Late Miocene increase in precipitation in the Western Cordillera of the Andes between 18-19°S latitudes inferred from shifts in sedimentation patterns

Fritz Schlunegger¹, Kevin P. Norton², Romain Delunel¹, Todd A. Ehlers³, Andrea Madella¹

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Surface Uplift and Climate Change in the Andes of Northern Chile

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Abstract

Modern climate in the Andes is characterized by strong N-S decreasing trends in precipitation rates. Here we use stratigraphic records to show that this pattern has been established since as early as 12-11 Ma, at least on the western Andean margin of Northern Chile. The stratigraphic architecture on the western Andean margin documents a transition between 19°-20°S latitude where matrix-supported debris flow deposits shift to fluvial conglomerates between 12-11 Ma. The deposition of fluvial sediments has been maintained to the present north of 19°-20°S, while the occurrence of post 11 Ma aeolian sand, matrix-supported breccias with conglomerate interbeds south of these latitudes implies ongoing sedimentation with less water and thus under drier conditions. We relate these changes to the tectonic development of the Andes. Existing palaeoclimate models suggest that an elevated plateau deflects the Andean jet towards the south, thereby focusing moisture from the equatorial Atlantic to the northeastern flanks of the Altiplano. In addition, the formation of the eastern Andean foothills most likely intercepted moisture transport, and shifted it farther to the east, thereby keeping the western Andean margin dry south of 19°-20°S latitudes. The sedimentological

35 data support a strong linkage between orographic precipitation and stratigraphy
36 whereby central Andean deformation controls the distribution of available
37 moisture on the western flank through a combination of orographic precipitation
38 and deflection of air masses.

39

40 **1. Introduction**

41 The central Andes between c. 16°S and 24°S latitude are characterized by large
42 climate contrasts with tropical conditions on the eastern side, and an extremely dry
43 climate on the western margin (Fig. 1A). Likewise, precipitation rates reveal large
44 N-S gradients, resulting in semiarid conditions on the western margin in northern
45 Peru and in a hyperarid climate in northern Chile. These trends where modern
46 precipitation rates decrease from N to S are particularly evident in water budgets,
47 where annual precipitation north of 17°S is large enough to supply the perennial
48 Lago Titicaca (Fig. 1B). However, dryer conditions farther south have resulted in
49 the establishment of salt pans in the Lago Poopó at 19°S and the accumulation of
50 evaporites in the Salar de Uyuni at 21°S. The effects of such precipitation
51 gradients are also evident on the western Andean margin (Fig. 1B), with the result
52 that modern runoff of streams with sources in the Western Cordillera (Houston,
53 2006) shifts from perennial north of 19°-20°S, to semi-perennial and ephemeral
54 south of 21°S, as reported by Nester et al. (2007, see their SI Table 2). Although
55 evidence for N-S precipitation gradients exist, the timing of their formation, and
56 the mechanisms through which these changes occurred have been contested.

57 In the past years, sedimentological data from the endorheic Central Depression,
58 situated south of c. 19.5°S between the Western and Coastal Cordilleras (Fig. 2),
59 have been used to infer a semi-arid/arid climate on the western Andean margin at
60 least since Late Miocene times, and possibly earlier (Jordan et al., 2014). In this
61 context, Armijo et al. (2015) interpreted that shifts to hyperarid conditions in
62 northern Chile occurred as a response to global cooling after the Mid-Miocene
63 climate optimum (Zachos et al., 2001). For the time after 4 Ma, Hartley (2003) and
64 also Jordan et al. (2014) suggested that a further enhancement of the already
65 existing hyperarid conditions in western South America established as
66 consequence of an additional phase of global climate cooling, but both authors did
67 not rule out the possibility of orographic effects on aridisation. Based on
68 sedimentological analyses carried out on deposits in the Central Depression south

69 of 20°S latitude, Jordan et al. (2014) proposed that the sediments record a 12-11
70 Ma-old change towards hyperarid conditions, which they tentatively related to
71 amplifications of rain shadow effects in response to the rise of the Andes.
72 However, it was also suggested that the rise of the Altiplano should cause an
73 increase in precipitation rates through a mechanism where an elevated plateau
74 leads to stronger insolation and thus to a positive feedback between greater
75 atmospheric convection on the Altiplano and attraction of more moisture from the
76 Amazon basin (Strecker et al., 2007). If correct, then sedimentary archives with
77 sources in the Western Cordillera and the adjacent Altiplano Plateau are expected
78 to record an increase in available moisture and precipitation rates and not a shift
79 towards drier conditions. It is the scope of this paper to explore this hypothesis and
80 to propose a mechanism for explaining possible spatially variable responses.

81 Here, we investigate stratigraphic records of precipitation change in the Andes of
82 Northern Chile. We focus on the time span between c. 25 Ma and the present
83 when the Andes rose to elevations high enough to modify and intercept moisture
84 pathways (Strecker et al., 2007; Ehlers and Poulsen, 2009), and when the
85 protrusion of the Andean foothills on the eastern margin, as revealed by e.g., Kley,
86 (1999) and Eichelberger et al. (2013), could have deviated the moisture source
87 farther to the east. Our investigations focus on stratigraphic archives of material
88 deposited in the western Andean margin, but derived through erosion from the
89 Western Cordillera (Figs. 1, 2). We use the sedimentary structures of these
90 deposits to reconstruct spatio-temporal shifts of precipitation rates in the source
91 areas. We conclude that in the Western Andean margin of Chile, the modern N-S
92 precipitation gradient and related water budgets (Fig. 1) has been recorded since as
93 early as 12-11 Ma, and that the establishment of this gradient might have been
94 conditioned by the history of crustal deformation of the eastern Andean margin
95 and particularly in response to the rise of the Altiplano Plateau.

96

97 **2. Setting**

98 In northern Chile, the western margin of the Andes represents a gently folded
99 ramp that connects the Pacific coast with the Western Cordillera, which forms the
100 modern arc and hosts andesitic stratovolcanoes of Miocene to modern ages
101 (Wörner et al. 2000). The Altiplano, situated east of the Western Cordillera (Fig.
102 1), is a highly elevated plateau with gentle hills at approximately 4 km a.s.l. and

103 extends ca. 2000 km in N-S direction and 400 km across strike. The Coastal
 104 Cordillera (labeled as ‘CC’ on Fig. 2) delineates the Andean margin to the west
 105 adjacent to the Pacific Ocean. In northern Chile, this morphotectonic domain is
 106 underlain by Mesozoic tholeiitic arc rocks and marine backarc sediments
 107 (Allmendinger et al. 2005; García and Hérail 2005). The Central Basin situated
 108 between c. 18°-19°S in the Tacna-Arica area (Fig. 2), and the Central Depression
 109 north of c. 20°S host an Oligo/Miocene suite of volcano-sedimentary rocks
 110 derived from the Western Cordillera (see e.g. Madella et al., 2016). Both
 111 sedimentary basins are located between the Coastal Cordillera and the Western
 112 Cordillera. Mapping reveals that the Central Basin comprises the Azapa (Az on
 113 Fig. 2), Oxaya (Ox), and El Diablo Formations (ED), while the time equivalent
 114 deposits in the Central Depression are mainly referred to as the Alto de Pica
 115 Formation (AdP on Fig. 2) (Farías et al., 2005; Jordan et al., 2014). On the
 116 foothills of the Western Cordillera, locally-sourced deposits are preserved in the
 117 Late Miocene Huyalas Formation in the Arica region (18°S latitude, Hy on Figure
 118 2; García et al., 2004), in the Lauca Basin on the eastern margin of the Western
 119 Cordillera (Fig. 2; Gaupp et al., 1999), and in the what has been referred to as the
 120 ‘Piedmont unit’ by Jordan et al. (2014) in the Alto de Pica area at c. 22°S.
 121 The depositional surfaces particularly of the Oxaya, El Diablo and Alto de Pica
 122 Formations form pediplains or ‘pampas’ that can be correlated several tens of
 123 kilometres along strike. The pampas have been incised by perennial and ephemeral
 124 streams with sources on the Altiplano and in the Western Cordillera since ca. 11
 125 Ma or possibly later (Hoke et al. 2007; Jordan et al., 2010), leading to the
 126 formation of >1500 m deep valleys or ‘Quebradas’. Progressive downcutting also
 127 resulted in the accumulation of strath terrace conglomerates, particularly in the
 128 Lluta valley at 18°S (Fig. 2).
 129 Modern precipitation on the Altiplano is linked to the position of the Inter-
 130 Tropical Convergence Zone (ITCZ) and the Andean jet that transfers humidity
 131 from the Atlantic Ocean to the Subandes via the Amazon Basin (lower left inset of
 132 Fig. 1A). Most precipitation falls along the foothills of the Subandes through
 133 orographic effects, and along the Santa Cruz bend and the Boomerang Hills (Fig.
 134 1A) that deflect the jet to the east (Campetella and Vera, 2002). This results in
 135 strong N-S decreasing trends of precipitation rates on the eastern Andean margin,
 136 but also on the Altiplano Plateau and the western Andean Cordillera (lower right

inset of Fig. 1A). The consequences of these precipitation gradients on annual precipitation budgets are illustrated on Figure 1B. It shows that modern annual water budgets drastically decrease from 17°S (latitude of Lago Titicaca) to 21°S (Salar de Uyuni). In addition, the upwelling of the cold Peruvian Current along the Pacific Coast creates a persistent temperature inversion that traps any Pacific moisture (Houston, 2006), resulting in hyperarid conditions along the Pacific coast.

144

145 **3. Datasets and Methods**

This paper is mainly based on sedimentological data from the western Andean margin of northern Chile between c. 18°-21°S latitude, where effects related to N-S precipitation gradients are recorded by the sedimentary archives. We start with the presentation of the sedimentological record at Francia (Fig. 2), and expand related observations along strike to extend the chronological record of change from N to S between 18° and 21°S latitudes. The sedimentary structures of deposits preserved in this section are analyzed at a scale of 1:250, where 10 meters of stratigraphic section was generalized to a 4 cm-long sedimentological log in the field. Sedimentological logging is based on field photos and standard sedimentological techniques. The deposits are interpreted in terms of changes in surface runoff. In addition, we re-calibrate the chronological record of the Francia section originally presented by von Rotz et al. (2005) based on additional ages obtained from interbedded tuffs in the region (Blanco and Tomlinson, 2013) that were recently synthesized by Jordan et al. (2014, see their Table 1).

160

161 **3.1. Revised chronology of the Francia section**

The Francia section comprises a 225-m-thick Late Miocene suite of sandstones and conglomerates with mudstone interbeds (Fig. 3A, Pinto et al., 2004; von Rotz et al., 2005; Farías et al., 2005; Hartley and Evenstar, 2010). A chronology of the section (El Diablo Formation) was established by von Rotz et al. (2005) through the magnetostratigraphic analyses of 35 palaeomagnetic sites, yielding a total of 12 reversals (Fig. 3B). These authors assigned ages between 11 and 8 Ma for the magnetopolarity stratigraphy (MPS) of this section, based on their preferred correlation of the Francia MPS with the magneto polarity time scale (MPTS). However, additional chronological data by Farías et al. (2005) and a recent

171 synthesis of published and new ages for the El Diablo Formation (Jordan et al.,
 172 2014) requires a revision of the Francia MPS, the results of which are shown on
 173 Figure 3C. Similar to Hartley and Evenstar (2010), we bracket the ages of the
 174 Francia section through (i) the 15.7 ± 0.7 Ma whole rock K-Ar age of a reworked
 175 tuff at the base of the formation near Camarones (Fig. 2) (García and Hérail,
 176 2005), (ii) the 16.2 ± 0.7 Ma whole rock K-Ar age assigned to the Ignimbrita Suca
 177 underlying this unit in the region (Muñoz and Sepúlveda, 1992), and particularly
 178 by the 19.3 ± 0.8 Ma age for the Ignimbrita Moquilla c. 13 km upstream of the
 179 Francia section in the same valley (Pinto et al., 2004) (Fig. 2), (iii) the 8.2 ± 0.5 Ma
 180 (Muñoz and Sepúlveda, 1992) and 9.0 ± 1 Ma (Naranjo and Paskoff, 1985) whole
 181 rock K-Ar ages of the Lava de Tana overlying the top units of the El Diablo
 182 Formation c. 5 km farther north of Francia (Pinto et al., 2004), and (iv) the
 183 11.7 ± 0.4 Ma whole rock K-Ar age assigned to ignimbrites interbedded within the
 184 El Diablo Formation c. 47 km farther to the southwest (Farías et al., 2005) (Fig. 2).
 185 We additionally benefit from the suite of 12 reversals encountered at Francia (von
 186 Rotz et al., 2005) plus numerous new ages of ash deposits embedded in this unit
 187 farther to the south, which were recently summarized by Jordan et al. (2014) in
 188 their Table 1. These constraints suggest an age of c. 13 Ma for the base of the
 189 section, and an age of approximately 11 Ma for the upper termination of the
 190 sedimentary suite (Figure 3C), consistent with the revised chronological
 191 framework for the region (Jordan et al., 2014). Note that we have not dated the
 192 volcano-clastic unit in the middle of the section because rip-up clasts and
 193 embedded boulders indicate a mixture, and possible recycling, of material derived
 194 from various sources.

195

196 **4. Results and Interpretation**

197 **4.1 Late Oligocene to Early/Middle Miocene deposits (28-20 Ma)**

198 In the area surrounding the Lluta and Azapa valleys (Fig. 2), Late Oligocene to
 199 Early Miocene deposits are preserved in the >300-m thick Azapa Formation (Fig.
 200 4). It comprises an alternation of 3-5 m-thick matrix-supported breccias and 2 m-
 201 thick massive-bedded (i.e., lacking distinct sedimentary flow structures) and
 202 parallel-laminated sandstones that are moderately sorted. Towards the west and
 203 the south, the Azapa deposits onlap and partially overlap the basement (Fig. 4A).
 204 This unit finally feathers out south of Camarones. The transport and deposition of

205 this material most likely occurred through debris flows and flash flood processes,
206 based on: (i) the matrix-supported fabric encountered in the breccia deposits,
207 which is characteristic for sediment transfer by cohesive and high-concentrated
208 debris flow processes where yield strengths of a highly viscous matrix and
209 buoyancy forces allow the transfer of large clasts (McArdell et al., 2007), (ii)
210 parallel laminations encountered in sandstone beds that suggest the occurrence of
211 sediment transport by non-confined floods in the upper flow regime (Fielding,
212 2006), and (iii) the massive-bedded fabric in some of the sandstone beds. This
213 suggests that the transfer of material was accomplished through highly
214 concentrated flows, or granular flows, where sediment deposition rates were much
215 faster than the time scales required to winnow and sort the grains upon transport
216 (Pettijohn et al., 1987).

217 The overlying c. 400-m-thick, Early Miocene Oxaya Formation (Fig. 4) is a stack
218 of ignimbrites and volcano-clastics and covers nearly the entire western margin in
219 the Lluta and Azapa areas (García et al., 2004). Towards the south, the Oxaya unit
220 thins out to a few tens of meters, where it is laterally encroached and replaced by
221 the Ignimbrita Suca and Moquella, and particularly by the deposits of the Alto de
222 Pica Formation (Farías et al., 2005) (Fig. 4). This latter unit spans the entire time
223 interval between c. 26 and 11 Ma and is thickest in the Aroma and Alto de Pica
224 regions (Pinto et al., 2004; Farías et al., 2005), where it comprises a >700 m-thick
225 alternation of sandstones, matrix supported breccias, and ignimbrites (Farías et al.,
226 2005; Jordan et al., 2010). The available sedimentological information inhibits a
227 clear assignment of the depositional processes. However, the frequent occurrence
228 of matrix-supported breccia interbeds (Jordan et al., 2014) allows an interpretation
229 that is similar to the breccia units within the Azapa Formation where debris flow
230 processes were a major sediment transport mechanism during that time.

231

232 **4.2. Middle to Late Miocene (20-10 Ma)**

233 The best accessible and most continuous exposures of Middle to Late Miocene
234 deposits are found along a road cut at 19°25'S/69°36'W near the village of
235 Camiña (Fig. 5A), where the sediments are encountered along the 225-m-thick
236 Francia section (Figs. 2, 3, and 4). Here, we present more information about the
237 sedimentary structures encountered of related deposits.

238

239 *Base of section, 13-12 Ma*

240 The base of the section, informally referred to as unit I in this paper, is c. 13 Ma
241 old (revised chronology of Fig. 3) and comprises a c. 50-m-thick suite of 3- to 5-
242 m-thick coarse-grained sandstone beds with m-thick conglomerate interbeds (Figs.
243 5B, 5C). The sandstones are massive-bedded (Sm-facies) and display both planar
244 and erosive basal contacts. Most of the sandstone beds have tabular geometries
245 and laterally extent over several tens to hundreds of meters. Pebble- to cobble
246 conglomerate interbeds display m-thick ribbon geometries with dm- to m-deep
247 erosional scours (Ch) and planar tops. The conglomerates display a matrix-
248 supported fabric (Gmm-facies; Fig. 5B). Some conglomerate beds contain
249 boulder-sized clasts. We interpret the sandstone beds of the Sm-facies to have
250 been deposited mainly by non-confined, high-concentrated flows, where large
251 matrix strengths (McArdell et al., 2007) and fast sediment transport rates resulted
252 in a poor sorting of the grains upon transport (Fielding, 2006). The result was the
253 formation of massive-bedded sandstones (Sm-facies) that lacks any evidence for
254 ripple marks, subaquatic dunes or any other sedimentary structures (Pettijohn et
255 al., 1987). The tabular geometries of these units suggest that the flows were non-
256 confined. In the same sense, the matrix-supported fabric encountered in the
257 conglomerate deposits (Gmm-facies; Fig. 5B) points towards a mechanism where
258 sediment transfer was controlled by the viscosity and yield strengths of the matrix
259 (McArdell et al., 2007), which resulted in large buoyancy forces and thus in the
260 transfer of outsized clasts. Such a mechanism is commonly found in cohesive and
261 highly concentrated debris flows (McArdell et al., 2007). These flows were most
262 likely confined in channels, as indicated by the m-deep scours and ribbon
263 geometries (e.g., Friend, 1983). In summary, unit I displays sedimentologic
264 properties that are characteristic for sediment transport by debris flows processes
265 either in channels, or as non-confined high-concentrated flows on mudflats. This
266 implies that runoff was most likely ephemeral.

267

268 *Middle part of section, c. 12 Ma*

269 The following 125 meters, dated to c. 12 Ma (see revised chronology on Fig. 3)
270 and referred to as unit II in this paper, are characterized by a succession of dm- to
271 m-thick sandstone beds with cm-thick intercalations of volcanic ash (Figs. 5D-
272 5G). A >20-m-thick welded ignimbrite with rip up clasts and boulders is found in

the middle of this unit. Most of the sandstone beds are parallel-laminated (Sp-facies) particularly in the upper part of unit II. However, some massive-bedded units (Sm-facies) and m-scale cross-bedded sandstones (Sc-facies) also occur (Fig. 5F). Some of the sandstone beds display cm-scale ripple cross-beds (Sr-facies) and aggrading ripple marks (Sa-facies; Fig. 5G). Massive-bedded sandstones display convolute bedding in a few locations (Scon-facies; Fig. 5E). Wavy laminations where laminae are arranged as sets of convex-upward oriented units occur at dm- to m-scales (Sw-facies; Fig. 5G) upon shallow erosive contacts. Steep erosional scours (Ch), where present, are up to 1-2 m deep and overlain by gravelly sandstones and conglomerates with a massive-bedded fabric and m-scale cross-beds (Gm- and Gc-facies) (Fig. 5F). The cross-beds are indicative for subaquatic megaripples, or dunes, that were formed by unidirectional flows in the lower flow regime within channels (Pettijohn et al., 1987). The ensemble of these observations is interpreted as channel fills where sediment transport occurred by confined flows. In contrast, the wavy laminations (Sw) and aggrading ripple marks (Sa) that are mainly encountered within the middle part of unit II suggest that these sandstone units represent antidunes where sediment transport was accomplished within the upper flow regime (Fielding, 2006). In these circumstances, standing waves resulted in the vertical aggradation of laminae and the formation of nearly symmetrical ripple marks, or dunes with shallow-dipping foresets (Sa- and Sw-facies). Likewise, the parallel laminations (Sp-facies) found in cm- to dm-thick sandstone beds also point towards the occurrence of upper flow regime conditions upon sediment transport. Finally, the sediment transport rates and deposition of massive-bedded sandstones beds (Sm-facies) occurred fast enough to inhibit the sorting of the material. Support for the inferred rapid deposition is also provided by the convolute bedding (Scon-facies) encountered in some of these sandstone beds, because these structures point to the occurrence of escape of in-excess water that was trapped in the pore space after episodic deposition (Fielding, 2006). In either case, the ensemble of sedimentary properties encountered within unit II imply that most of the material was transported and deposited by low-concentrated flows within a relatively short time of possibly a few minutes' duration (Fielding, 2006), which is characteristic for a highly episodic and thus flashy transport regime. Similar to unit I, this also implies that runoff was most likely ephemeral.

307

308 *Top of Francia section, 12-11 Ma*

309 A remarkable change is recorded at c. 12-11 Ma (Fig. 3; 5H-5K) with the first
310 appearance of well-sorted conglomerate fabrics in the uppermost 50 meters of the
311 section. This unit III is an amalgamation of clast-supported conglomerate beds
312 with thicknesses between several decimeters to >1 meter. The conglomerate clasts
313 are imbricated in places (Gi-facies). Crossbeds (Gc-facies) with dm-deep scours
314 and dm- to m-high forests with shallow dips also occur. Most of the conglomerate
315 beds, however, reveal a massive-bedded, clast-supported fabric (Gm-facies). Both
316 Gc- and Gm-facies follow upon erosional scours 1-2 m deep (Ch). Interbedded
317 parallel laminated sandstone beds (Sp-facies) with ripple marks (Sr-facies) are
318 encountered at the base of unit III. The ensemble of these features suggests
319 deposition by non-confined braided streams where units with a massive-bedded
320 fabric (Gm-facies) represent longitudinal bars that bordered a network of braided
321 and shallow channels (Ch). Alternatively, the cross-bedded conglomerates (Gc-
322 facies) most likely represented subaquatic transverse bars at the confluence
323 between braided tributary branches (Lunt et al., 2004). Sediment transport and
324 conglomerate deposition was accomplished during high stage flows, but in the
325 lower flow regime, where conglomerates reveal cross-beds and a massive-bedded
326 fabric. Imbricated clasts (Gi) and parallel laminated sandstones, however, suggest
327 that upper flow regime conditions also occurred. Most important, the inferred
328 occurrence of braided channels suggests that sediment transport and deposition
329 was accomplished by streams where runoff was most likely perennial (Rust,
330 1978).

331

332 *Facies relationships north and south of Francia section*

333 North of 19°-20°S (i.e., north of Francia section) units I and II encountered at
334 Francia onlap the top of the Oxaya Formation, while the upper member overlap
335 the Oxaya unit and can be mapped along the entire Western Andean margin of
336 Northern Chile (Figs. 2, 4). This change in the sedimentation characteristics
337 suggests a shift in the depositional style where episodic, and thus ephemeral high-
338 concentrated runoff changed to a mechanism where sediment transport was most
339 likely accomplished by perennial streams as is the case today at the same latitudes
340 (e.g., Lluta River, Ministerio de Obras Publicas, Chile). However, south of 19°-

341 20°S (i.e. south of Francia section), units I-III are laterally replaced by the breccia
342 deposits of the Alto de Pica unit (Jordan et al., 2014), suggesting that an episodic
343 and most likely ephemeral and flashy runoffs have been maintained (Fig. 4).

344

345 **4.3. Late Miocene to Pliocene (c. 10-5 Ma) and modern**

346 The sedimentary sequences show that perennial rivers persisted after 12-11 Ma
347 north of 19°-20°S, while farther south runoff conditions stayed ephemeral relative
348 to the north, with only short-term periods characterized by perennial runoff
349 conditions. These along-strike changes in depositional styles and related
350 implications for palaeoclimate reconstructions are elaborated in the next sections.

351

352 *Sedimentary characterization of deposits north of 19°-20°S*

353 At the foothills of the Western Cordillera at 18°S latitude, Late Miocene to
354 Pliocene deposits are preserved in the Huyalas Formation (Hy on Fig. 2). These
355 deposits have been locally derived from the adjacent Western Cordillera (García
356 and Hérail, 2005) and thus formed the Piedmont unit of this mountain range. Age
357 assignments for the Huyalas unit are mainly based on K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$
358 chronologies for biotite and amphibole minerals in interbedded tuff deposits.
359 Published ages reveal a large spread from 16.8±1.5 Ma (García et al., 1966) to
360 10.55±0.05 Ma (Wörner et al., 2000; Figs. 4A, 4B). However, García and Hérail
361 (2005) noted that their previously reported ages of 16.8±1.5 Ma (García et al.,
362 1966) and 14.8±0.9 Ma (García and Hérail, 2005) for two independent sites are
363 likely to represent overestimates due to low K contents in biotites (<6%) and
364 atmospheric Ar contributions of >75%. They proposed that age assignments
365 between 11.4±0.5 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$ age measured on amphiboles; García and Hérail;
366 2005) and 8.8±1.0 Ma (K-Ar age measured on biotites; García and Hérail; 2005)
367 are likely to be correct.

368 The Huyalas Formation comprises a >100-m-thick suite of conglomerates,
369 sandstones and mudstones. It starts with a ca. 50-m-thick alternation of coarse-
370 grained sandstone beds and fine-grained, clast-supported conglomerates with
371 erosive bases up to 50 cm deep (Kohler, 1999). Individual conglomerate interbeds
372 are matrix supported. This fabric suggests that the material was deposited on a
373 braid plain where deposition occurred by a combination of fluvial and debris flow
374 processes (see above for justification). The overlying c. 20-m-thick succession is

375 an alternation of diatomites, siltstones and fine-grained sandstone beds several
376 centimeters thick with root casts, suggesting the presence of a shallow lake
377 (Kohler, 1999). These deposits are overlain by m-thick matrix-supported breccia
378 beds that were deposited by one or multiple debris flows.

379 In the Lluta valley, situated on the western margin at the same latitude, clast
380 supported and imbricated fluvial conglomerates with sources in the Western
381 Cordillera mainly occur as Late Miocene to Pliocene fluvial strath terrace
382 conglomerates perched on valley walls, and Quaternary cut-and-fill sequences on
383 the valley bottom. This is particularly evident at 18°23'S/69°55'W (Fig. 4D),
384 where fluvial conglomerates with sources in the Western Cordillera occur as
385 terrace deposits situated ca. 450 m above the current valley floor, and ca. 200 m
386 beneath the 11 Ma-old top of the El Diablo formation (Fig. 4D, upper photo).
387 Mapping reveals that this terrace is laterally encroached by a rhyolite (Fig. 4D),
388 that either corresponds to the 2.7-Ma-old Lauca ignimbrite (Wörner et al., 2000),
389 or alternatively to a similar unit c. 4 km farther east dated to approximately 4.1 Ma
390 (García et al., 2004). In either case, these ages imply a Pliocene age for the terrace
391 deposits as these sediments are younger than the top of the El Diablo
392 conglomerates (11 Ma), but older than the rhyolite. In the same valley, massive-
393 bedded and clast-supported conglomerates are perched on hillslopes. Imbricated
394 clasts are found in a few locations. These conglomerates either underlay the 4.1
395 Ma or 2.7 Ma-old ignimbrite (García et al., 2004) (Fig. 4D, lower photo) or are
396 embedded in it. The clast-supported and partially imbricated fabric suggests that
397 runoff of the Lluta River was most likely perennial at that time (Rust, 1978), as is
398 also observed today for streams north of 19°-20°S (e.g., Ministerio de Obras
399 Publicas, Chile, Nester et al., 2007; Houston, 2006).

400 Also at the same latitude, but on the eastern margin of the Western Cordillera,
401 deposition of the c. 120-m-thick suite within the Lauca basin (Fig. 2) spanned a
402 time interval between c. 8 Ma and 1 Ma. Gaupp et al. (1999) categorized the
403 sediments into four stages of basin evolution, which comprises: (i) a c. 8-6.5 Ma-
404 old and >25 m-thick succession of alluvial and fluvial red sandstone beds and
405 conglomerates with interbedded lacustrine mudstones (stage I), (ii) a c. 40-m-thick
406 and 6.5-4 Ma-old suite of predominantly mudstones with gypsum and sandstone
407 interbeds (stage II environment with ephemeral lakes and mudflats), (iii) a <35-m-
408 thick suite of fluvial braidplain conglomerates representing stage III, and finally

409 (iv) a <35-m-thick suite of Pleistocene fluvio-glacial sandstone-conglomerate
410 alternations that were grouped into stage IV (see Gaupp et al., 1999, for
411 sedimentological details). The Lauca basin deposits are consistent with the records
412 at Huyalas in the sense that they imply the predominant occurrence of fluvial
413 sediment transport, which points to a relatively wet palaeoclimate with varying
414 precipitation and temperature conditions (see Gaupp et al., 1999, for further
415 details).

416

417 *Sedimentary characterization of deposits at 19.5°S latitudes*

418 At 19.5°S, the streams that deposited the unit I at Francia ended in a lake
419 (lacustrine diatomites, Fig. 4C) that was situated within the Central Depression
420 east of the Coastal Cordillera (Kirk-Lawlor et al., 2013) (see ‘lake deposits’ on
421 Fig. 2 between 19.5°S and 20°S). New mapping reported by Kirk-Lawlor et al.
422 (2013) has shown that this palaeolake covered a large area of >200 km² and
423 extended c. 40 km along the eastern border of the Coastal Cordillera.
424 Accumulation of these sediments occurred until at least 6.4 Ma, or possibly 3.5
425 Ma. After this time, streams started to incise into the Coastal Cordillera (Kirk-
426 Lawlor et al., 2013), thereby lowering the base level, terminating the lacustrine
427 period, and shifting the lacustrine environment into a salar (Hoke et al., 2007).

428

429 *Sedimentary characterization of deposits south of 19°-20°S*

430 South of 19°-20°S, however, 13-11 Ma deposits are made up of breccias and
431 matrix-supported conglomerates with sources in the Western Cordillera (Farías et
432 al., 2005). Younger deposits occur as ‘Piedmont units’ perched on the hillslopes of
433 the Western Cordillera (Jordan et al., 2014). They comprise a suite of 11-5.5 Ma
434 sandstones of presumably aeolian origin (Blanco and Tomlinson, 2013) and
435 matrix-supported breccias (Jordan et al., 2014), suggesting the occurrence of
436 concentrated flows and debris flows. 5 Ma to modern deposits comprise
437 subangular to rounded breccias and conglomerates (also sourced from the Western
438 Cordillera) and lake deposits, followed by a suite of conglomerates, sandstones of
439 presumably aeolian origin, mudstones, and salt-pan gypsum deposits thereafter
440 (Jordan et al., 2014, their Fig. 8 and Table 3). The available information about
441 clast textures suggests that deposition most likely occurred by mixed fluvial/debris
442 flow processes, but where sediment transport by debris flows predominated. Also

443 south of 19°-20°S, sediment transport in streams with sources in the Western
444 Cordillera tends to be mainly accomplished by debris flows in an ephemeral
445 environment (Nester et al., 2007, see their SI Table 2).

446

447 **5. Discussion**

448 **5.1. Changes in sedimentological structures imply shifts in runoff volumes**

449 Changes in the structures of clastic sediments, and particularly of conglomerates
450 and breccias, provide first order constraints on the physical processes and
451 hydrological conditions upon material transport and deposition. In this context, a
452 shift from debris flow to fluvial deposition either occurs in response to landscape
453 flattening, or to an increase in the ratio between water runoff volumes and
454 sediment discharge (Rickenmann and Koschni, 2010). In the case of the western
455 Andean margin, ongoing tectonic deformation and landscape steepening over the
456 entire Western Cordillera (Farías et al., 2008; Jordan et al., 2010) implies that
457 channel gradients most likely increased, or at least did not flatten. We are thus left
458 with a scenario where a substantial increase in water runoff volumes relative to
459 sediment flux (Heller and Paola, 1992; Rickenmann and Koschni, 2010) has to be
460 invoked to explain the stratigraphic change from debris flow to fluvial deposition
461 at 12-11 Ma. This could have been accomplished through a change to a wetter
462 climate with lower precipitation variability in the Western Cordillera where the
463 sources of the sediments were (Hartley and Evenstar, 2010). We base this
464 interpretation on modern examples in the USA where discharge patterns of
465 streams with sources under wetter climate conditions are less variable (Turcotte
466 and Green, 1993). Alternatively, lower discharge variabilities could also have been
467 accomplished by the growth of glaciers in response to global cooling and/or higher
468 rainfall rates. Glaciers are capable of buffering seasonal variations in precipitation,
469 which in turn could explain the observed changes in the stratigraphic records at
470 Francia. Finally, larger runoff values could also have been accomplished through
471 connection of the groundwater supply from the plateau to the western Andean
472 streams (Magaritz et al., 1989). In either case, the inferred change in sediment
473 transport invokes a substantial increase in water runoff volumes at the Francia
474 section, which occurred at c. 12-11 Ma.

475 After 11 Ma, differences in depositional styles encountered north and south of
476 19°-20°S suggest that relatively large runoff volumes have consistently been

477 maintained in the north (Fig. 6A), while conditions became most likely drier
478 farther south (Jordan et al., 2014). However, in the Altos de Pica region, the
479 occurrence of conglomerate interbeds within the 'Piedmont units' could also
480 indicate that climate turned wetter south of 19°-20°S, at least for short time
481 periods (Jordan et al., 2014). In the same sense, at Huyalas (Hy on Fig. 6A)
482 situated north of 19°-20°S, the occurrence of debris flow units could be indicative
483 of a shift towards drier conditions with episodic runoff. Nevertheless, the
484 ensemble of lithofacies between 18°-21°S latitudes does suggest that after 12-11
485 Ma, relative runoff volumes were larger north of 19°-20°S than farther south (Fig.
486 6A).

487

488 **5.2 Possible controls on shifts in inferred runoff volumes**

489 The observed changes in channel runoff could have geomorphic, climatic, or
490 tectonic origins. Headward retreat of streams and enlargement of drainage basins
491 through stream cannibalization would correspond to a surface control on the
492 inferred stratigraphic changes, because larger basins tend to be more fluvial, while
493 debris flows preferentially occur in smaller and steeper catchments (Rickenmann
494 and Koschni, 2010). While we cannot fully exclude such a scenario, we consider it
495 unlikely that such a mechanism alone is capable of explaining the observed N-S
496 trending stratigraphic changes mainly because they occurred over a broad spatial
497 scale and at nearly the same time. Indeed, if these changes were the response of
498 intrinsic enlargements of individual basins through headward retreat, then related
499 stratigraphic responses would have been strongly time transgressive, which cannot
500 be confirmed with the available dataset. We do acknowledge, however, that basin
501 enlargements might have amplified effects related to a possible climate driver
502 through a positive feedback, where large runoff yields larger basins through more
503 efficient erosion. In the same sense, we do not consider a shift towards a cooler
504 global climate in the middle Miocene (Zachos et al., 2001) as a potential control
505 on the inferred shift in precipitation patterns because the stratigraphic changes
506 occurred later. Instead, we use the results of regional palaeoclimate models to
507 suggest a tectonic control on the inferred shifts in available moisture on the
508 Altiplano and the Western Cordillera (Fig. 6A) where the sediment sources were
509 situated.

510 Model-predicted precipitation changes due to the rise of the Andes from two
511 independent regional palaeoclimate modeling studies conducted at ~55 km
512 resolution (Ehlers and Poulson, 2009; Insel et al. 2012) show that on the Altiplano
513 and Western Cordillera, precipitation increased farther north compared to the
514 south as the plateau elevations increased from 50-100% of the modern elevation.
515 The magnitude of predicted precipitation change varies between models (Fig. 6B),
516 but the general latitudinal trend of precipitation change from north to south is
517 robust between the different models. For example, as topography increases from
518 50% to modern elevations, both models predict wetter conditions north of 19°-
519 20°S. Farther south, slightly wetter conditions (Ehlers and Poulsen 2009) to no
520 precipitation change (Insel et al. 2012) is suggested compared to modern
521 conditions (Fig. 6B). Most important, both model and stratigraphic evidence do
522 support a scenario where climate turned wetter in the north relative to the south.
523 Alternatively, a $>4^\circ$ southward shift of the atmospheric circulation pattern could
524 explain the increased average runoff north of 19.5°S. The latitudinal position of
525 the sub-tropical high-pressure belt is mainly driven by insolation (Houghton,
526 1984) and it is known to vary periodically at both seasonal (Garreaud et al., 2003)
527 and orbital time scales. It is possible that such a shift could have occurred through
528 a mechanism where the rise of the Altiplano and the related increase in insolation
529 would lead to a stronger Andean jet and thus to a southward shift in climate belts.
530 However, we discard a possible weakening of the Humboldt current as driving
531 force for the inferred climate change since this would also have resulted in a
532 wetter climate in the south, which is not consistent with our observations.
533 These N-S precipitation gradients have been maintained during Quaternary
534 climate oscillations when major precipitation was localized farther north, forming
535 large lakes on the Altiplano that extended as far south as the Salar de Uyuni (Fritz
536 et al., 2004). We thus propose that the change to the current precipitation pattern
537 was established at 12-11 Ma at the latest. The timing of the 12-11 Ma
538 stratigraphic changes is also broadly consistent (although with large uncertainties)
539 with available chronological information for the eastward growth of the Andean
540 foothills near Santa Cruz (Fig. 1). There, the rotation and formation of the
541 Bolivian orocline together with the protrusion of the Santa Cruz Bend was most
542 likely accomplished through sinistral transpressional faulting along the
543 Cochabamba fault zone (Kley, 1999). We note, however, that the eastward

extension of this fault is not clear and its possible role on oroclinal bending has been contested (Eichelberger et al., 2013). In either case, low temperature thermochronometric records were used to infer an age <20 Ma for the protrusion of the bend, associated by a phase of rapid cooling and deformation between 10-5 Ma (Eichelberger et al., 2013). As a consequence, the west-east promontory of high topography would have intercepted moisture transport, thereby deviating the moisture source farther to the east. We thus suggest that the ensemble of these tectonic processes offer a suitable mechanism to explain the inferred 12-11 Ma-old amplification of N-S precipitation gradients when on the Western Cordillera palaeoclimate turned wetter north of 19°-20°S, while conditions became hyperarid south of this latitude (Fig. 6A). The consequences are also observed today where modern precipitation totals in the North (Fig. 1) are large enough to maintain the Lago Titicaca, northern Bolivia, in freshwater conditions and to sustain a perennial runoff of streams with sources in the Western Cordillera (including rivers in the Lluta and Camarones valleys).

5.3. Implications for reconciling contrasts in palaeoclimate reconstructions, and conclusions

The inferred tectonic controls on precipitation patterns in the Andes of Northern Chile offers the possibility to reconcile possible contrasts in palaeoclimate reconstructions (Fig. 7). For the region south of 19°-20°S latitudes, Jordan et al. (2014) proposed a scenario (based on an analysis of sedimentary archives and palaeosols preserved in the Central Depression, Fig. 2) where precipitation rates decreased from originally c. 200 mm/yr prior to 11-12 Ma, to < 50 mm/yr thereafter. Support for this palaeoclimate change scenario is offered by Rech et al. (2006), who interpreted the change from 20 Ma-old calcic vertisols to >3 m-thick salic gypsisols at 10-8 Ma as a response to a decrease in precipitation rates from >200 mm/yr prior to 20 Ma, to <20 mm/yr thereafter.

For the western Andean margin north of 19°-20°S latitude, the results of our investigations, although largely incomplete regarding the time covered by the archives, yield comparable conclusions, but with opposite directions (Fig. 6A). In particular, if we consider a precipitation rate of 200 mm/yr for the transfer of debris flow breccias as reference (Jordan et al., 2014), then our inferred shifts from debris flows to fluvial processes at c. 12-11 Ma would imply an increase in

precipitation rates from the reference value of 200 mm/yr prior to 12 Ma (Jordan et al., 2014), to the current c. 300 mm/yr thereafter (Fig. 7). Following the results of paleocirculation models, the rise of Andes resulted in attraction of more moisture through strengthening of the Andean jet, thereby increasing moisture availability in the high Andes in the North while maintaining dry conditions in the south. In such a conceptual framework, oscillations of global climate would additionally modulate the precipitation pattern (Gaupp et al., 1999), but would not exert the major driving mechanism for increasing the contrasts in moisture availability north and south of 19°-20°S latitudes.

While our interpretation is non-unique, the stratigraphic data implies that the currently observed N-S trends in precipitation rates in Northern Chile (Fig. 1) was already established between 12 and 11 Ma at the latest (Fig. 6A, 7). As a second consequence, our interpretation implies a strong linkage between orographic precipitation and stratigraphy where the tectonic history of the Andes controls the distribution of available moisture and sediment transport on the western flank.

Figure caption

Figure 1: (A) Central Andes with pattern of annual precipitation. The inset map (lower left side) shows the course of the Andean jet. I, II and III in lower right inset indicate sections (swaths 1, 8 and 16 of Figure 1B) for which 10-km-wide precipitation swath-profiles are shown in inset figure on the lower right side. Precipitation data are the 2B31 TRMM data set (Mulligan, 2006). Satellite images © Google 2015. (B) Mean annual water budgets for 16 swaths across the Andes. See Figure 1A for location of swaths 1, 8 and 16. Water budgets were calculated as the products between mean annual precipitation rates (extracted from the 2B31 TRMM data set) and the area. No evapo-transpiration nor infiltration has been considered for these calculations. Swath 1 corresponds to section I on the lower right inset of Figure 1A, swath 8 to section II, and swath 16 to section III.

Figure 2: Overview map showing the western margin of the Andes of northern Chile. The Figure illustrates the tectono-geomorphic domains (including the Coastal Cordillera labeled as CC and the Western Cordillera from where the investigated sediments have been derived). The figure also

shows the spatial distribution of Late Miocene fluvial and mixed fluvial/debris flow conglomerates that are preserved in the El Diablo Formation (modified after von Rotz et al., 2005). Red dots are sites where ash layers and ignimbrites were dated (see Fig. 4 for reference to related publications). Satellite images © Google 2015

Figure 3: (A) Stratigraphic column of the Francia section, (B) correlation of the Francia magnetostratigraphy (MPS, von Rotz et al., 2005) with the magnetopolarity time scale (MPTS), and (C) revised chronological correlation of the section (this paper). Please see Figs. 1A and 2 for location of section.

Figure 4: (A) Stratigraphic relationships and (B) chronological diagram for deposits in northern Chile since the Miocene. Stratigraphic data were taken from: a) Kohler (1999), b) von Rotz et al. (2005), c) Farias et al. (2005), d) Jordan et al. (2014), e) García and Hérail (2005) and f) Kirk-Lawlor et al. (2013). Radiometric ages are compiled from: 1: Wörner et al. (2000), 2: Walfort et al. (1995), 3: Muñoz and Sepúlveda (1992), 4: Naranjo and Paskoff (1985), 5: Parraguez (1998), 6: Farías et al. (2005), and 7: García and Hérail (2005). (C) Schematic section across the Western Andean margin. (D) Photos illustrating strath terrace deposits in the Lluta area. The upper photo shows the fluvial terrace that formed in the Lluta valley c. 450 meters above the modern valley floor and c. 400 m beneath the top of the El Diablo Formation (18°23'S, 69°55'W). Mapping revealed that a 4.1 or 2.7 Ma-old rhyolite onlaps the base of the terrace, which implies a Pliocene age for the conglomerates on top of the terrace. The lower photo shows a strath terrace that was deposited by the Lluta River. The photo is taken in the Lluta valley at 18°23'S, 69°56'W. See Fig. 2 for location of sections (A) and (C).

Figure 5: (A) Overview with spatial arrangements of stratigraphic units I-III encountered in the Francia section (Fig. 3) along the road to Camiña. (B) and (C): Sedimentological details of unit I; (D)-(G): sedimentological fabric of unit II, and (H)-(K) sedimentological properties of unit III. See Figs. 1A, 2 and 4 for location of section, and Figs. 3 and 5A for location of photos. Satellite image © Google 2015.

Figure 6: (A) Wheeler (Chronologic) diagram of the deposits situated along the western Andean escarpment between 18°S and 21°S along the same transect as Figs. 4A and 4B. The data was compiled from: García et al. (2004) and García and Hérail (2005) for the Andes at 18°S, Nester et al. (2007) for modern streams, Kirk-Lawlor et al. (2013) for lakes at 19.5°S, and Farías et al. (2005) and Jordan et al. (2014) for sections at 20°-21°S. The hatched space shows time interval when the North was wet compared to the South. Prior to 12 Ma dry conditions prevailed along the entire segment. This is inferred from the occurrence of debris flow deposits along the entire Andean margin. (B) N-S pattern of model-based increase in mean precipitation rates averaged over the Altiplano and the Western Cordillera. This was calculated from 50% of the current plateau elevation to its modern height from two different regional atmospheric general circulation models (crosses: Ehlers and Poulsen, 2009; grey area: Insel et al. 2012). The spatial resolution of the simulations was ~55 km.

Figure 7: (A) Chronology of paleoclimate development inferred from stable isotope records (Zachos et al., 2001). (B) Estimations of precipitation rates for the Andes north of 19°-20°S based on sedimentary archives (this work) and inferred from sections presented by Gaupp et al. (1999; evolutionary stages I-IV), and reconstructed palaeoclimate shifts south of 19°-20°S proposed by Jordan et al. (2014). See text for further details. AdP: Altos de Pica Formation; A: Azapa Formation; Ox: Oxaya Formation; I-II of ED and III of ED: units I-II of El Diablo Formation and unit III of El Diablo Formation (Fig. 4); St: strath terraces in Lluta valley; I-IV: evolutionary stages of Lauca basin (See Fig. 2 for location of basin).

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680

681 **References**

682 Allmendinger, R.W., González, G., Yu, J., Hoke, G., Isacks, B., 2005. Trench-
683 parallel shortening in the Northern Chilean Forearc: tectonic and climatic
684 implications. *Geol. Soc. Am. Bull.* 117, 9-104.

685 Armijo, R., Lacassin, R., Coudrier-Curveur, A., Carrizo, D., 2015. Coupled
686 tectonic evolution of Andean orogeny and global climate. *Earth-Sci. Rev.* 143,
687 1-35.

688 Blanco, P.N., and Tomlinson, A.J., 2013, Carta Guatacondo, Región de Tarapacá:
689 Santiago, Chile, Servicio Nacional de Geología y Minería, Subdirección
690 Nacional de Geología, Carta Geológica de Chile, Serie Geología Basica 15, 109
691 p., 1 map, scale 1:100,000.

692 Campetella, C.M., Vera, C.S., 2002. The influence of the Andes Mountains on the
693 South American low-level flow. *Geophys. Res. Lett.* 29, 1826.
694 doi:10.1029/2002GL015451

695 Ehlers, T.A., Poulsen, C.J., 2009. Influence of Andean uplift on climate and
696 paleoaltimetry estimates. *Earth Planet. Sci. Lett.* 281, 238-248.

697 Eichelberger, N., McQuarrie, N, Ehlers, T.A., Enkelmann, E., Barnes, J.B., Lease,
698 R.O., 2013. New constraints on the chronology, magnitude, and distribution of
699 deformation within the central Andean orocline. *Tectonics* 32, 1432-1453.

700 Farías, M., Charrier, F., Comte, D., Martinod, J., Hérail, G., 2005. Late Cenozoic
701 deformation and uplift of the western flank of the Altiplano: evidence from the
702 depositional, tectonic, and geomorphic evolution and shallow seismic activity
703 (northern Chile at 19°30'S). *Tectonics* v. 24, TC4001.

704 Farías, M., Charrier, F., Carretier, S., Martinod, J., Fock, A., Campbell, D.,
705 Cáceres, J., Comte, D., 2008. Late Miocene high and rapid surface uplift and its
706 erosional response in the Andes of central Chile (33°-35°S). *Tectonics* 27,
707 TC1005.

708 Fielding, C.R., 2006. Upper flow regime sheets, lenses and scour fills: Extending
709 the range of architectural elements for fluvial sediment bodies. *Sediment. Geol.*
710 190, 227-240.

711 Friend, P.F., 1983. Towards the field classification of alluvial architecture or
712 sequence. In: Collinson, S.D., Lewin, J. (Eds.), *Modern and ancient fluvial*
713 *systems*. Int. Ass. Sed. Spec. Publ. 6, 345-354.

714 Fritz, S.C., Baker, P.A., Lowenstein, T.K., Seltzer, Rigsby, C.A., 2004.
715 Hydrologic variation during the last 170,000 years in the southern hemisphere
716 tropics of South America. *Quat. Res.* 61, 95-104.

717 García, M., Hérail, G., Charrier, R., 1996. The Cenozoic forearc evolution in
718 Northern Chile: the border of the Altiplano of Belén (Chile). Third International
719 Symposium on Andean Geodynamics: Institut Français de Recherche pour le
720 Développement en Coopération (ORSTOM), Saint Malo, France, pp. 359–362.

721 García, M., Gardeweg, M., Clavero, J., Hérail, G., 2004. Hoja Arica, Región de
722 Tarapacá. Servicio Nacional de Geología y Minería: Carta Geológica de
723 Chile, Serie Geología Básica, 84, mapa escala 1:250'000, Santiago.

724 García, M., Hérail, G., 2005. Fault-related folding, drainage network evolution and
725 valley incision during the Neogene in the Andean Precordillera of Northern
726 Chile. *Geomorphology* 65, 279-300.

727 Garreaud, R., Vuille, M., Clement, A.C., 2003. The climate of the Altiplano:
728 observed current conditions and mechanisms of past change. *Palaeogeogr.*
729 *Palaeoclimatol. Palaeoecol.* 194, 5-22.

730 Gaupp, R., Kött, A., Wörner, G., 1999. Palaeoclimatic implications of Mio-
731 Pliocene sedimentation in the high-altitude intra-arc Lauca Basin of northern
732 Chile. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 194, 5-22.

733 Hartley, A.J., 2003. Andean uplift and climate change. *J. Geol. Soc. London* 160,
734 7–10.

735 Hartley, A.J., Evenstar, L., 2010. Cenozoic stratigraphic development in the north
736 Chilean forearc: Implications for basin development and uplift history of the
737 Central Andean margin. *Tectonophysics* 495, 67-77.

738 Heller, P.L., Paola, C., 1992. The large-scale dynamics of grain-size variation in
739 alluvial basins, 2: Application to syntectonic conglomerates. *Basin Res.* 4, 73-
740 90.

741 Hoke, G.D., Isacks, B.L., Jordan, T.E., Tomlinson, A.J., Blanco Pavez, N.
742 Razmezani, J., 2007. Geomorphic evidence for post-10 Ma uplift of the western
743 flank of the central Andes 18°30'-22°S. *Tectonics* 26, TC5021.

744 Houston, J., Hartley, A.J., 2003. The central Andean west-slope rainshadow and
 745 its potential contribution to the origin of hyper-aridity in the Atacama Desert.
 746 *Int. J. Climatol.* 23, 1453-1464.

747 Houston, J., 2006. Variability of precipitation in the Atacama Desert: It's causes
 748 and hydrological impact. *Int. J. Climatol.* 26, 2181-2198.

749 Insel, N., Poulsen, C.J., Ehlers, T.A., 2010. Influence of the Andes Mountains on
 750 South American moisture transport, convection, and precipitation. *Clim. Dyn.*
 751 35, 1477-1492. doi:10.1007/s00382-009-0637-1.

752 Jordan, T.E., Nester, P.L., Blanco, N., Hoke, G.D., Dávila, F., Tomlinson, A.J.,
 753 2010. Uplift of the Altiplano-Puna plateau: A view from the west. *Tectonics* 29,
 754 TC5007.

755 Jordan, T.E., Kirk-Lawlor, N.E., Blanco, N., Rech, J.A., Cosentino, N.J., 2014.
 756 Landscape modification in response to repeated onset of hyperarid paleoclimate
 757 states since 14 Ma, Atacama Desert, Chile. *Geol. Soc. Amer. Bull.* 126, 1016-
 758 1046. doi: 10.1130/B30978.1

759 Kirk-Lawlor, N.E., Jordan, T.E., Rech, J.A., Lehmann, S.B., 2013. Late Miocene
 760 to Early Pliocene paleo- hydrology and landscape evolution of Northern Chile,
 761 19° to 20° S. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 387, 76–90,
 762 doi:10.1016/j.palaeo.2013.07.011.

763 Kley, J., Müller, J., Tawackoli, S., Jacobshagen, V., Manutsoglu, E., 1997. Pre-
 764 Andean and Andean-age deformation in the eastern Cordillera of southern
 765 Bolivia. *J. South Amer. Earth Sci.* 10, 1-19.

766 Kohler, I., 1999. Syntektonische kontinentale Sedimentation auf der
 767 Westabdachung der Anden Nordchiles (18° bis 19°S). *Profil* 17, 1-165,
 768 Universität Stuttgart.

769 Lunt, L.A., Bridge, J.A., Tye, B., 2004. A quantitative, three-dimensional
 770 depositional model of gravelly braided rivers. *Sedimentology* 51, 377-414.

771 Magaritz, M., Aravena, R., Peña, H., Suzuki, O., Grilli, A., 1989. Water chemistry
 772 and isotopic study of streams and springs in northern Chile. *J. Hydrol.* 108,
 773 323–341.

774 Mcardell, B.W., Bartelt, P., Kowalski, J., 2007. Field observations of basal forces
 775 and fluid pore pressure in a debris flow. *Geophys. Res. Lett.* 34, L07406,
 776 doi:10.1029/2006GL029183.

777 McQuarrie, N., 2002. Initial plate geometry, shortening variations, and evolution
778 of the Bolivian orocline. *Geology*, 30(10), pp.867-870.

779 Mulligan, M., 2006. Global gridded 1 km TRMM rainfall climatology and
780 derivatives. Version 1.0 Database: <http://www.ambiotek.com/1kmrainfall>.

781 Muñoz, N., Sepúlveda, P., 1992. Estructuras compresivas con vergencia al oeste
782 en el borde oriental de la Depresión Central, Norte de Chile (19°15'S). *Rev.*
783 *Geol. Chile* 19, 241-247.

784 Naranjo, J., Paskoff, R., 1985. Evolución cenozoica del piedemonte andino en la
785 Pampa del Tamarugal, norte de Chile (18°-21°S). *Congreso Geol. Chileno* 4,
786 149-165, Antofagasta.

787 Nester, P.L., Gayó, E., Latorre, C., Jordan, T.E., Blanco, N., 2007. Perennial
788 stream discharge in the hyperarid Atacama Desert of northern Chile during the
789 latest Pleistocene. *Proc. Natl. Acad. Sci.* 104, 19,724–19,729,
790 doi:10.1073/pnas.0705373104.

791 Parraguez, G., 1998. Sedimentología y geomorfología producto de la tectónica
792 cenozoica, en la Depresión Central, I Región de Tarapacá, Chile: Memoira de
793 Título de tesis, Unversidad de Chile, 108 p.

794 Pettijohn, P.J., Potter, P.E., Siever, R., 1987. *Sands and Sandstones*. 2nd Edn.,
795 Springer-Verlag, New York, 553 p.

796 Pinto, L., Hérail, G., Charrier, R., 2004. Sedimentación sintectónica a las
797 estructuras neógenas en la Precordillera de la zona de Moquella, Tarapacá
798 (19°15'S, norte de Chile). *Rev. Geol. Chile* 31, 19–44.

799 Rech, J.A., Currie, B.S., Michalski, G., Cowan, A.M., 2006. Neogene climate
800 change and uplift in the Atacama Desert, Chile. *Geology* 34, 761-764.

801 Rickenbach, D., Koschni, A., 2010. Sediment loads due to fluvial and debris flows
802 during the 2005 flood events in Switzerland. *Hydrol. Proc.* 24, 993-1007.

803 Rust, B.R., 1978. Depositional models for braided alluvium, in Miall, A.D., Ed.,
804 *Fluvial Sedimentology*. *Mem. Canad. Assoc. Petrol. Geol.* 5, 221-245.

805 Strecker, M.R., Alonso, R.N., Bookhagen, B., Carrapa, B., Hilley, G.E., Sobel,
806 E.R., Trauth, M.H., 2007. Tectonics and climate of the southern Central Andes.
807 *Ann. Rev. Earth Planet. Sci.* 35, 747-787.

808 Turcotte, D.L., Greene, L., 1993. A scale-invariant approach to flood- frequency
809 analysis. *Stoch. Hydro. Hydraul.* 7, 33–40.

810 Von Rotz, R., Schlunegger, F., Heller, F., 2005. Assessing the age of relief growth
811 in the Andes of northern Chile: Magneto-polarity chronologies from Neogene
812 continental sections. *Terra Nova* 17, 462–471.

813 Walfort, B., Hammerschmidt, K., Wörner, G., 1995. New Ar/Ar ages from
814 Tertiary volcanics in the northern Chilean Andes (18°S): Implication for
815 tectonic and magmatic evolution. *Terra Nova, Abstracts*, v. 7, p. 354.

816 Wörner, G., Hammerschmidt, K., Henjes-Kunst, F., Lezaun, J., Wilke, H., 2000.
817 Geochronology (40Ar/39Ar, K-Ar and He-exposure ages) of Cenozoic
818 magmatic rocks from Northern Chile (18-22°S): implications for magmatism
819 and tectonic evolution of the central Andes. *Rev. Geol. Chile* 27, p. 205-240.

820 Zachos, J., Pagani, M., Sloan, L., Thomas, E., Billups, K., 2001. Trends, rhythms,
821 and aberrations in global climate 65 Ma to present. *Science* 292, 686-693.

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825

Figure 1

A

B

Figure 1

Figure 2

B: Basement; CC: Coastal Cordillera, Az: Azapa Formation, Ox: Oxaya Formation; MoqD: upper member of Moquelle Formation; IM: Ignimbrita Moquella, IS: Ignimbrita Suca; I-II of ED: Units I and II of El Diablo Formation; III of ED: Unit III of El Diablo Formation; AdP: Alto de Pica Formation; Hy: Huyalas Formation; LdT: Lava de Tana; L: Lauca Ignimbrite.

Figure 2

Chronology of Francia section

Figure 3

Figure 4

Figure 4

Figure 5

Figure 5

Figure 5 continued

Abbreviations:

- Ch: Channels (m-deep erosional scours)
- Gmm: Matrix-supported conglomerates
- Gm: Massive-bedded and clast-supported conglomerates
- Gc: Cross-bedded conglomerates
- Gi: Imbricated conglomerates, clast supported
- Sm: Massive-bedded sandstones
- Sp: Parallel-laminated sandstones
- Sc: Cross-bedded sandstones
- Sr: Ripple cross-laminations in sandstones
- Sw: Wavy laminations in sandstones
- Sa: Sandstones beds, aggrading ripples, antidunes
- Scon: Sandstone beds, convolute bedding

Figure 5(continued)

Figure 6

A

B

Figure 6

Figure 7

Figure 7