were used to identify the regional geology and previously recognized surfaces and deposits in the study area. Surfaces for the alluvial fans and gravel deposits are numbered independently: Qf1 (oldest) and Qf2 (youngest) and Qg1 (oldest) and Qg2 (youngest), respectively. A lowercase ‘a’ (upper) or ‘b’ (lower) was added to the surface name when subdivisions within major surfaces were apparent, e.g. Qg2a and Qg2b.

3.2. Cosmogenic exposure dating

3.2.1. Sampling

In order to determine the timing of alluvial surface abandonment and debris flow deposition, rock samples were collected for ^{36}Cl surface exposure analysis. Alluvial fan and gravel deposit surface samples were collected from broad, flat, elevated fan surfaces that displayed minimal evidence for erosion, anthropogenic effects, or deflation or inflation of the surface and were located away from eroding slopes and ephemeral surface channels. Sampling sites were carefully selected on the highest elevation of the fan surface to avoid complications of topographic shielding and to minimize the potential for sediment shielding during the exposure history of the surface. Each alluvial surface and gravel surface sample was composed of 25 andesite or dacite cobbles of ~3 cm diameter, which were crushed and amalgamated. This process aids in minimizing spurious nuclide concentrations owing to nuclide inheritance or weathering. Samples for debris flows were collected by removing 0.5-1 kg of the upper surface from individual boulders. Sampled boulders were embedded in the surrounding sediment and were elevated above the surface, likely retaining their original position since deposition and minimizing the likelihood of shielding by sediment cover. The location of the boulders in boulder trains provides further evidence that the boulders have remained *in situ*. Multiple boulders were sampled on each surface to assess the possibility of weathering or nuclide inheritance.

Cosmogenic inheritance is one of the major complications concerning TCN exposure dating of alluvial surfaces and is a chief concern when dealing with younger alluvial surfaces, as well as units that are composed of reworked sediment from older deposits (e.g., Owen et al., 2011; Blisniuk et al., 2012). Previous studies have assessed the nuclide inheritance of sediment (Anderson et al., 1996; Hancock et al., 1999; Jungers et al., 2013), surficial erosion (Hidy et al., 2010), and complexity of depositional history (Matmon et al., 2006; Jungers et al., 2013) through the use of TCN depth profiles. Below a certain depth, typically ~2 m, the cosmogenic nuclide concentration of sediment is recognized as the inherited component derived from previous periods of exposure. In order to constrain potential nuclide inheritance in surficial samples, a depth profile was sampled from a gravel pit near the town of Toconao that exposed

an ~8-m section of alluvial fan unit Qf1 (Figs. 3, 4). Six samples were collected from several depths, determined by clast abundance of the required size fraction, to a depth of 6 m. At each depth sampled, 25 andesitic pebbles were collected using the same criteria as surface samples. The surface of the gravel pit immediately surrounding the exposure was buried by tailings piles so determination of whether the present surface is the original (top) depositional surface was unclear. To determine whether the exposed fan surface represented the original surface, multiple exploratory pits were constructed along the same fan surface. A characteristic sand layer occurs ~0.5 m below the surface, containing abundant interspersed ignimbrite clasts. This unit was observed in all of the exploratory pits along the fan surface, suggesting the unit is laterally continuous. This same unit is exposed in the gravel pit at ~0.55 m, which supports the interpretation that the exposed surface adjacent to the gravel pit represents the original surface.

3.2.2. Cosmogenic laboratory methods and age determinations

Extraction, chemical preparation, and sample dissolution for ^{36}Cl followed the procedures of Stone et al. (1996) and were carried out at the University of Cincinnati. Procedures were slightly modified to accelerate the dissolution process (J. Radler, Purdue University, personal communication, 2014) and are outlined here. Each sample was crushed and sieved to the 125-250 μm fraction, of which 80 g was leached overnight in dilute HNO_3 and subsequently rinsed and dried to remove contaminants and eliminate the meteoric ^{36}Cl component. After drying, 30 g of the leached material was spiked with ~1.0 mg of enriched ^{35}Cl carrier and dissolved in concentrated TMG (trace metal grade) HF and TMG HNO_3 . Following complete dissolution, AgCl was precipitated through the addition of AgNO_3 . The AgCl precipitate was collected and redissolved through the addition of NH_4OH and eluted through anion exchange columns to remove interfering ions. The AgCl was precipitated through the addition of AgNO_3 and HNO_3 . The AgCl was dried, weighed, and loaded into Cu cathode holders packed with AgBr for AMS analysis at Purdue Rare Isotope Measurement (PRIME), where $^{35/37}\text{Cl}$ and $^{36/37}\text{Cl}$ ratios and Cl concentrations were determined. Major and trace element compositions were analyzed by Na_2O_2 fusion

and ICP-MS at Activation Laboratories Ltd. in Ontario, Canada, and at Bureau Veritas Commodities Canada Ltd., British Columbia, Canada.

All ^{36}Cl exposure ages were calculated using the community standard CRONUS-Earth online calculator (Marrero et al., 2015a; Marrero et al., 2015b; Phillips et al., 2015; <http://web1.ittc.ku.edu:8888/2.0/>; v.2.0). Input tables are provided as supplemental files. Ages were calculated using the time-dependent ‘Lal/Stone’ (Lal, 1991; Stone, 2000) scaling scheme that uses ^{36}Cl production rates of 52.5 ± 3.6 at $\text{g Ca}^{-1} \text{ y}^{-1}$, 151 ± 10 at $\text{g K}^{-1} \text{ y}^{-1}$, and 649 ± 201 neutrons $\text{g}^{-1} \text{ y}^{-1}$ (Marrero et al., 2015a). These factors are shown locally to yield ages that are consistent with other, independent cosmogenic measurements (Ward et al., 2015).

3.2.3. Cosmogenic depth profile interpretation

Modeling of cosmogenic ^{36}Cl depth profiles is complicated by the depth-dependent behavior of the various production mechanisms. Most importantly, production by thermal and epithermal neutrons peaks in the shallow subsurface (here, ~20 cm depth) and makes up ~18% of total production at that depth in the rock considered here, before declining with further depth (Gosse and Phillips, 2001). This low-energy production also depends sensitively on the chemistry of the samples and so is difficult to generalize. This makes analytical solutions to the case of production during gradual accumulation of a deposit much less tractable than in the case of simple spallogenic production, which can be well-approximated by a simple exponential with depth. It also complicates numerical integrations (e.g., Schimmelpfennig et al., 2009). Including the time-dependent nature of the production mechanisms owing to variations in the geomagnetic field, solar flux, and atmospheric pressure further complicates the situation and makes the kind of straightforward profile analysis common in ^{10}Be studies (e.g., Anderson et al., 1996; Balco et al., 2012) much more difficult with ^{36}Cl . Full numerical simulation of all of the above effects through time and with depth, across different scenarios of burial rate and inheritance, is a major undertaking and is beyond the scope of this study. Nonetheless, we would like to analyze our depth

profile so as to (i) assess the potential contributions of inheritance and progressive burial to the measured concentrations of ^{36}Cl and (ii) determine whether a simple history of deposition and surface abandonment explains the measured concentration profile or if a more complex approach is required. Therefore, we employed a simplified two-part modeling approach to analyze the depth profile, maintaining the important features of the complexity described above and remaining compatible with the surface exposure ages computed for the surficial deposits using the regionally appropriate scaling scheme.

First, the Age Calculation Engine (ACE) software of Zweck et al. (2013) was used to compute a time-averaged production rate-depth profile, including the depth-dependent low-energy production pathways, over the timescale relevant to the samples in the profile (ca. 240 ka). Components used within the calculation included time-varying production rate scaling scheme based on latitude-altitude-pressure factors of Stone (2000), the sea level curve of Fairbanks (1989) and Shackleton (2000), the paleolatitude record of Ohno and Hamano (1992) and Yang et al. (2000), and the geomagnetic intensity model of Guyodo and Valet (1999). These choices are generally compatible with the scaling scheme used for the surface exposure ages. The profile was computed by generating a synthetic sample set, having the same bulk chemistry, location, and range of depths as the real samples, but distributed at much greater depth resolution, with 1-cm spacing in the near subsurface to resolve the low-energy production peak, with increased spacing at depth where production rate profiles are more gradual. Second, the production rate curve was used in conjunction with a MATLAB script, which allowed simple manipulation of cosmogenic inheritance and accumulation history. Essentially, at each time step in the accumulation history, the script tracks the production at each depth in the deposit, where the depths evolve through time. Production at each time step is determined by simple interpolation of the production rate curve generated from our synthetic sample profile and so is relevant to the location and chemistry of the real profile, while allowing more flexibility in accumulation pattern.

3.3. Luminescence sampling and processing methods

Where suitable sand deposits were observed, optically stimulated luminescence (OSL) samples were collected to corroborate surface exposure ages. Two OSL samples were collected; one from a distinct sand layer at ~3.8 m in the depth profile at Toconao, and another was collected at ~2 m from an exposed sand layer within a perched gravel deposit (Qg2b) on the upper reaches of the plateau. Samples were collected by hammering 20-cm-long, 5-cm-diameter black PVC tubes into cleaned sediment faces and extracting the tubes from the exposure wall. Tube ends were capped, sealed with duct tape, and placed horizontally in bags for transport. The samples were processed under safelight conditions at the University of Cincinnati.

A 3-cm-thick layer of sediment was removed from each end of the tube to retrieve the center plug of sediment. Elemental concentrations for dose rate calculations were determined through instrumental neutron activation analysis on ~10-g aliquots of sediment at Activation Laboratories Ltd. in Ontario, Canada. Sediment from the center of the tubes was leached in HCl and H₂O₂ to remove carbonates and organic matter, respectively. The leached samples were rinsed in deionized H₂O, dried, and sieved to collect the 90-250 μ m fraction. This fraction was etched in HF for 70 minutes to remove feldspar contaminants in addition to the outer, α -irradiated layer from quartz grains. Any subsequent fluoride precipitates were removed using concentrated HCl. The quartz sample was rinsed in deionized H₂O and C₂H₃O₂, then dried and sieved to the 90-150 μ m fraction. Aliquots of each sample were loaded on 10-mm stainless steel disks for measurement. Luminescence signals were measured on a Riso TL/OSL reader (model DA-20) using an array of blue light emitting diodes (470 nm, 50 mW cm⁻²). Signals were filtered through a BG-39 filter, detected through a Hoya U-340 filter, and measured using a 52-mm-diameter EMI 9235 QA photomultiplier tube. All aliquots were checked for feldspar contamination through infrared stimulations with infrared light-emitting diodes (870 nm, 150 mW cm⁻²). Equivalent dose (D_e) measurements were determined on multiple aliquots for each sample using the single aliquot regenerative (SAR) method protocol developed by Murray and Wintle (2000) and Jain et al. (2008). Growth curves were generated by fitting a saturated exponential function.

The D_e values for each aliquot were examined using the Riso Analyst 3.22b software. Aliquots that displayed poor recuperation (>10%) were not included in age calculations. The D_e of all eligible aliquots were used to determine a weighted mean D_e , which was then divided by the calculated dose rate to determine the weighted mean age for the sample. The cosmic dose rate was calculated using the methods of Prescott and Hutton (1994), and the methods of Adamiec and Aitken (1998) and Mejdahl (1979) were used to combine cosmic dose rate with β and γ doses from elemental concentrations to calculate the total dose rate.

4. Results

4.1. Mapping of alluvial features

Mapping of alluvial features along the western flank of the Chajnantor Plateau revealed several major surfaces (Fig. 3). The majority of the ignimbrite shield of the Chajnantor consists of a bare bedrock surface; however, sparse gravel deposits are preserved below the volcanic peaks. These gravel deposits are locally thick (~50 m), diamictic, and likely consist of debris flow and lahar deposits (Ramírez and Gardeweg, 1982), glacial and periglacial deposits, and volcanogenic sediment. The oldest of the gravel surfaces, Qg1, is highly dissected by fluvial bedrock channels (Fig. 3). Andesitic clasts range from 20 to 40 cm and are highly ventifacted and pitted, suggestive of an older surface. The younger gravel surface Qg2a is significantly more coherent than the surface of Qg1. Surface Qg2a displays evidence of shallow, surface channels; however, it is not dissected by bedrock channels as is the case with Qg1. Gardeweg and Ramírez (1982) mapped the southern deposit of Qg2a near Cerro Lascar as a lahar deposit, and that origin is adopted in this manuscript. The northern deposit of Qg2a likely has a similar origin; however, the surface appears to be more denuded. Clasts range from 0.2 to 2 m and are predominantly andesitic, and the surface appears to be undergoing local reworking with the eroded material generating surface Qg2b

(Fig. 3). The Qg2b unit displays a much smaller clast size (1-3 cm) than the source deposit, and abundant coarse sand lenses are observed.

Two major alluvial fan surfaces were identified through morphostratigraphy and weathering characteristics of surface clasts (Figs. 3, 4). The older of the two surfaces, Qf1, is an upper abandoned alluvial surface that rises 5-15 m above adjacent modern channels and locally extends onto the surface of the ignimbrite shield. The surface is limited in its extent, consisting of coherent deposits alongside smaller fragmented remnants dissected by fluvial channels. A younger alluvial fan surface, Qf2, represents a lower abandoned alluvial surface that is elevated 2-10 m above the modern channel. While Qf2 is locally dissected by fluvial networks that emanate from bedrock channels in the ignimbrite shield, the surface is generally coherent and displays minimal fragmentation (Fig. 3). Thin veneers of andesite gravels are observed on the ignimbrite slopes and are interpreted as remnants of surface Qf2. Where stratigraphic sections are visible, the deposits comprise subangular to subrounded, poorly sorted, clast-supported, pebble to cobble-grade gravels with a coarse sand matrix. The deposits are largely structureless with minor sand lenses and are considered to represent hyperconcentrated flow deposits, similar to those described in Wells and Harvey (1987) and Hartley et al. (2005). Along the southern margin of the Chajnantor Plateau below Cerro Lascar, the Qf2 surface transitions from hyperconcentrated flows to broad, debris flow deposits (Fig. 3).

Entrained in bedrock channels on the ignimbrite shield and within modern and ancient channels incised into the alluvial surfaces are numerous lobe-type debris flow deposits, Qf3 (Figs. 2, 3). These debris flow deposits consist of poorly sorted, randomly oriented, boulder-grade andesite and ignimbrite clasts that range from 0.3 to 4 m in diameter, and where locally preserved, a coarse sand-grade matrix reflecting deposition from cohesive flows (Wells and Harvey, 1987). Where the coarse sand-grade matrix has been eroded, the deposits are represented by boulder fields. The majority of the boulders are entrained within a coarse sand matrix or the sediment of the channel, although a minor fraction is observed resting on the surface (Fig. 2). Weathering characteristics of the boulders are insufficient to determine a relative chronology of the discrete lobe deposits (Fig. 4).

4.2. Surface exposure and luminescence ages

Eleven ^{36}Cl analyses on alluvial fan ($n = 10$) and gravel surface ($n = 1$) samples yielded ages that range from 49.6 ± 4.4 to 540 ± 130 ka (Tables 1, 2; Fig. 3.). Seven of these 11 ^{36}Cl samples yielded exposure ages between 122 and 194 ka. Eight ^{36}Cl samples from debris flow boulders yielded exposure ages that range from 12.4 ± 2.1 to 229 ± 53 ka. Six of these eight samples yielded ages between 12.4 and 28.3 ka (Tables 1, 2; Fig. 3). Errors include the analytical uncertainty and propagated production rate uncertainties. The two optically stimulated luminescence samples, Chaj-OSL1 and Toc-OSL1, yield ages of 40.51 ± 2.00 and 42.86 ± 2.03 ka, respectively (Tables 3, 4; Fig. 3).

4.3. TCN depth profile

Measured nuclide concentrations for all six samples in the depth profile exhibit the expected theoretical exponential decrease but do not appear to capture the ‘*humped*’ nature of the ^{36}Cl depth profile near the surface (Fig. 5). The ‘*humped*’ curve near the surface typically occurs within the top 20-30 cm of the profile, so the possibility exists that the sparse sampling within the exposure did not have resolution to capture the hump (Gosse and Phillips, 2001). Moreover, spallation dominates the ^{36}Cl production in these rocks, and as a result, the contribution of the thermal component is muted. The nuclide concentrations of the depth profile samples are in good agreement with the nuclide concentrations for the surface samples, AFDP-1-14 and AFDP-2-14, from surface Qf1 (Table 1; Fig. 5). The exposure age of 150 ± 13 ka for sample AFDP-1-14 is younger than the expected age of the depth profile, which may be attributed to a minor amount of post-depositional erosion. Modeled solutions to the depth profile were produced using the methodology described above in section 3.2.3. Values for the rate of deposition, timing of deposition, and inheritance were parameterized to explore various depositional histories. The modeled best-fit solution uses ca. 25 ka of deposition at a rate of $\sim 0.24 \text{ mm a}^{-1}$ and zero inheritance and assumes a bulk

density for the deposit of 2.0 g/cm^3 . The modeled profile passes through eight of the nine samples (including all samples at depth) and suggests surface abandonment at ca. 200 ka (Fig. 5). Using the ‘Lal/Stone’ scaling scheme (Lal, 1991; Stone, 2000), the model predicts a surface abandonment age of 185 ka, while using the ‘Lifton/Sato’ scaling (Lifton et al., 2014) the model predicts surface abandonment at ca. 214 ka. These abandonment ages are taken to be in good agreement with one another given scaling uncertainties over this time span. Other parameterizations of bulk density, deposition rate, and inheritance also yield model fits within this range.

4.4. Landform associations

Quaternary gravels: The dissected gravel surface, Qg1, on the upper reaches of the plateau is defined by a single exposure age of $540 \pm 130 \text{ ka}$. No surface exposure ages were collected from gravel surface Qg2a or Qg2b. On the basis of surface characteristics and morphology, the more coherent nature of the deposit suggests that it is younger than Qg1. A luminescence age from sample Chaj-OSL1 of $40.51 \pm 2.00 \text{ ka}$ at 2 m depth within gravel surface Qg2b corroborates this interpretation, and represents a maximum age for the Qg2b surface (Table 1; Fig. 3). However, the volcanic nature of the sediment suggests that this luminescence age is an underestimate of the actual age of deposition (Bonde et al., 2001; Tsukamoto et al., 2003).

Quaternary alluvial fans: Three exposure ages from surface Qf1 yield ages of 150 ± 13 , 181 ± 26 , and $194 \pm 12 \text{ ka}$ and a mean age for the surface of $175 \pm 22.6 \text{ ka}$ (Table 1; Fig. 3). The OSL sample collected at a depth of 3.8 m yields an age of $42.86 \pm 2.03 \text{ ka}$, which contradicts the surface exposure age. Previous luminescence studies have highlighted discrepancies in the OSL components of volcanic and sedimentary quartz (e.g., Tsukamoto et al., 2003). Bonde et al. (2001) reported average OSL age underestimates of ~90% using volcanic quartz from a paleosol, while Tsukamoto et al. (2007) reported D_e underestimates of ~80% using quartz phenocrysts from a tephra. From the prevalence of volcanic lithologies in our study area, it is reasonable to infer that the discrepancy with the surface exposure age is

caused by an underestimation of the luminescence age. As such, the cosmogenic ^{36}Cl exposure ages are taken as the preferred ages. The surface exposure age and depth profile concentrations suggest that surface Qf1 formed prior to MIS 6 and was abandoned by MIS 6d-6e.

Seven surface exposure ages from surface Qf2 range from 49.6 ± 4.4 to 144 ± 13 ka. Four samples from the Qf2 surface yield ages between 122 ± 12 and 144 ± 13 ka, and two samples yield exposure ages of 72.1 ± 8.9 and 82 ± 11 ka. Sample AF-1-14 with an exposure age of 49.6 ± 4.4 ka was collected from a channel depression in the Qf2 surface and has potentially been reworked or covered by sediment causing the younger exposure age. Samples AF-2-14 and AF-3-14 were collected at lower elevations along the fan surface; as such the alluvial surface may have been more prone to reworking and results in younger exposure ages (72.1 ± 8.9 and 82 ± 11 ka, respectively). Alternatively, the samples may reflect a period of reworking when the surface was reactivated post-abandonment. Four surface exposure ages between 122 ± 12 and 144 ± 13 ka yield a mean age of 134.5 ± 9.2 ka for the Qf2 surface, which suggests that the surface likely formed during MIS 6 and was abandoned by MIS 5e-6a. The younger exposure ages of 72.1 ± 8.9 and 82 ± 11 ka may reflect localized reoccupation of the surface ca. MIS 4-5a.

Quaternary debris flows: Eight samples from debris flow boulders on the Qf3 surface yield exposure ages that range from 12.4 ± 2.1 to 229 ± 53 ka. Six of these eight samples yield ages between 12.4 and 28.3 ka (Table 1; Fig. 3). Samples DF3-B1-13 and DF1-B3-14 yield surface exposure ages of 110 ± 26 and 229 ± 53 ka, respectively. When compared to the other six samples, these appear to be anomalously old. Sample DF3-B1-13 was collected from near the plateau cliff-face and may be representative of a reworked rock-fall boulder. Owing to the short travel distance, it is likely that insufficient erosion would have taken place to reset the inherited nuclide concentration, thus resulting in the old age. Sample DF1-B3-14 was collected from an andesite boulder, which is more resistant to erosion than ignimbrite. Transport through a cohesive debris flow may have been insufficient to produce enough erosion to reset the nuclide concentration. An alternative explanation is that the boulders are derived from much older events, more probable for the sample of andesite, as it is located ~5 km from the

toe of the plateau, adjacent to the salar. Six surface exposure ages between 12.4 and 28.3 ka yield a mean age of 20.07 ± 6.26 ka for the Qf3 surface. Three surface exposure ages between 12.4 and 20.2 ka and two surface exposure ages of 26.3 and 28.3 ka yield mean ages of 15.8 ± 3.99 and 27.3 ± 1.41 ka, respectively, for the individual debris lobe deposits (Fig. 3). A single exposure age of 18.4 ± 3.7 ka defines the third debris flow lobe sampled. Surface exposure ages on boulders of the debris flow deposits suggest that surface Qf3 formed and was subsequently abandoned during MIS 2.

Alluvial surfaces and debris flow deposits along the western flank of the Chajnantor Plateau appear to have been abandoned and incised during three main periods. The oldest alluvial fan surface (Qf1) likely formed prior to MIS 6 and was abandoned by MIS 6d-6e. Alluvial fan surface Qf2 likely formed during MIS 6 and was subsequently abandoned by MIS 5e-6a. However, the surface may have been locally active until ca. MIS 4-5a or reoccupied at that time. The youngest surface, consisting of debris flow deposits, likely formed and was abandoned during MIS 2.

5. Discussion

5.1. Geologic sources of uncertainty in surface exposure ages

Two primary sources of geologic uncertainty that affect cosmogenic exposure ages are (i) post-depositional erosion or burial of the sample by sediment or snow that reduces the apparent exposure age, and (ii) inheritance of nuclide inventory from previous periods of exposure that increases apparent exposure ages. These sources of geologic uncertainty are of particular concern when dealing with alluvial systems that rework sediment from upslope. Long residence times of sediment stored in the upper reaches of catchments yield high concentrations of nuclide inventories that are transferred into the younger alluvial system upon reworking. As a consequence, the high nuclide inventories result in an increased inherited nuclide signal within the younger alluvial deposit.

Scatter among some of the cosmogenic ages from the alluvial surfaces and debris flow boulders suggests that both of these sources may be affecting our results. In the context of their depositional setting, ages on debris flow boulders DF3-B1-13 (110 ± 26 ka) and DF1-B3-14 (229 ± 53 ka) likely reflect incomplete resetting of the nuclide inventory. Alluvial fan sample AF-1-14 yields an exposure age of 49.6 ± 4.4 ka, while the adjacent sample AF-2-14 yields an exposure age of 82 ± 11 ka. Both samples were collected from the Qf2 surface that yields a mean age of 134.5 ± 9.2 ka, which suggests that they have been subject to sediment cover, erosion, or reworking post-abandonment producing the younger ages.

5.2. Uncertainty in depth-profile modeling

Our depth-profile modeling approach captures the most important time-transient behavior — the evolution of production with depth — but makes a few simplifying assumptions. One, it assumes that the time-averaged production rate at this site over a timescale of ca. 240 ka is representative of that at each time-step within the evolution of our model profile. Variation of a few percent is likely over this time. Note that the production profile used is an integration of the time-varying production rates over that timescale, not a time-independent scaling. So long as this timescale is relevant to the period of time each sample in the profile spent in the near-surface, high-production rate portion of the profile, this assumption is unlikely to result in significant error. An identical chemistry is assumed for all points in our synthetic profile, based on an average of the real profile samples. However, sample chemistry varies very little across this profile and should not be expected to affect the age results by more than ~1% (Table 2). Geological uncertainty (e.g., transient cover, periods of profile erosion, and changes to water content) is expected to be the most significant source of error and is relatively difficult to constrain, as it is for the surficial samples. The tight fit of the modeled depth profile to the depth sample concentrations suggests that there is minimal inherited nuclide signal, at least within the Qf1 deposits (Fig. 5). This suggests that

the alluvial fan displays a relatively simple depositional history, as a steady state deposition rate can explain the majority of the measured concentrations.

5.3. Chronology of alluvial fan development

Alluvial fans and debris flows within the study area provide an archive to elucidate the nature of climate-modulated sediment transfer in the region. Depositional processes, inferred from the morphology and sedimentology of the alluvial fans, appear to be dominated by hyperconcentrated flows with minor contributions from debris flows. Cosmogenic exposure ages across the bajada, though scattered, reveal active landscape modification throughout the mid- to late Pleistocene and record a dynamic history of sediment storage and transport over the past ca. 540 ka.

The geomorphology of the study area and the morphology of the alluvial fans, coupled with our exposure age chronology, indicate two main phases of alluvial fan development. A major phase of alluvial fan aggradation occurred prior to or during MIS 6 and likely resulted in the main period of bajada development, evident from the extensive Qf1 and Qf2 fans. Aggradation on the Qf1 and Qf2 fan surfaces appears to have ceased by the transition to MIS 5e-6a and was followed by incision and subsequent abandonment of the fan surfaces (Fig. 6). A period of proximal fan head trenching and distal fan progradation, with localized reoccupation of the Qf2 fan surface, persisted throughout MIS 5 and ceased ca. MIS 4-5a. Fan head trenching and incision of the fan surfaces suggests that MIS 5 was a time of considerable landscape readjustment, likely reflecting changes in sediment availability within the source area catchments and a reduction in sediment supply to the alluvial fans (e.g., Harvey and Wells, 1994). Reduction in sediment supply within the source area catchments can result in the transition from a sediment-rich depositional regime to sediment-poor fluvial regime (Harvey and Wells, 1984; Harvey, 1984b, 1989). This switch may have contributed to the abandonment of the Qf1 and Qf2 fan surfaces and initiated fan head trenching and surface incision during MIS 5.

A period of quiescence in fan formation followed, which ended with the initiation of the second phase of fan aggradation during MIS 2. The second major phase of alluvial fan development resulted in the formation and abandonment of the Qf3 surface during MIS 2, characterized by localized, proximal aggradation of debris flow deposits within the fluvial channels (Fig. 6). Geomorphic activity along the alluvial apron has significantly slowed in the modern, reflected in the cessation of alluvial aggradation and the presence of minor trenching and fan incision. Backfilling of channels within the catchments implies transport mechanisms that are insufficient to deliver sediment to the fan surfaces, supporting the low level of geomorphic activity. A reduction in sediment supply in the upper catchments, coupled with reduced mean flow velocity conditions, may have contributed to the transition from hyperconcentrated flow to debris flow deposits ca. MIS 2 (Pierson and Costa, 1987). The transition to debris flow deposition may have occurred prior to MIS 2, as indicated by the presence of lobes of denuded boulders in adjacent deposits.

These phases of alluvial development do not appear to have been restricted to the eastern margin of the Salar de Atacama, as equivalent ^{10}Be concentrations expressed as exposure ages (120-220 ka, MIS 5e-7c) on channel gravels and sediments, P_1 through P_4 , are observed along the western margin adjacent to the Cordillera Domeyko (Fig. 1; Placzek et al., 2014). Placzek et al. (2014) reported two ^{10}Be concentrations that yielded exposure ages of 60 ± 10 and 80 ± 10 ka (MIS 4-5a) from sediment along the bajada below Chajnantor, P_5 and P_6 , similar to the two younger ages on the Qf2 surface, 72.1 ± 8.9 ka and 82 ± 11 ka (MIS 5a; Figs. 1, 6). These younger ages support the interpretation that the alluvial surfaces along the western flank of the Chajnantor Plateau may have been subject to reoccupation or reworking around the transition to MIS 4 (ca. 71 ka).

5.4. Climatic controls on alluvial fan formation

Over the past few decades, numerous authors have proposed several process-response models of alluvial aggradation to changing environmental conditions (e.g., Church and Ryder, 1972; Bull, 1977;

Ponti, 1985; Barnard et al., 2006). Current research proposes two prevailing process-response models concerning alluvial aggradation and fan formation associated with catchments in desert environments, referred to hereafter as the arid and humid models. The arid model argues that alluvial fan aggradation is associated with a relative increase in aridity (e.g., Bull, 1977, 1991, 2000; Wells et al., 1987; McDonald et al., 2003; Pope and Wilkinson, 2005), whereas the humid model argues that fan aggradation is associated with a relative increase in precipitation (e.g., Ponti, 1985; Dorn et al., 1987; Harvey et al., 1999a, 1999b; Spelz et al., 2008). The arid model of aggradation is widely accepted for alluvial systems of the southwestern United States, as developed by Bull (1991; 2000). As highlighted by Spelz et al. (2008) and Owen et al. (2014), the arid model of fan formation proposes that fan aggradation coincides with the transition from glacial to interglacial cycles when increased temperatures and reduced precipitation result in reduced vegetation and decreased soil moisture, which leads to an increased sediment-supply-triggering aggradation. Aggradation ceases when the system crosses a deposition-erosion threshold and sediment supply to the fan decreases — reflecting stabilization of the hillslope colluvium by vegetation and/or exhaustion of the colluvium in the source region (e.g., Bull, 2000). Following the cessation of aggradation, the geomorphic system transitions to a transport-limited regime with fluvial systems incising through the alluvial surface. Under the arid alluvial fan model, fan sediments should have interglacial ages of deposition, whereas abandoned surfaces can form anytime thereafter when the deposition-erosion threshold is crossed, prior to the onset of the next glacial period (Spelz et al., 2008; Owen et al., 2014).

Alternatively, the humid model proposed by Harvey et al. (1999a, 1999b) argues that aggradation of alluvial fan sequences (stacked and telescopic) coincides with wetter conditions related to glacial cycles. As highlighted in Spelz et al. (2008), the humid model implies that aggradation and incision occur during wetter climatic cycles, associated with high geomorphic activity. Geomorphic activity decreases during arid interglacial periods, as demonstrated in the American Southwest by Harvey et al. (1999a) where alluvial fan development was restricted to minor fan head trenching (Spelz et al., 2008). According to the humid model of alluvial fan development, geomorphic thresholds that modulate aggradational and

erosional regimes should be crossed at a high frequency (Spelz et al., 2008), and alluvial aggradation should cease with the onset of arid interglacial conditions. Spelz et al. (2008) emphasized that both process-response models are valid in different desert environments; proposing that the arid model is most appropriate in semiarid deserts that undergo major changes in vegetation corresponding to glacial cycles, whereas the humid model is most appropriate in hyperarid deserts, where climate only transitions between hyperarid and arid conditions over glacial and interglacial cycles. In such transport-limited environments, increased runoff is required to drive the system toward aggradation. Alternating cycles of erosion and deposition should correspond with glacial and interglacial cycles under both process-response models.

Well-established archives of climate variability in our study area are restricted to the past ca. 50 ka, and records extending beyond 100 ka are sparse. This paucity of archives that record variations in climate beyond 100 ka makes deciphering process-response models of fan aggradation problematic. Additional challenges arise when correlating fan formation with climatic shifts owing to the increased uncertainties associated with dating older alluvial landforms (e.g., Owen et al., 2014). Nevertheless, inferences can be made toward the nature of alluvial fan development in the study area.

Exposure ages for alluvial fan surfaces Qf1 and Qf2 suggest abandonment ca. MIS 6d-6e and MIS 5e-6a, respectively (Fig. 6). The presence of a salt crust (S5, 35.4-41.5 m) in the Salar de Uyuni sediment core ca. MIS 6, reflecting increased aridity on the Bolivian Altiplano (Fornari et al., 2001; Placzek et al., 2013) coincides with the timing of surface abandonment. We propose that the cessation of alluvial aggradation for fans Qf1 and Qf2 is associated with decreased geomorphic activity that corresponds with a shift toward more arid conditions. A lacustrine layer (L5, 41.5-46.7 m) dated to ca. 191 ka implies wetter conditions on the Altiplano prior to abandonment of the Qf1 and Qf2 fans (Fornari et al., 2001). Nuclide concentrations in our depth-profile model for fan Qf1 suggest deposition over ca. 25 (Fig. 5) and supports aggradation of fan Qf1 during periods of increased precipitation. Therefore, we propose that the Qf1 and Qf2 fans likely formed according to the humid model. Localized reoccupation of alluvial surface Qf2 during MIS 5 coincides with the Ouki and Salinas (80-120 ka) lake cycles on the Altiplano (Placzek et al., 2006a, 2013) and a lacustrine layer (L4, 22-35.4 m) in the Salar de Uyuni

sediment core (Fornari et al., 2001) reflecting increased precipitation on the Altiplano, supporting our humid model interpretation. Exposure ages for debris flow boulders of surface Qf3 suggest aggradation and abandonment ca. MIS 2, synchronous with the glaciation of the Chajnantor Plateau (Ward et al., 2015) and regional glaciations (Zech et al., 2006, 2007; Blard et al., 2013), the Tauca highstand and prior Sajsi lake cycle of the Altiplano (Placzek et al., 2006, 2013), and a highstand in the Salar de Atacama (Fig. 6; Bobst et al., 2001). The synchronicity of climate proxies that reflect increased precipitation with the aggradation of surface Qf3 suggests that the alluvial fan deposit formed according to the humid model of fan development. Veit et al. (2016) described aggradation/incision cycles in the Río Lluta Valley, linking river incision ca. 15.4 k cal a BP to increased precipitation and runoff. Alluvial fans in the Pampa del Tamarugal in addition to the Qf3 fans in our study area experienced aggradation coincident with the timing of incision in the Río Lluta catchment. As highlighted by Veit et al. (2016), these contradictory responses to increased precipitation arise from the nature of the geomorphological system. Fans in the endoheric basins of the Atacama Desert are generally inactive during periods of increased aridity, with phases of aggradation coinciding with increased precipitation. In perennial fluvial systems such as the Río Lluta, relative increases in precipitation result in increased and/or more consistent discharge, allowing for incision (Veit et al., 2016).

Formation of alluvial fans in our study area is largely controlled by climate-modulated sediment supply. We recognize that the uncertainty associated with dating older alluvial landforms and the limited availability of temporally relevant paleoclimate archives hinders our ability to adequately test whether the alluvial fans in the study region fully comply with the humid model of formation. However, this highlights the need to further explore what conditions are conducive to alluvial fan formation in hyperarid environments during the late Quaternary.

6. Conclusions

We document the existence of three alluvial fan surfaces, defined by geomorphic characteristics and morphostratigraphy, along the western flank of the Chajnantor Plateau. The alluvial fans comprise hyperconcentrated flow and boulder-rich debris flow deposits that reflect deposition transitioning between cohesive and noncohesive regimes. Cosmogenic ^{36}Cl exposure ages indicate that abandonment of alluvial fan surfaces Qf1, Qf2, and Qf3 date to 175 ± 22.6 ka (MIS 6), 134.5 ± 9.18 ka (MIS 6), and 20.07 ± 6.26 ka (MIS 2), respectively. A ^{36}Cl concentration-depth profile through alluvial fan Qf1 suggests a simple depositional history with steady state aggradation over ca. 25 ka followed by surface abandonment ca. 200 ka. Our data support a strong climatic control on alluvial fan formation in the region. The paucity of temporally relevant paleoclimate archives and regional fan chronologies coupled with the uncertainty associated with dating older alluvial landforms hinders our ability to conclusively determine climate conditions for alluvial fan formation. We propose that alluvial fans in this near-hyperarid environment evolve according to the humid model of fan formation. However, additional alluvial chronologies coupled with more extensive dating of the alluvial fans is required to adequately assess this suggestion. Nevertheless, the alluvial fans in our study area provide an archive of climate modulated sediment transfer and erosion throughout the Late Quaternary.

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References

- Adamiec G., Aitken M.J., 1998. Dose-rate conversion factors: new data. *Ancient TL* 16, 37–50.
- Amundson, R., Dietrich, W., Bellugi, D., Ewing, S., Nishiizumi, K., Chong, G., Owen, J., Finkel, R., Heimsath, A., Stewart, B., 2012. Geomorphologic evidence for the late Pliocene onset of hyperaridity in the Atacama Desert. *Geol. Soc. Am. Bull.* 124 (7-8), 1048–1070.
- Anderson, R. S., Repka, J. L., Dick, G. S., 1996. Explicit treatment of inheritance in dating depositional surfaces using in situ ^{10}Be and ^{26}Al . *Geology* 24 (1), 47-51.
- Arriagada, C., Cobbold, P. R., Mpodozis, C., Roperch, P., 2002. Cretaceous to Paleogene compressional tectonics during deposition of the Purilactis Group, Salar de Atacama, paper presented at Fifth International Symposium on Andean Geodynamics (ISAG), Inst. de Rech. pour le De´v., Toulouse, France.
- Arriagada, C., Cobbold, P.R., and Roperch, P., 2006. Salar de Atacama basin: A record of compressional tectonics in the central Andes since the mid-Cretaceous. *Tectonics* 25 (1), NIL_17-NIL35.
- Baker, P.A., Rigsby, C.A., Seltzer, G.O., Fritz, S.C., Lowenstein, T.K., Bacher, N.P., Veliz, C., 2001a. Tropical climate changes at millennial and orbital timescales on the Bolivian Altiplano. *Nature* 409, 698–701.
- Balco G., Soreghan G.S., Sweet D.E., Marra K.R., Bierman P.R., 2012. Cosmogenic-nuclide burial ages for Pleistocene sedimentary fill in Unaweep Canyon, Colorado, USA. *Quat. Geochronol.* 18, 149-157.
- Barnard, P.L., Owen, L.A., Finkel, R.C., 2008. Quaternary fans and terraces in the Khumbu Himalaya, south of Mt. Everest: their characteristics, age and formation. *J. Geol. Soc.* 163, 383-400.
- Betancourt, J.L., Latorre, C., Rech, J.A., Quade, J., Rylander, K.A., 2000. A 22,000-yr record of monsoonal precipitation from northern Chile's Atacama Desert. *Science* 289, 1546-1550.
- Blard, P.-H., Lave, J., Farley, K.A., Ramirez, V., Jimenez, N., Martin, L.C.P., Charreau, J., Tibari, B., Fornari, M., 2014. Progressive glacial retreat in the Southern Altiplano (Uturuncu volcano, 22°S) between 65 and 14ka constrained by cosmogenic ^3He dating. *Quat. Res.* 82, 209-221.
- Blisniuk, K., Oskin, M., Fletcher, K., Rockwell, T., Sharp, W., 2012. Assessing the reliability of U-Series and ^{10}Be dating techniques on alluvial fans in the Anza Borrego Desert, California. *Quat. Geochronol.* 13, 26-41.

- Bobst, A. L., Lowenstein, T. K., Jordan, T.E., Godfrey, L.V., Ku, T., Luo, S., 2001. A 106ka paleoclimate record from drill core of the Salar de Atacama, northern Chile. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 173 (1-2), 21-42.
- Bookhagen, B. and Strecker, M.R., 2008. Orographic barriers, high-resolution TRMM rainfall, and relief variations along the eastern Andes. *Geophys Res. Lett.* 35 (6), L06403.
- Bookhagen, B. and Strecker, M.R., 2012. Spatiotemporal trends in erosion rates across a pronounced rainfall gradient: Examples from the southern Central Andes. *Earth Planet. Sci. Lett.* 327-328, 97-110.
- Bonde, A., Murray, A., Friedrich, W.L., 2001. Santorini: Luminescence dating of a volcanic province using quartz?. *Quat. Sci. Rev.* 20 (5-9), 789-793.
- Bull, W.B., 1977. The alluvial-fan environment. *Prog. Phys. Geol.* 1, 222-270.
- Bull, W.B., 1991. *Geomorphic responses to Climate Change*. Oxford University Press, New York.
- Bull, W.B., 2000. Correlation of fluvial aggradation events to times of global climate change. In: Noller, J.S., Sowers, J.M., Lettis, W.R (Eds.), *Quaternary Geochronology: Methods and Applications*. American Geophysical Union, Washington, D.C., pp. 456-464.
- Bull, W. B., Schick, A. P., 1979. Impact of climatic changes in an arid watershed: Nahal Yael, southern Israel. *Quat. Res.* 11 (2), 153-171.
- Caviedes, C., 1973. A climatic profile of the north Chilean desert at latitude 20° south. In: Amiran, D.H.K., and Wilson, A.W. (Eds.), *Coastal Deserts, Their Natural and Human Environments*. The University of Arizona Press, Tucson, AZ. pp. 115-121.
- Church M., Ryder J.M., 1972. Paraglacial sedimentation: consideration of fluvial processes conditioned by glaciation. *Geol. Soc. Am. Bull.* 83 (10), 3059-3072.
- Clark, J.D.A., 2005. Antiquity of aridity in the Chilean Atacama Desert. *Geomorphology* 73, 101-144.
- De Silva, S.L., 1989. Geochronology and stratigraphy of the ignimbrites from 21°30'S to 23°30'S portion of the central Andes of northern Chile. *J. Volcanol. Geotherm. Res.* 37, 93-131.
- Díaz, F.P., Latorre, C., Maldonado, A., Quade, J., Betancourt, J.L., 2012. Rodent middens reveal episodic, long distance plant colonizations across the hyperarid Atacama Desert over the last 34,000 years. *J. Biogeogr.* 39 (3), 510-525.
- Dorn, R.I., DeNiro, M.J., Ajie, H.O., 1987. Isotopic evidence for climatic influence on alluvial-fan development in Death Valley, California. *Geology* 15, 108-110.
- Dühnforth, M., Densmore, A.I., Ivy-Ochs, S., Allen, P.A., Kubik, P.W., 2007. Timing and patterns of debris flow deposition on Shepherd and Symmes creek fans, Owens Valley, California, deduced from cosmogenic ¹⁰Be. *J. Geophys. Res.* 112 (F3), F03S15.
- Dunai, T.J., González López, G.A., Juez-Larré, J., 2005. Oligocene–Miocene age of aridity in the Atacama Desert revealed by dating of erosion-sensitive landforms. *Geology* 33 (4), 321–324.

- Durcan, J.A., King, G.E., Duller, G.A.T., 2015. DRAC: Dose rate and age calculator for trapped charge dating. *Quat. Geochronol.* 28, 54-61.
- Enzel, Y., Amit, R., Grodek, T., Ayalon, A., Lekach, J., Porat, N., Bierman, P., Blum, J.D., Erel, Y. 2012. Late Quaternary weathering, erosion, and deposition in Nahal Yael, Israel: An “impact of climatic change on an arid watershed”? *Geol. Soc. Am. Bull.* 124 (5–6), 705–722.
- Fairbanks, R.G., 1989. A 17,000 year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep ocean circulation. *Nature* 342, 637–642.
- Flint, S., Turner, P., Jolley, E., Hartley, A.J., 1993. Extensional tectonics in convergent margin basin: An example from the Salar de Atacama, Chilean Andes. *Geol. Soc. Am. Bull.* 105 (5), 603-617.
- Francis, P.W., McDonough, W.F., Hammill, M., O’Callaghan, L.J., Thorpe, R.S., 1984. The Cerro Purico volcanic complex, North Chile. In R.S. Harmon, B.A. Barriero (Eds.), *Andean Magmatism-Chemical and Isotopic constraints*, pp. 150–161. Shiva Publishing, Orpington, U.K.
- Frankel, K.L., Brantley, K.S., Dolan, J.F., Finkel, R.C., Klinger, R.E., Knott, J.R., Machette, M.N., Owen, L.A., Phillips, F.M., Slate, J.L., Wenicke, B.P., 2007a. Cosmogenic ^{10}Be and ^{36}Cl geochronology of offset alluvial fans along the northern Death Valley fault zone: Implications for transient strain in the eastern California shear zone. *J. Geophys. Res.* 112 (B6), B06407.
- Fornari, M., Risacher, F., Féraid, G., 2001. Dating of paleolakes in the central Altiplano of Bolivia. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 172, 269-282.
- Gardeweg, P.M., Ramirez C.F., 1987. The La Pacana Caldera and the Atana ignimbrite: A major ash-flow and resurgent caldera complex in the Andes of northern Chile. *Bull. Volcanol.* 49 (3), 547-566.
- Garreaud, R.D., Aceituno, P., 2001. Interannual rainfall variability over the South American Altiplano. *J. Climate* 14, 2779-2789
- Geyh, M.A., Grosjean, M., Nunez, L.A., Schotterer, U., 1999. Radiocarbon reservoir effect and the timing of the late-Glacial/Early Holocene humid phase in the Atacama Desert (northern Chile). *Quat. Res.* 52 (2), 143-153.
- Gosse, J.C., Phillips, F.M. 2001, Terrestrial in situ cosmogenic nuclides: theory and application. *Quat. Sci. Rev.* 20 (14), 1475-1560.
- Grosjean, M., van Leeuwen, J.F.N., van der Knaap, W.O., Geuh, M.A., Amman, B., Tanner, W., Messerli, B., Nunez, L.A., Valero-Garcés, B.L., Viet, H., 2001. A 22,000 ^{14}C year BP sediment and pollen record of climate change from Laguna Miscanti (23 °S), northern Chile. *Global Planet. Change* 28 (1-4), 35–51.
- Guest, J. E., 1969. Upper Tertiary ignimbrites in the Andean Cordillera, a part of the Antofagasta province northern Chile. *Geol. Soc. Am. Bull.* 80 (3), 337-372.
- Gunster, N., Skowronek, A., 2001. Sediment–soil sequences in the Granada Basin as evidence for

- long- and short-term climatic changes during the Pliocene and Quaternary in the Western Mediterranean. *Quat. Int.* 78 (1), 17-32.
- Guyodo, Y., Valet, J.-P., 1999. Global changes in intensity of the earth's magnetic field during the past 800 kyr. *Nature* 399, 249–252.
- Hall, S.R., Farber, D.L., Audin, L., Finkel, R.C., Mériaux, A.-S., 2008. Geochronology of pediment surfaces in southern Peru: Implications for Quaternary deformation of the Andean forearc. *Tectonophysics* 450 (1-4), 186-205.
- Hancock, G.S., Anderson, R.S., Chadwick, O.A., Finkel, R.C., 1999. Dating fluvial terraces with ^{10}Be and ^{26}Al profiles: application of the Wind River, Wyoming. *Geomorphology* 27 (1-2), 41-60.
- Hartley A.J., Chong, G., 2002. Late Pliocene age for the Atacama Desert: implications for the desertification of western South America. *Geology* 30 (1), 43–46.
- Hartley, A., Flint, S., Turner, P., Jolley, E.J., 1992. Tectonic controls on the development of a semi-arid, alluvial basin as reflected in the stratigraphy of the Purilactis Group (Upper Cretaceous-Eocene), northern Chile. *J. South Am. Earth Sci.* 53 (3-4), 275-296.
- Hartley, A.J., Mather, A.E., Jolley, E., Turner, P., 2005. Climatic controls on alluvial-fan activity, Coastal Cordillera, northern Chile. In: Harvey, A.M., Mather, A.E., Stokes, M. (Eds.), *Alluvial Fans: Geomorphology, Sedimentology, Dynamics*. Geol. Soc. Spec. Pub. 251, London, U.K., pp. 95-115.
- Harvey, A.M., 1984. Aggradation and dissection sequences on Spanish alluvial fans: influence on morphological development. *Catena* 11, 289-304.
- Harvey, A.M., 1989. The occurrence and role of arid zone alluvial fans. In: Thomas, D.S.G. (Ed.), *Arid Zone Geomorphology*, Belhaven Press, London, pp. 136-158.
- Harvey, A.M., 1990. Factors influencing Quaternary alluvial fan development in southeast Spain. In: Rachoki, A.H., Church, M. (Eds.), *Alluvial Fans: a Field Approach*. Wiley, Chichester, pp. 247-269.
- Harvey, A.M., 1996. The role of alluvial fans in the mountain fluvial systems of southeast Spain: implications of climatic change. *Earth Surf. Proc. Land.* 21, 543-553.
- Harvey, A.M., 1997. The role of alluvial fans in arid-zone fluvial systems. In: Thomas, D.S.G. (Ed.), *Arid Zone Geomorphology; Process, Form and Change in Drylands*. Belhaven Press, London, pp. 231-259.
- Harvey, A.M., and Wells, S.G., 1994. Late Pleistocene and Holocene changes in hillslope sediment supply to alluvial fan systems: Zzyzx, California. In: Millington, A.C., Pye, K. (Eds.), *Environmental Change in Drylands: Biogeographical and Geomorphological Perspectives*. Wiley, London, pp. 67-84.
- Harvey, A.M., Silva, P.G., Mather, A.E., Goy, J.L., Stokes, M., Zazo, C., 1999a. The impact of Quaternary sea-level and climatic change on coastal alluvial fans in the Cabo de Gata ranges, southeast Spain. *Geomorphology* 28, 1-2.

- Harvey, A.M., Wigand, P.E., Wells, S.G., 1999b. Response of alluvial fan systems to the late Pleistocene to Holocene climatic transition: contrasts between the margins of pluvial Lakes Lahontan and Mojave, Nevada and California, USA. *Catena* 36, 255-281.
- Harvey, A.M., Mather, A.E., Stokes, M., 2005. Alluvial fans: geomorphology, sedimentology, dynamics – introduction. In: Harvey, A.M., Mather, A.E., Stokes, M. (Eds.), *Alluvial Fans: Geomorphology, Sedimentology, Dynamics*. Geol. Soc. Spec. Pub. 251, London, U.K., pp. 1-7.
- Haug, E.W., Kraal, E.R., Sewall, J.O., Dijk, M.V., Chong Diaz, G., 2010. Climatic and geomorphic interactions on alluvial fans in the Atacama Desert, Chile. *Geomorphology* 121 (3-4), 184-196.
- Hedrick, K., Owen, L.A., Rockwell, T.K., Meigs, A., Costa, C., Caffee, M.W., Masana, E., Ahumada, E., 2013. Timing and nature of alluvial fan and strath terrace formation in the Eastern Precordillera of Argentina. *Quat. Sci. Rev.* 80, 143-168.
- Hidy, A.J., Gosse, J.C., Pederson, J.L., Mattern, J.L., Finkel, R.C., 2010. A geologically constrained Monte Carlo approach to modeling exposure ages from profiles of cosmogenic nuclides: an example from Lees Ferry, Arizona. *Geochem. Geophys.* 11 (9), Q0AA10.
- Hoke, G.D., Isacks, B.L., Jordan, T.E., Yu, J.S., 2004. Groundwater-sapping origin for the giant quebradas of northern Chile. *Geology* 32 (7), 605-608.
- Houston, J., 2006. The great Atacama flood of 2001 and its implications for Andean hydrology. *Hydrol. Processes* 20 (3), 591-610.
- Houston, J., Hartley, A.J., 2003. The Central Andean West-slope Rainshadow and its Potential contribution to the Origin of Hyper-aridity in the Atacama Desert. *Int. J. Climatol.* 23 (12), 1453-1464.
- Irwin, R.P., Tooth, S., Craddock, R.A., Howard, A.D., Baptista de Latour, A., 2014. Origin and development of theater-headed valleys in the Atacama Desert, northern Chile: Morphological analogs to martian valley networks. *Icarus* 243, 296-310.
- Jain, M., Choi, J.H., Thomas, P.J., 2008. The ultrafast OSL component in quartz: Origins and Implications. *Radiat. Meas.* 43 (2-6), 709-714.
- Jolley, E.J., Turner, P., Williams, G.D., Hartley, A.J., Flint, S., 1990. Sedimentological response of an alluvial system to Neogene thrust tectonics, Atacama Desert, northern Chile. *J. Geol. Soc.* 147 (5), 769-784.
- Jordan, T.E., Mpodozis, C., Munoz, N., Blanco, N., Pananont, P., Gardeweg, M., 2007. Cenozoic subsurface stratigraphy and structure of the Salar de Atacama Basin, northern Chile. *J. South Am. Earth Sci.* 23 (2-3), 122-146.
- Jungers M.C, Heimsath, A.M., Amundson, R., Balco, G., Shuster, D., Chong, G., 2013. Active erosion-deposition cycles in the hyperarid Atacama Desert of Northern Chile. *Earth Planet. Sci. Lett.* 371-372, 125-133.
- Lal, D., 1991. Cosmic ray labeling of erosion surfaces: in situ nuclide production rates and erosion models. *Earth Planet. Sci. Lett.* 104 (2-4), 424-439.

- 994
995 Latorre, C., Betancourt, J.L., Rylander, K.A., Quade, J., 2002. Vegetation invasions into Absolute
996 Desert: A 45,000-yr rodent midden record from the Calama-Salar de Atacama Basins, northern
997 Chile (22-24°S). *Geol. Soc. Am. Bull.* 114 (3), 349-366.
998
999 Latorre, C., Betancourt, J.L., Rylander, K.A., Quade, J., Matthei, O., 2003. A 13.5-kyr vegetation
1000 history from the arid prepuna of northern Chile (22–23°S). *Palaeogeogr. Palaeoclimatol.*
1001 *Palaeoecol.* 194, 223–246.
1002
1003 Latorre, C., Betancourt, J.L., Rech, J.A., Quade, J., Holmgren, C., Placzek, C., Maldonado, A.J.C., Vuille,
1004 M., Rylander, K., 2005. Late Quaternary history of the Atacama Desert. In Smith, M., Hesse, P.
1005 (Eds.), *23°S: the Archeology and Environmental History of the Southern Deserts*, pp. 73–90.
1006 National Museum of Australia Press, Canberra, Australia.
1007
1008 Latorre, C., Betancourt, J.L., Arroyo, M.T.K., 2006. Late Quaternary vegetation and climate history of a
1009 perennial river canyon in the Rio Salado basin (22°S) of northern Chile. *Quat. Res.* 65, 450-466.
1010
1011 Linters, J.D., Cook, K.H., 1997. On the origin of the Bolivian High and related circulation features of
1012 the South American climate. *J. Atmos. Sci.* 54, 656-677.
1013
1014 Lifton, N., Sato, T., Dunai, T.J., 2014. Scaling in situ cosmogenic nuclide production rates using
1015 analytical approximations to atmospheric cosmic-ray fluxes. *Earth Planet. Sci. Lett.* 386, 149-
1016 160.
1017
1018 Lindsay, J.M., de Silva, S., Trumbull, R., Emmermann, R., Wemmer, K., 2000. La Pacana caldera, N.
1019 Chile: A re-evaluation of the stratigraphy and volcanology of one of the world's largest resurgent
1020 calderas. *J. Volcanol. Geotherm. Res.* 106 (1-2), 145-173.
1021
1022 Lisiecki, L. E., Raymo, M. E., 2005. A Pliocene-Pleistocene stack of 57 globally distributed benthic
1023 $\delta^{18}\text{O}$ records. *Paleoceanography* 20 (1), PA1003.
1024
1025 Lowenstein, T.K., Hein, M.C., Bobst, A.L., Jordan, T.E., Ku, T.L., Luo, S., 2003. An assessment of
1026 stratigraphic completeness in climate-sensitive closed-basin lake sediments: Salar de Atacama,
1027 Chile. *J. Sediment. Res.* 73 (1), 91-104.
1028
1029 Maldonado, A., Betancourt, J.L., Latorre, C., Villagran, C., 2005. Pollen analyses from a 50,000-yr rodent
1030 midden series in the southern Atacama Desert (25°30'). *J. Quaternary Sci.* 20, 493-507.
1031
1032 Marinovic, N., Lahsen, A., 1984. Geología de la Hoja Calama, Región de Antofagasta. Carta
1033 Geológica de Chile, Escala 1:250.000, 58. Servicio Nacional de Geología y Minería, Santiago de
1034 Chile, p. 140.
1035
1036 Marrero, S.M., Phillips, F.M., Caffee, M., Gosse, J., 2015a. CRONUS-Earth cosmogenic ^{36}Cl
1037 calibration. *Quat. Geochronol.* 31, 199-219.
1038
1039 Marrero, S.M., Phillips, F.M., Borchers, B., Lifton, N., Aumer, R., Balco, G., 2015b. Cosmogenic
1040 Nuclide Systematics and the CRONUScalc Program. *Quat. Geochronol.* 31, 160-187.
1041
1042 Mather, A.E., Hartley, A.J., 2005. Flow events on a hyper-arid alluvial fan: Quebrada Tambores, Salar
1043 de Atacama, northern Chile. In: Harvey, A.M., Mather, A.E., Stokes, M. (Eds.), *Alluvial Fans:*
1044 *Geomorphology, Sedimentology, Dynamics*. *Geol. Soc. Spec. Pub.* 251, London, U.K., pp. 9-29.

- Matmon, A., Nichols, K., Finkel, R., 2006. Isotopic insights into smoothening of abandoned fan surfaces, Southern California. *Quat. Res.* 66 (1), 109-118.
- McDonald, E.V., McFadden, L.D., Wells, S.G., 2003. Regional response of alluvial fans to the Pleistocene to Holocene climatic transition, Mojave Desert, California. In: Enzel, Y., Wells, S.G., Lancaster, N. (Eds.), *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, Geol. Soc. Spec. Pap. 368, Boulder, Colorado, pp. 189-205.
- McFadden, L.D., Ritter, J.B., Wells, S.G., 1989. Use of multiparameter relative-age methods for age estimation and correlation of alluvial fan surfaces on a desert piedmont, Eastern Mojave Desert, California. *Quat. Res.* 32 (3), 276-290.
- McPhillips, D., Bierman, P.R., Crocker, T., Rood, D.H., 2013. Landscape response to Pleistocene-Holocene precipitation change in the Western Cordillera, Peru: ^{10}Be concentrations in modern sediments and terrace fills. *J. Geophys. Res.: Earth Surface* 118 (4), 2488-2499.
- Mejdahl, V. 1979. Thermoluminescence dating: beta-dose attenuation in quartz grains. *Archaeometry* 21 (1), 61-72.
- Mpodozis, C., Arriagada, C., Basso, M., Roperch, P., Cobbold, P., Reich, M., 2005. Late Mesozoic to Paleogene stratigraphy of the Salar de Atacama basin, Antofagasta, northern Chile: Implications for the tectonic evolution of the central Andes. *Tectonophysics* 399 (1-4), 125-154.
- Muñoz, N., Charrier, R., Reutter, K., 1997. Evolución de la Cuenca Salar de Atacama: Inversión tectónica y relleno de una cuenca de antepaís de retroarco. *Proceedings 8th Congreso Geológico Chileno* 1, 195-199.
- Murray A.S., Wintle A.G., 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. *Radiat. Meas.* 32 (1) 57-73.
- Nalpas, T., Dabard, M.-P., Ruffet, G., Vernon, A., Mpodozis, C., Loi, A., Hérail, G., 2008. Sedimentation and preservation of the Miocene Atacama Gravels in the Perdinales-Chañaral Area, Northern Chile: Climatic or tectonic control?. *Tectonophysics* 459 (1-4), 161-173.
- Nicholas, A.P., Quine, T.A., 2007. Modeling alluvial landform change in the absence of external environmental forcing. *Geology* 35 (6), 527-530.
- Nishiizumi, K., Caffee, M.W., Finkel, R.C., Brimhall, G., Mote, T. 2005, Remnants of a fossil alluvial fan landscape of Miocene age in the Atacama Desert of northern Chile using cosmogenic nuclide exposure age dating. *Earth Planet. Sci. Lett.* 237 (3-4), 499-507.
- Ohno, M., Hamano, Y., 1992. Geomagnetic poles over the past 10,000 years. *Geophys Res. Lett.* 19 (16), 1715-1718.
- Owen, L.A., Frankel, K.L., Knott, J.R., Reynhout, S., Finkel, R.C., Dolan, J.F., Lee, J., 2011. Beryllium-10 terrestrial cosmogenic nuclide surface exposure dating of Quaternary landforms in Death Valley. *Geomorphology* 125 (4), 54-557.
- Owen, L.A., Clemmens, S.J., Finkel, R.C., Gray, H., 2014. Late Quaternary alluvial fans at the eastern end of the San Bernardino Mountains, Southern California. *Quat. Sci. Rev.* 87, 114-134.

1096
1097 Pananont, P., Mpodozis, C., Blanco, N., Jordan, T. E., Brown, L. D., 2004. Cenozoic evolution of the
1098 northwestern Salar de Atacama Basin, northern Chile. *Tectonics* 23 (6), TC6007.
1099
1100 Perkins, J.P., Finnegan, N.J., de Silva, S.L., 2015. Amplification of bedrock canyon incision by wind.
1101 *Nat. Geosci.* 8, 305-310.
1102
1103 Phillips, F.M., Zreda, M.G., Flinsch, M.R., Elmore, D., Sharma, P., 1996. A reevaluation of
1104 cosmogenic ³⁶Cl production rates in terrestrial rocks. *Geophys Res. Lett.* 23 (9), 949–952.
1105
1106 Phillips, F.M., Argento, D.C., Balco, G., Caffee, M.W., Clem, J., Dunai, T.J., Finkel, R., Goehring, B.,
1107 Gosse, J.C., Hudson, A.M., Jull, A.J.T., Kelly, M.A., Kurz, M., Lal, D., Lifton, N., Marrero,
1108 S.M., Nishiizumi, K., Reedy, R.C., Schaefer, J., Stone, J.O.H., Swanson, T., Zreda, M.G., 2015.
1109 The Cronus-Earth Project: A Synthesis. *Quat. Geochronol.* 31 (36), 119-154.
1110
1111 Pierson, T.C., Costa, J.E., 1987. A rheologic classification of subaerial sediment-water flows. *Geol.*
1112 *Soc. Am. Rev. Eng. Geol.* 7, 1-12.
1113
1114 Placzek, C., Quade, J., Patchett, P.J., 2006. Geochronology and stratigraphy of Late Pleistocene lake
1115 cycles on the Southern Bolivian Altiplano: implications for causes of tropical climate change.
1116 *Geol. Soc. Am. Bull.* 118 (5), 515–532.
1117
1118 Placzek C.J., Quade, J., Betancourt, J.L., Patchett, P.J., Rech, J.A., Latorre, C., Matmon, A., Holmgren,
1119 C., English, N.B., 2009. Climate in the dry central Andes over geologic, millennial, and
1120 interannual timescales. *Ann. Missouri. Bot. Gard.* 96 (3), 386-397.
1121
1122 Placzek, C.J., Matmon, A., Granger, D.E., Quade, J., Niedermann, S., 2010. Evidence for active
1123 landscape evolution in the hyperarid Atacama from multiple terrestrial cosmogenic nuclides.
1124 *Earth Planet. Sci. Lett.* 295 (1-2), 12–20.
1125
1126 Placzek, C.J., Quade, J., Patchett P.J., 2013. A 130 ka reconstruction of rainfall from the Bolivian
1127 Altiplano. *Earth Planet. Sci. Lett.* 363, 97-108.
1128
1129 Placzek, C.J., Granger, D.E., Matmon, A., Quade, J., Ryb, U., 2014. Geomorphic process rates in the
1130 central Atacama Desert, Chile: insights from cosmogenic nuclides and implications for the onset
1131 of hyperaridity. *Am. J. Sci.* 314 (10), 1462-1512.
1132
1133 Ponti, D.J., 1985. The Quaternary alluvial sequence of the Antelope Valley, California. *Geo. Soc. Spec.*
1134 *Pap.* 203, 79-96.
1135
1136 Pope, R.J.J., Wilkinson, K.N., 2005. Reconciling the roles of climate and tectonics in Late Quaternary fan
1137 development on the Spartan piedmont, Greece. In: Harvey, A.M., Mather, A.E., Stokes, M.
1138 (Eds.), *Alluvial Fans: Geomorphology, Sedimentology, Dynamics*. *Geol. Soc. Spec. Pub.* 251,
1139 London, U.K., pp. 133-152.
1140
1141 Prescott J.R., Hutton, J.T. 1994., Cosmic ray contributions to dose rates for luminescence and ESR
1142 dating: large depths and long-term time variations. *Radiat. Meas.* 23 (2-3), 497–500.
1143
1144 Quade, J., Rech, J.A., Betancourt, J.L., Latorre, C., Quade, B., Rylander, K.A., Fisher, T., 2008.

1145 Paleowetlands and regional climate change in the central Atacama Desert, northern Chile. *Quat.*
1146 *Res.* 69 (3), 343–360.
1147
1148 Ramírez, C.F., Gardeweg, M., 1982. Hoja Toconau, Región de Antofagasta: Carta Geológica de
1149 Chile, 1:250,000 Servicio Nacional de Geología y Minería de Chile, Boletín v. 54.
1150
1151 Raymo, M. E., Oppo, D.W., Curry, W., 1997. The mid Pleistocene climate transition: A deep sea
1152 carbon isotopic perspective. *Paleoceanography* 12(4), 546-559.
1153
1154 Rech, J.A., Quade, J., Betancourt, J.L., 2002. Late Quaternary paleohydrology of the central Atacama
1155 Desert (lat. 22–24° S), Chile. *Geol. Soc. Am. Bull.* 114 (3), 334-348.
1156
1157 Rech, J.A., Pigati, J.S., Quade, J., Betancourt, J.L., 2003. Re-evaluation of mid-Holocene deposits at
1158 Quebrada Purica, northern Chile. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 194, 207-222.
1159
1160 Rech, J.A., Currie, B.S., Michalski, G., Cowan, A.M., 2006. Neogene climate change and uplift in the
1161 Atacama Desert, Chile. *Geology* 34 (9), 761–764.
1162
1163 Ritter, J.B., Miller, J.R., Enzel, Y., Wells, S.G., 1995. Reconciling the roles of tectonism and climate
1164 in Quaternary alluvial fan evolution. *Geology* 23 (3), 245-248.
1165
1166 Schimmelpfennig, I., Benedetti, L., Finkel, R., Pik, R., Blard, P.H., Bourlés, D., Burnard, P.,
1167 Williams, A., 2009. Sources of in situ ³⁶Cl in Basaltic rocks. Implications for calibration of
1168 production rates. *Quat. Geol.* 4 (6), 441-461.
1169
1170 Schmitt, A.K., de Silva, S.L., Trumbull, R.B., Emmermann, R. 2001, Magma evolution in the Purico
1171 ignimbrite complex, northern Chile: evidence for zoning of a dacitic magma by injection of
1172 rhyolitic melts following mafic recharge. *Contrib. Mineral. Petr.* 140 (6), 680-700.
1173
1174 Shackleton, N.J., 2000. The 100,000-year ice-age cycle identified and found to lag temperature, carbon
1175 dioxide, and orbital eccentricity. *Science* 289 (5486), 1897-1902.
1176
1177 Spelz, R.M., Fletcher, J.M., Owen, L.A., Caffee, M.W., 2008. Quaternary alluvial-fan development,
1178 climate and morphologic dating of fault scarps in Laguna Salada, Baja California, Mexico.
1179 *Geomorphology* 102, 578-594.
1180
1181 Stone, J.O., 2000, Air pressure and cosmogenic isotope production. *J. Geophys. Res.* 105 (B10), 23753-
1182 23759.
1183
1184 Stone, J.O., Fifield, L.K., Allan, G.L., Cresswell, R.G., 1996. Cosmogenic chlorine-36 production
1185 from calcium spallation. *Geochim. Cosmochim. Acta.* 60 (1), 679-692.
1186
1187 Tsukamoto, S., Rink, W.J., Watanuki, T., 2003. OSL of tephric loess and volcanic quartz in Japan and an
1188 alternative procedure for estimating D_e from a fast OSL component. *Radiat. Meas.* 37 (4-5), 459-
1189 465.
1190
1191 Tsukamoto, S., Murray, A.S., Huot, S., Watanuki, T., Denby, P.M., Bøtter-Jensen, L., 2007.
1192 Luminescence property of volcanic quartz and the use of red isothermal TL for dating tephros.
1193 *Radiat. Meas.* 42 (2), 190-197.
1194
1195 Veit, H., May, J.-H., Madella, A., Delunel, R., Schlunegger, F., Szidat, S., Capriles, J.M., 2016. Palaeo-

- geoeological significance of Pleistocene trees in the Lluta Valley, Atacama Desert. *J. Quaternary Sci.* 31 (3), 203-213.
- Ventra, D., Nichols, G.J., 2014. Autogenic dynamics of alluvial fans in endoheric basins: Outcrop examples and stratigraphic significance. *Sedimentology* 61 (3), 767-791.
- Ventra, D., Díaz, G.C, De Boer, P.L., 2013. Colluvial sedimentation in a hyperarid setting (Atacama Desert, northern Chile): Geomorphic controls and stratigraphic facies variability. *Sedimentology* 60, 1257-1290.
- Viseras, C., Calvache, M.L., Soma, J.M., Fernandez, J. 2003. Differential Features of Alluvial Fans Controlled by Tectonic or Eustatic Accommodation Space. Examples From The Betic Cordillera, Spain. *Geomorphology* 50 (1-3), 181-202.
- Vuille, M., 1999. Atmospheric circulation anomalies over the Bolivian Altiplano during dry and wet periods and extreme phases of the Southern Oscillation. *Int. J. Climatol.* 19 (14), 1579-1600.
- Ward, D.J., Cesta, J.M., Galewsky, J., Sagredo, E., 2015. Late Pleistocene glaciations of the arid subtropical Andes and new results from the Chajnantor Plateau, northern Chile. *Quat. Sci. Rev.* 128, 98-116.
- Wells S.G., Harvey, A.M., 1987. Sedimentologic and geomorphic variations in storm generated alluvial fans, Howgill Fells, northwest England. *Bull. Geol. Soc. Am.* 98, 182-198.
- Wells, S.G., McFadden, L.D., Harden, J., 1990. Preliminary results of age estimations and regional correlations of Quaternary alluvial fans within the Mojave Desert of Southern California. In: Reynolds, R.E., Wells, S.G., Brady, R.J.I. (Eds.), *At the End of the Mojave: Quaternary Studies in the Eastern Mojave Desert*. Special Publications of the San Bernadino County Museum Association, Redlands, CA, pp. 45-53.
- Yang, S., Odah, H., Shaw, J., 2000. Variations in the geomagnetic dipole moment over the last 12000 years. *Geophys. J. Int.* 140 (1), 158-162.
- Zech, R., Kull, C., Veit, H., 2006. Late Quaternary glacial history in the Encierro Valley, northern Chile (29°S), deduced from 10Be surface exposure dating. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 234, 277-286.
- Zech, R., Kull, C., Kubik, P.W., Veit, H., 2007. Exposure dating of Late Glacial and pre-LGM moraines in the Cordon de Doña Rosa, Northern/Central Chile (~31°S). *Clim. Past.* 3, 1-14.
- Zweck, C., Zreda, M., Anderson, K., Bradley, E., 2012. The theoretical basis of ACE, an Age Calculation Engine for cosmogenic nuclides. *Chem. Geol.* 291, 199-205.

Figure Captions

Fig. 1. (A) Location map of the study area in the subtropical Andes of Chile. (B) Hillshade digital elevation model of the Salar de Atacama Basin and Chajnantor Plateau (ASTER 30-m resolution). Dashed box denotes the extent of the study area. Circles represent locations of prior studies discussed later in section 5.3. (C) Transect topography (black, m) through A-A'. Annual precipitation (blue, mm a⁻¹) derived from Tropical Rainfall Measurement Mission (TRMM) project 2B31 (Bookhagen and Strecker, 2008, 2012).

Fig. 2. Characteristic geomorphic features of the study area. (A) Ignimbrite shield of the Chajnantor Plateau ornamented with Cerro Negro and Cerro de Macón. Bedrock channels incised into the ignimbrite shield, some of which reach the glacial margin. Alluvial fan surface Qf2 of the Chajnantor bajada in the foreground. (B) Modern channel incising alluvial surface Qf2; sparse vegetation suggests ephemeral flow. Adjacent surfaces rise 0.5 to 3 m above the modern channel. (C) Quaternary gravel Qg2a preserved on the Chajnantor Plateau below Cerro Toco. Clasts are predominantly andesite lithology. Minimum deposit thickness of ~50 m observed in quarry exposure. (D) Bedrock channel carved ~4 m into the Cajón ignimbrite, sedimentary fill along channel base. (E) Debris flow deposits entrained in channels dissecting alluvial fan surface Qf2 along the Chajnantor bajada. (F) Boulder field associated with a debris flow deposit (Qf3) along the Andean Flank adjacent to Cerro Lascar. Boulders range from 0.2 to 3 m.

Fig. 3. (A) Surficial geomorphology along the western flank of the Chajnantor Plateau including sampling locations, surface exposure ages, and luminescence ages. For regional context, refer to Fig. 1. (B-G) Insets of sample locations with ³⁶Cl surface exposure ages (red) and luminescence ages (blue).

Fig. 4. (A) Weathering characteristics measured on surface clasts of alluvial fan surface Qf1 and Qf2. Percentages are based on 100 surface clasts. Horizontal axis denotes arbitrary sampling location on the

respective surface. (B) Weathering characteristics for measured boulders on debris flow surface Qf3. Numbers above the histograms correspond to the number of boulders measured. Horizontal axis denotes arbitrary sampling location on the respective surface.

Fig. 5. (A) ^{36}Cl depth profile for alluvial surface Qf1 with profile sketch and sampling locations for context. Subplots represent variables used in the model simulation, rate of deposition (B) and timing of deposition (C). Modeled best-fit ^{36}Cl depth profile (black) for 25 ka of deposition at a rate of 2.4 mm a^{-1} with zero inheritance. The profile suggests an exposure age of ca. 200 ka. Surface samples are AFDP-1 and AFDP-2, and depth samples are AFDP-35 through 600.

Fig. 6. Distribution of ^{36}Cl surface exposure ages for geomorphic surfaces along the western flank of the Chajnantor Plateau, open squares represent the mean and 1σ uncertainty. Hollow circles denote ages that may be suspect. Distribution of ^{10}Be concentrations converted to exposure ages along the margins of the Salar de Atacama described in Placzek et al. (2014). Triangles represent quartz clasts; inverted triangles denote alluvial sediment and channel gravels. Refer to text for context. Probability density ‘camel’ plots for regional glacial histories of (17) Blard et al. (2014), (18) Ward et al. (2015), (19) Zech et al. (2006), and (20) Zech et al. (2007b) with rescaled values from Ward et al. (2015). Sedimentary hydrologic proxy records from Salar de Atacama (Bobst et al., 2001) and Salar de Uyuni (Fornari et al., 2001) with ages after Placzek et al. (2013). Lake cycles from the Bolivian Altiplano as described in Placzek et al. (2013). 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.

Supplementary Table 1. Table containing input data for age calculation with CRONUScalc.

Table 1

Table 1

Attributes, site information, AMS measured ^{36}Cl concentrations, and exposure ages for alluvial fan and debris flow boulder samples

Sample no.	Latitude (°S)	Longitude (°W)	Elevation (m)	Deposit ^a	Material	Thickness (cm)	Topographic shielding	Cl _{rock} (ppm)	$^{36}\text{Cl}/\text{Cl}$ (10 ⁻¹⁵)	^{36}Cl Inventory (10 ⁶ at g ⁻¹)	Exposure age ^b (ka)
AF-1-13	23.14525	68.00323	2627	AF	Andesite	3.0	0.999	10.7 ± 0.39	2974.66 ± 86.61	2.963 ± 0.088	135 ± 12
AF-2-13	23.14688	68.00283	2626	AF	Andesite	3.0	0.999	10.0 ± 0.43	3434.82 ± 72.03	3.164 ± 0.070	144 ± 13
AF-3-13	23.14417	68.01749	2498	AF	Andesite	3.0	0.999	25.9 ± 0.91	2925.17 ± 84.89	3.007 ± 0.097	137 ± 14
AF-4-13	23.14431	68.01879	2482	AF	Andesite	3.0	0.999	49.5 ± 2.36	2158.94 ± 78.61	3.048 ± 0.138	122 ± 12
AF-1-14	23.11884	68.03659	2440	AF	Andesite	3.0	0.999	32.4 ± 0.78	1108.43 ± 17.97	1.210 ± 0.023	49.6 ± 4.4
AF-2-14	23.11917	68.03588	2443	AF	Andesite	3.0	0.999	83.4 ± 0.79	1295.34 ± 20.33	2.489 ± 0.041	82 ± 11
AF-3-14	23.11884	68.01939	2469	AF	Andesite	3.0	0.999	73.7 ± 3.80	1043.27 ± 25.57	2.189 ± 0.084	72.1 ± 8.9
AF-5-14	23.27193	67.99696	2477	AF	Andesite	3.0	0.999	122.5 ± 4.34	2438.32 ± 35.88	6.259 ± 0.195	181 ± 26
AF-6-14	23.27420	67.97386	2644	QG	Andesite	3.0	0.999	133.7 ± 10.18	5325.59 ± 65.16	14.676 ± 0.901	540 ± 130
AFDP-1-14	23.17648	68.01833	2456	AF	Andesite	3.0	0.999	20.1 ± 0.71	3054.95 ± 64.77	3.182 ± 0.075	150 ± 13
AFDP-2-14	23.17556	68.01605	2468	AF	Andesite	3.0	0.999	13.4 ± 0.35	4033.35 ± 83.86	3.778 ± 0.081	194 ± 12
DF3-B1-13	23.14434	68.01790	2481	DF	Ignimbrite	4.0	0.999	344.5 ± 57.30	1124.9 ± 33.43	7.011 ± 1.071	110 ± 26
DF3-B2-13	23.14347	68.01958	2475	DF	Ignimbrite	5.0	0.999	345.5 ± 43.42	204.62 ± 6.62	1.273 ± 0.150	18.4 ± 3.7
DF1-B1-14	23.11797	68.04921	2409	DF	Ignimbrite	4.0	0.999	234.3 ± 4.88	336.15 ± 8.87	1.474 ± 0.047	26.3 ± 3.6
DF1-B2-14	23.11773	68.05075	2404	DF	Ignimbrite	5.0	0.999	287.2 ± 10.32	348.97 ± 8.56	1.832 ± 0.073	28.3 ± 4.3
DF1-B3-14	23.11505	68.06631	2375	DF	Andesite	2.0	0.999	153.8 ± 11.95	2454.30 ± 45.62	7.575 ± 0.497	229 ± 53
DF6-B1-14	23.22236	68.01277	2450	DF	Ignimbrite	2.5	0.999	192.4 ± 11.64	276.64 ± 6.14	1.022 ± 0.057	20.2 ± 2.6
DF6-B2-14	22.22234	68.01281	2434	DF	Ignimbrite	5.5	0.995	200.3 ± 13.03	188.32 ± 4.79	0.720 ± 0.044	12.4 ± 2.1
DF6-B3-14	23.22248	68.01318	2439	DF	Ignimbrite	3.5	0.999	276.0 ± 21.00	207.77 ± 4.54	1.050 ± 0.074	14.8 ± 2.5
<u>Depth Profile</u>											
AFDP-14-35	23.17689	68.01970	2450	AF	Andesite	3.0	N/A	9.8 ± 0.17	3182.55 ± 68.87	2.880 ± 0.062	N/A
AFDP-14-65	23.17689	68.01970	2450	AF	Andesite	3.0	N/A	13.6 ± 0.22	2317.45 ± 39.89	2.163 ± 0.037	N/A
AFDP-14-200	23.17689	68.01970	2450	AF	Andesite	3.0	N/A	17.6 ± 0.65	576.52 ± 11.75	0.589 ± 0.013	N/A
AFDP-14-400	23.17689	68.01970	2450	AF	Andesite	3.0	N/A	18.2 ± 0.78	174.49 ± 5.36	0.180 ± 0.006	N/A
AFDP-14-500	23.17689	68.01970	2450	AF	Andesite	3.0	N/A	13.9 ± 0.56	136.06 ± 4.04	0.126 ± 0.004	N/A
AFDP-14-600	23.17689	68.01970	2450	AF	Andesite	3.0	N/A	12.3 ± 0.36	123.64 ± 4.22	0.112 ± 0.004	N/A

^a Landform sampled, AF: alluvial fan, DF: debris flow, QG: Quaternary gravel.^b Exposure ages calculated by means of time-varying Lal/Stone scaling, uncertainties include 1 σ analytical uncertainty and propagated production rate scaling uncertainty. See text for further description.

Table 2

Table 2Major and trace element concentrations for alluvial fan cobbles and debris flow boulder samples^a

Sample no.	SiO ₂ (wt.%)	TiO ₂ (wt.%)	Al ₂ O ₃ (wt.%)	Fe ₂ O ₃ (wt.%)	MnO (wt.%)	MgO (wt.%)	CaO (wt.%)	Na ₂ O (wt.%)	K ₂ O (wt.%)	P ₂ O ₅ (wt.%)	CO ₂ (wt.%)	B (ppm)	Sm (ppm)	Gd (ppm)	U (ppm)	Th (ppm)	H ₂ O (wt.%)
AF-1-13	61.85	0.76	18.30	5.07	0.08	1.82	5.30	3.96	2.00	0.03	0.030	13.0	2.53	2.37	1.4	6.8	0.030
AF-2-13	62.65	0.74	18.27	4.85	0.07	1.82	5.24	3.99	2.11	0.02	0.000	13.0	2.96	2.41	1.4	7.3	0.000
AF-3-13	64.42	0.641	16.82	4.41	0.067	1.48	4.52	3.93	2.15	0.02	0.335	19.4	2.80	2.10	1.7	6.0	0.335
AF-4-13	64.10	0.666	16.63	4.82	0.073	1.57	4.72	3.68	2.01	0.03	0.315	20.0	2.90	2.00	1.6	5.9	0.315
AF-1-14	63.92	0.677	17.54	4.77	0.070	1.77	5.11	3.95	2.06	0.05	0.385	19.8	2.80	2.00	1.7	5.7	0.385
AF-2-14	62.78	0.729	17.69	5.23	0.072	2.20	5.53	3.85	1.91	0.06	0.395	14.3	3.00	2.00	1.4	4.9	0.395
AF-3-14	62.00	0.630	18.56	4.29	0.070	1.65	5.47	3.96	2.01	0.05	0.550	20.0	2.84	2.46	1.8	5.8	0.550
AF-5-14	63.18	0.673	17.28	5.39	0.077	2.27	5.42	3.51	2.29	0.05	0.375	15.9	3.40	2.70	3.7	10.4	0.375
AF-6-14	63.02	0.691	16.54	5.54	0.079	2.70	5.58	3.43	2.20	0.08	0.365	25.4	3.80	2.60	3.7	9.8	0.365
AFDP-1-14	63.00	0.63	19.13	4.19	0.07	1.63	5.69	4.09	2.03	0.02	0.000	12.0	2.65	2.18	1.5	6.3	0.000
AFDP-2-14	62.67	0.65	19.01	4.24	0.07	1.62	5.59	4.07	2.03	0.02	0.000	14.0	2.72	2.21	1.8	6.1	0.000
DF3-B1-13	60.16	1.069	14.34	8.71	0.124	3.26	5.07	2.64	2.45	0.17	0.550	31.1	6.30	4.90	6.3	12.1	0.550
DF3-B2-13	59.64	1.107	15.43	8.92	0.133	3.27	5.45	3.08	2.50	0.02	0.460	28.5	6.10	5.10	5.5	11.9	0.460
DF1-B1-14	63.62	0.999	13.67	7.87	0.117	3.13	4.13	2.61	3.24	0.04	0.520	25.3	6.00	4.70	7.9	17.0	0.520
DF1-B2-14	62.58	1.043	14.23	8.05	0.113	2.99	4.24	2.69	3.15	0.04	0.420	20.7	5.80	4.50	7.6	17.1	0.420
DF1-B3-14	60.48	0.750	17.47	6.19	0.096	2.83	5.86	3.46	1.85	0.04	0.385	13.2	3.20	2.30	1.4	4.2	0.385
DF6-B1-14	65.98	0.491	17.55	3.49	0.055	0.91	4.32	3.91	2.79	0.01	0.265	17.1	5.10	3.70	4.4	13.9	0.265
DF6-B2-14	67.40	0.555	16.67	3.74	0.054	0.94	4.01	3.86	3.17	0.02	0.170	10.5	5.10	3.70	5.2	15.5	0.170
DF6-B3-14	64.83	0.587	17.37	4.14	0.069	1.49	4.55	3.40	3.27	0.03	0.350	14.3	5.40	3.00	4.4	21.3	0.350
Depth Profile																	
AFDP-14-35	62.44	0.60	18.66	4.02	0.06	1.53	5.54	4.01	2.04	0.03	0.400	13	2.47	2.22	1.8	6.5	0.400
AFDP-14-65	62.42	0.61	18.21	3.93	0.06	1.49	5.32	4.01	2.13	0.03	0.800	13	2.60	2.32	2.1	7.7	0.800
AFDP-14-200	61.82	0.62	18.76	4.17	0.06	1.63	5.55	4.03	2.00	0.02	0.550	13	2.68	2.32	1.5	6.3	0.550
AFDP-14-400	61.26	0.61	19.38	4.16	0.07	1.50	5.87	4.10	1.96	0.02	0.400	12	2.57	2.16	1.5	5.8	0.400
AFDP-14-500	61.18	0.65	18.95	4.35	0.07	1.53	5.72	4.03	1.98	0.03	0.650	12	3.09	2.39	1.6	7.1	0.650
AFDP-14-600	62.07	0.59	19.33	3.87	0.06	1.40	5.67	4.09	2.00	0.03	0.350	13	3.07	2.11	1.4	5.5	0.350

^aElemental concentrations from Na₂O₂ Fusion and ICP-MS of whole sediment measured at Activation Laboratories, Ontario, and Bureau Veritas Commodities Canada Ltd., British Columbia. Uncertainties are taken to be ± 10%.

Table 3

Table 3

Elemental concentrations, water content, and cosmic dose rate for total dose rate estimation

Sample no.	K (%) ^a	U (ppm) ^a	Th (ppm) ^a	Water content (%) ^b	Cosmic (μGy/ka) ^c	Total dose rate (Gy/ka) ^d
Chaj-OSL1	1.61 ± 0.08	2.50 ± 0.25	9.20 ± 0.17	5 ± 0.5	240 ± 24	2.85 ± 0.14
Toc-OSL1	1.36 ± 0.07	2.80 ± 0.28	8.20 ± 0.37	5 ± 0.5	149 ± 15	2.53 ± 0.13

^aElemental concentrations from FUS-ICP of whole sediment measured at Activation Laboratories, Ontario. Uncertainties are taken to be ±10 %.

^bWater content taken as 5% to account for the uncertainty over the timescale.

^cCosmic dose rate calculated using the methods of Prescott and Hutton (1994).

^dMethods of Adamiec and Aitken (1998) and Mejdahl (1979) used to combine cosmic dose rate with β and γ doses from elemental concentration to calculate the total dose rate.

Table 4

Table 4
Sample location, equivalent dose (D_E), dose rate, and calculated ages for OSL samples; measurements used the single aliquot regenerative dose (SAR) method; ages calculated by means of software by Rainer Grün and DRAC (Durcan et al., 2015)

Sample no.	Latitude (°S)	Longitude (°W)	Elevation (m)	Depth (cm)	Disks used *	Equivalent dose w/ ±1 σ error (Gy) Weighted Mean	Dose rate (Gy/ka)	Age w/ ±1 σ error (ka) Weighted Mean
Chaj-OSL1	23.06929	67.92152	3443	200	22(40)	115 ± 1.1	2.85 ± 0.14	40.51 ± 2.00
Toc-OSL1	23.17689	68.01970	2450	380	23(40)	108.2 ± 1.0	2.53 ± 0.13	42.86 ± 2.03

*Total number of aliquots measured listed inside the parentheses, number of aliquots that satisfy recuperation criteria and are considered for age calculation listed outside the parentheses.

Figure 1(Color)

Figure 2(Color)

Figure 3(Color)

Figure 4(Color)

Figure 5(Color)

Figure 6(Color)

Figure 1(Greyscale)

Figure 2(Greyscale)

Figure 3(Greyscale)

Figure 4(Greyscale)

Figure 5(Greyscale)

Figure 6(Greyscale)