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Holocene hydroclimate variability along the Southern Patagonian margin (Chile) reconstructed from Cueva Chica speleothems

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1 Introduction

Patagonia is ideally situated to reconstruct past migrations of the Southern westerly winds (SWWs) due to its southerly maritime location. It is the most southerly landmass and is almost surrounded by oceans: the Pacific Ocean to the west, the Atlantic Ocean to the east, the Southern Ocean to the South (Fig. 1). The dominating feature of the Southern mid-high latitudes is prevailing SWWs, which are responsible for driving the circulation of the Southern Ocean, the eastwards Antarctic Circumpolar Current (ACC) and the Peru-Chile Current (PCC) (Lamy et al., 2002, 2015). The SWWs are an important driver of Southern Ocean upwelling (Bakun et al., 2010) and their strength and latitudinal position are predominantly controlled by sea surface temperatures (SST) (Lamy et al., 2010). The SWWs have the strongest velocities between $\sim 50^\circ$ and 55° S. They seasonally migrate each year in response to the expansion and contraction of the Antarctic Sea ice (Aceituno et al., 1993) and the expansion of the Southern Polar Sea (Garreaud et al., 2013). Nevertheless, the latitudinal position and intensity of the SWWs changed in the past (Fletcher and Moreno, 2011), leading thus to different responses of the vegetation to past climate changes along the Chilean continental margin.

Several paleoenvironmental studies in the last decade showed that the SWWs impacted the climate variability of the Southern Patagonia during the last glacial termination and the Holocene periods (Moreno et al., 2012; 2018a, Zolitschka et al., 2019) as a consequence of deglaciation and long-term SST changes (Lamy et al., 2010, Shevenell et al., 2011, Ivanovic et al., 2018) in the South Pacific Ocean. This led to different environmental responses of ecosystems (Mansilla et al., 2016; 2018, McCulloch et al., 2001; 2019; 2020; 2021) to past climate change throughout the Chilean landmass, with records showing periods of warmer climate when the SWWs shifted poleward and cooler climate, when the SWWs shifted equatorward (Toggweiler et al., 2006). The westerlies' pattern and fluctuations even increased since the mid-Holocene along with higher SST gradients, leading to rapid climate cycles (RCC) recorded in some derived-pollen data from lacustrine records (Moreno et al., 2016, 2018b).

Whether long- or short-term changes in the SWWs were recorded in the Patagonian records, the sensitivity of such terrestrial records depends on gradients including the latitudinal position, distance to the coast as well as exposure to SWWs. In the complex geographical settings of Patagonia, a dissimilar distribution of rainfall is noticeable: orographic precipitation is produced as the SWWs reach the western coast and are driven upwards over the Andean Cordillera leading to a hyper-humid region in the west (~ 8000 mm/a) with the moisture decreasing eastwards, creating a rain shadow effect (~ 500 mm/a) (Schneider et al., 2003). This East-West precipitation gradient is reflected nowadays in the ecotones that range from Moorland and Evergreen Forest in the west to Steppic open field landscape in the east (Tuhkanen et al., 1990).

Among the various lacustrine (Moreno et al., 2018a, 2019; Zolitschka et al., 2019), peat (Mansilla et al., 2016, 2018; McCulloch et al., 2020, 2021) and soil (Alloway et al., 2017) records that helped reconstruct the vegetation history and their climatic feedback in Southern Patagonia, very few are based on the study of geochemical proxies directly linked to the hydroclimate variability of rainfall or temperature. The only two geochemical proxy-based records from carbonates (Solari et al., 2010; Schimpf et al., 2011) are discontinuous and span only the late-Holocene period. Speleothems (e.g., stalagmites, flowstone) are secondary chemical carbonates deposited in caves and represent the most suitable terrestrial archives for establishing high resolution proxy time-series in paleoclimate research (Genty et al., 2003; Cheng et al., 2012; Fairchild and Baker, 2012). Beyond qualitative trends provided by the isotopic signature and trace-

elements trapped in their laminae, speleothems can also provide quantitative estimates of paleoclimate data (e.g. temperature, rainfall amount) either through cave-specific proxy calibration (Tremaine et al., 2011) or through the isotopic analysis of fluid inclusions trapped within the carbonate speleothems at the time of their formation (van Breukelen et al., 2008; Matthews et al., 2021; Wassenburg et al., 2021). Moreover, the application of clumped isotopes on associated calcite (Eiler 2007; 2011) provides an additional control on the cave temperature and its variation at the time of speleothem formation (Affek et al., 2008; Matthews et al., 2021).

This study presents a new flowstone core S6-S6bis and a stalagmite S8 from The Cueva Chica located in the south-eastern part of the Cerro Benitez (S51°33'; W 72°33'), Ultima Esperanza region in Patagonia, Chile. A series of radiometric dating, petrographical and geochemical analyses (stable isotopes, chemical composition), combined with monitoring data of the cave was conducted to produce trace elements and isotope time-series. The Chica speleothems provide a high-resolution and well-dated terrestrial record spanning the hydroclimate history of the Southern Patagonia for the last 12 ka BP. We evaluate the timing and intensity of past climatic change throughout the Holocene and discuss both in conjunction with available marine and regional paleoenvironmental records in order to better constrain the regional paleoclimatic evolutions of SWWs. Lower resolved quantitative temperatures estimates are combined with higher resolved proxies to bolster our qualitative paleoclimatic reconstructions.

2 Environmental settings of Cerro Benitez and Chica cave

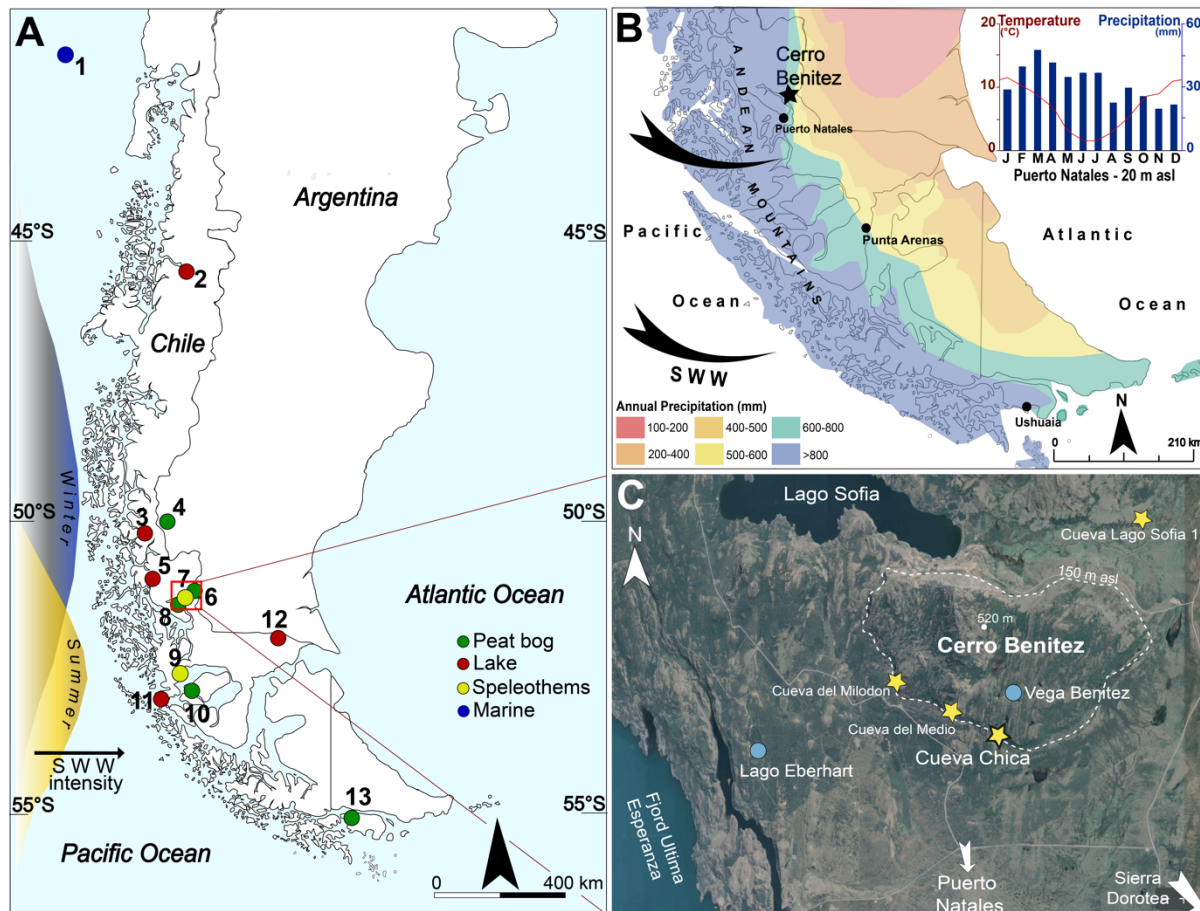
Cerro Benitez, a small hill of 520 m a.s.l., is located 20 km north of Puerto Natales, in the Eastern Andes, 10 km inland from the East coast of the Última Esperanza fjord (Fig. 1). The hill is comprised of conglomerates of the Cerro Toro formation (Romans et al., 2011), smoothed by successive glacial advances draining the South-Eastern part of the South Patagonian icefield. An East-West running glacial valley, today partially occupied by Lago (lake) Sofia, borders the Northern flanks of Cerro Benitez. Lago Sofia is the remnant of a large ice-dammed proglacial lake (proglacial Lago Puerto Consuelo) formed after the glacier retreated around 17.5 ka (Sagredo et al., 2011; Stern et al., 2011). The hill is surrounded by late Pleistocene erosional platforms and palaeoshorelines (Girault et al., 2022).

Cerro Benitez is influenced by the SWWs, which deliver precipitation from the Southeast Pacific (Fig. 1). The nearest meteorological station at Puerto Natales (51°44'S, 72°31'W, 2 m a.s.l.) receives 513 mm precipitation annually, which are well distributed through the whole year. Precipitation maxima occur in March, April, August, and November. The mean annual air temperature (MAAT) is between 6° and 7°C with an annual range of 9°–10°C (Dirección Meteorológica de Chile, www.meteochile.gob.cl/inicio.php). The limited moisture compared to the Western side of the Andes, results in a distinct forest-steppe ecotone covering the Eastern slope of the mountains, and particularly Cerro Benitez (Villa-Martínez and Moreno 2007; Moreno et al., 2009).

In the Southern flanks of Cerro Beníte, a few meters above the late Pleistocene 155 m asl lake terrace, a series of caves and rock shelters were partially sediment filled. This infilling occurred after several glacier retreats and advances from 30 ka to 18 ka BP (Girault et al., 2022). More than thirteen underground galleries comprise fossil remnants of megafauna in cave deposits (Villavicencio et al., 2016), which today comprise the oldest paleontological and archaeological material in the region (Prieto, 1991; Nami and Nakamura, 1995; Borrero and Franco, 1997; Martin et al., 2015; Martin and Borrero 2017). Two caves (Cave Lago Sofia 1 and el Medio Cave) feature two of the oldest archaeological records of Patagonia, dated between 13.6 and 10.6 ka cal. BP, while Chica Cave features the oldest paleontological record of the region, a *Lama gracilis* astragale dated to 18.500-17.930 cal. BP.

Chica Cave (S51°34'23.55"; W72°35'8.09") opens at 162 m a.s.l. and develops in complex mudstone/sandstone strata of the Cerro Toro formation. With its 73 m long relict conduits Chica cave is the most karstified site studied here. The cave is characterized by phreatic morphologies as well as unconsolidated and laminated sediments that are partially sealed with a thick calcite flowstone (Suppl., fig. 1). Seasonal drip and meteoric water flow on flowstone is still active in the Eastern gallery. As no stalagmites are preserved in the cave due to vandalism, a 40x6 (Lxl) cm core “S6” and a 5x5 (Lxl) cm core “S6bis” were retrieved from the flowstone at the end of the Western gallery (Suppl., fig.1). Additionally, a broken stalagmite (sample “S8”) has been collected on site. Temperature loggers placed both outside and inside the cave between December 2018 and December 2019, show an average temperature of $4.6 \pm 0.6^\circ\text{C}$ inside the

119 cave and 9.4°C outside in summer and 2°C in winter (Suppl., fig. 2). The catchment area of the cave
 120 comprises hilly slopes covered by scattered trees of *Nothofagus*. Some areas are covered by open grassland
 121 and peat vegetation. It is noteworthy that the catchment of the cave is downstream the Vega Benitez peat
 122 bog (McCulloch et al., 2021).
 123



124
 125 **Fig. 1.** Overview of southernmost Chile with the study site. **A)** Map of southernmost Chile with sites of discussed paleoclimate
 126 records: 1 – Marine cores GeoB3313-1 (Lamy et al., 2002) and ODP1233 (Lamy et al., 2015), 2 – lake Mallin Pollux
 127 (Markgraf et al., 2007), 3 - Lago Argentino (Kaplan et al., 2016), 4 - Cerro Frias peat (Tonello et al., 2009), 5 - Lago Cipreses
 128 (Moreno et al., 2018), 6 - Vega Benitez peat (McCulloch et al., 2021), 7 - **Chica cave** (this study), 8 - Lago Eberhart peat
 129 (Moreno et al., 2012), 9 – Marcelo Arevalo (MA) cave (Schimpf et al., 2011), 10 - GC peat bog (Killian et al., 2003; Lamy et
 130 al., 2010), 11 - lake Tamar (Lamy et al., 2010), 12 - lake Potrok Aike (Schäbitz et al., 2013), 13 - Pta Burselem & Isla Navarino
 131 peats (McCulloch et al., 2019; 2020). Latitudinal changes of average westerly wind intensities in summer and winter between
 132 1960 & 2000 AD are shown in yellow and blue (source: NCEP-NCAR). **B)** Annual precipitation over southernmost Chile
 133 (Schneider et al., 2003). Insert shows an ombrothermal diagram of Puerto Natales (www.climate-data.org). **C)** Study area
 134 Cerro Benitez with caves (yellow stars) and peatbogs (blue circles) mentioned in this study.
 135

136 3 Methods

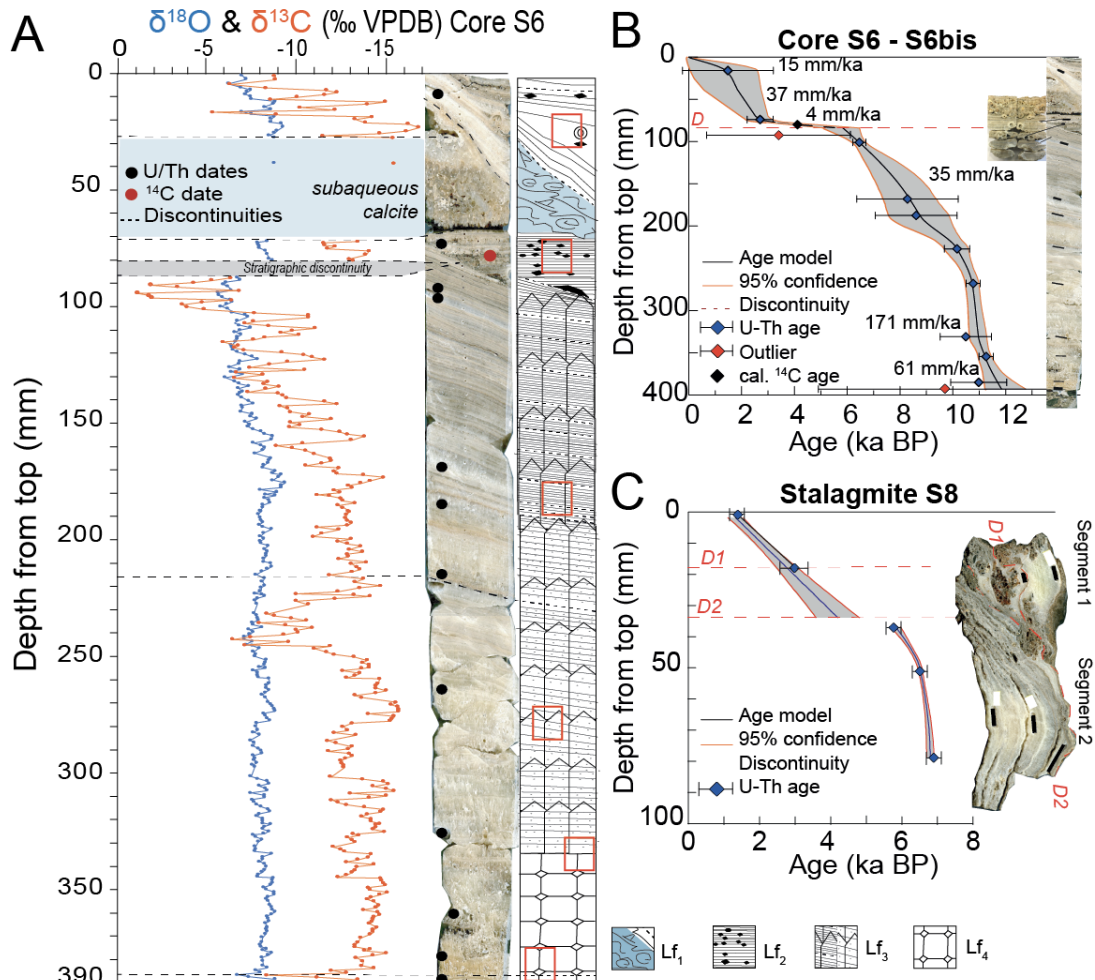
137 3.1. Petrography

138 Both S6 and S6bis cores and stalagmite S8 were cut along their growth axes, polished using 120–4000 μm
 139 silicon carbide sandpaper. Twelve thin sections (80–100 μm thick) were prepared at the department for
 140 Sediment and Isotope Geology, Ruhr University Bochum, Germany. A petrographic analysis was conducted
 141 with a Leica optical microscope to determine the stratigraphical characteristics of core S6 (e.g., fabrics,
 142 structures, inclusions, discontinuities, hiatuses).
 143

144 3.2. U-series and ^{14}C dating

145 Seventeen samples of between 30 and 230 mg were extracted from the sectioned cores S6 and S6bis and
 146 stalagmites S8 for U/Th dating using 0.5 to 0.8 mm tungsten carbide dental bits mounted on a Sherline 2010
 147 micromill. Samples were dissolved in isolation using 1.5M HNO_3 , spiked with a measured quantity of ^{233}U -
 148 ^{229}Th or ^{236}U - ^{233}U - ^{229}Th mixed synthetic isotopic tracer (Hellstrom, 2003; Drysdale et al., 2012) and allowed
 149

150 to equilibrate on a hotplate overnight. U and Th were separated from the carbonate matrix using Eichrom
 151 TRU ion exchange resin and established methods (Luo et al, 1997), modified to collect U and Th in the same
 152 fraction using 0.1M HCl–0.2M HF (Hellstrom, 2003). Mass spectrometric analysis was undertaken at the
 153 University of Melbourne using either a Nu Instruments Plasma MC-ICP-MS with standard collector block
 154 or a similar instrument with modified collector block placing simultaneously using the high- and low-mass
 155 SEM ion counters in each case (Hellstrom, 2003). Ion counter gain was determined using ion counter and
 156 Faraday measurements of ^{235}U and ^{229}Th for the standard collector block and $^{236}\text{U}/^{233}\text{U}$ and ^{229}Th for the
 157 modified collector bloc mass bias was determined using the bulk earth natural $^{238}\text{U}/^{235}\text{U}$ ratio of 137.88, this
 158 technique being relatively insensitive to error in this ratio due to the simultaneous measurement of
 159 $^{229}\text{Th}/^{230}\text{Th}$ and $^{233}\text{U}/^{234}\text{U}$. Ion counter baselines were measured at half mass units and interpolated for ^{234}U
 160 and ^{230}Th . Initial $^{230}\text{Th}/^{232}\text{Th}$ activities were determined by stratigraphic constraint (Hellstrom, 2006) where
 161 possible, giving a value of 0.93 ± 0.08 .
 162 An additional ^{14}C date was obtained through the analysis of 10 mg of charcoal sampled on core S6bis. The
 163 preparation and measurement were completed at the Radiochronology lab from Laval University, Québec,
 164 Canada. Radiocarbon concentrations are given as fractions of the modern standard, $\delta^{14}\text{C}$, and conventional
 165 radiocarbon age, following the conventions of Stuiver and Polach (1977). Result was corrected for isotope
 166 fractionation according to the conventions of Stuiver and Polach (1977), with $\delta^{13}\text{C}$ value measured on the
 167 graphite prepared, using the AMS spectrometer. Stratigraphic correspondence (Suppl., fig. 3) combined with
 168 U-Th dates from core S6bis permits the inclusion of the radiocarbon date into the age model, after the
 169 calibration (cal. BP) of the measured age (Suppl., fig. 5), using OxCal program (Bronk Ramsey, 2009) and
 170 the atmospheric data from Reimer et al. (2020).



171 **Fig. 2.** Lithofacies, stable isotope and age models for the studied samples. **A.** Stable isotope depth profile along the S6 core
 172 together with U/Th and ^{14}C dates and lithofacies. Explanation of lithofacies and red rectangles locating the thin sections are
 173 displayed in the supplementary (Suppl. S2). **B.** Age-depth model of core S6 and S6bis. **C.** Age-depth model for stalagmite S8.
 174 Age-depth models are based on COPRA (Breitenbach et al., 2012).
 175
 176
 177

3.3. Calcite and water stable isotopes measurements

178 Stable isotope samples were taken along the growth axis of core S6 and stalagmite S8 (Fig. 2) for $\delta^{13}\text{C}$ and
179 $\delta^{18}\text{O}$ measurements. Overall, 310 samples were measured in core S6 and 68 in stalagmite S8. Samples were
180 drilled along the stalagmite growth axis at 1 mm resolution using a Merchantek Micromill mounted on a
181 Leica microscope. Between every sample, the drill bit and sampling surface were cleaned with compressed
182 air. In addition, two recent calcite samples underneath active drip water were collected nearby the sampling
183 area and measured for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$. The samples were analysed using a Nu Carb carbonate device coupled
184 to a Nu Perspective mass spectrometer (MS) at the Vrije Universiteit Brussel, Belgium. Part of the samples
185 were measured at the Sediment and Isotope Geology department, Ruhr University Bochum, Germany, using
186 a Thermo Gasbench II connected to a Thermo MAT253 Plus MS in continuous flow mode. All $\delta^{18}\text{O}$ and
187 $\delta^{13}\text{C}$ values are calibrated against Vienna Pee-Dee Belemnite (V-PDB) and are reported in permil (‰).
188 Analytical uncertainties were better than 0.1 ‰ (1σ) for oxygen and 0.05 ‰ (1σ) for carbon on both
189 instruments.

190 Sixteen water samples were collected from the Cueva Chica and nearby caves, springs and lakes in December
191 2016, 2017, and 2018 for $\delta^{18}\text{O}$ and δD measurements in hermetically sealed glass bottles. Isotope
192 measurements were performed at the Vrije Universiteit Brussel on a Picarro L2130-i analyzer using the
193 cavity ring-down spectroscopy (CRDS) technique (Van Geldern and Barth Johannes, 2012). All values are
194 reported in permil relative to Vienna Standard Mean Ocean Water (VSMOW). Analytical uncertainties (1σ)
195 were better than 1 ‰ for δD and 0.1 ‰ for $\delta^{18}\text{O}$.

196

197 3.4. μXRF element data

198 The elemental distribution of Ca and Sr has been determined in core S6 using a μX -ray fluorescence (μXRF)
199 scanning. The scanning maps were selected along the stable isotope track using an Bruker M4 Tornado
200 μXRF spectrometer at the GFZ (Potsdam, Germany). The μXRF is equipped with a 50W Rh X-ray source
201 that was operated at 50 kV and 600 μA with a spot size of 20 μm . Measurements were performed under
202 vacuum every 50 μm and a counting time of 25 ms. The resulting element intensities are given in counts per
203 second (cps) and Sr element intensities were normalized over Ca to minimize sample density effects. The
204 Sr/Ca ratios were scaled by multiplying the intensity ratios by 1000, and transferred on to the age scale
205 through the age–depth model.

206

207 3.5. Fluid inclusions stable isotope (H-O) analyses

208 Fluid inclusion δD and $\delta^{18}\text{O}$ values from core S6 were obtained at the Max Planck Institute for Chemistry
209 in Mainz, Germany. Nine samples, consisting of 1 to 2 chips (Suppl., fig.7) weighing in total 0.6-0.8 g were
210 placed in a crushing cell and pre-heated at 115°C in a moisturized nitrogen flow for 20 minutes before
211 crushing. Liberated fluid inclusion water was analysed in a Picarro 2140i analyser. The contribution of the
212 ~13000 ppm water background in the N_2 carrier gas, which serves to remove sample-to-sample memory
213 effects in this analytical system, is mathematically subtracted from the isotope values analysed for the
214 crushed sample (Affolter et al 2014; De Graaf et al., 2020). A detailed description of the analytical technique
215 as it is set up at MPIC can be found in De Graaf et al. (2020). The Cueva Chica data were collected in two
216 different series. For each series more than 10 aliquots of standard water were injected in the crusher system,
217 analysed identically to the crushed samples and reproduced within 0.09 ‰ for $\delta^{18}\text{O}$ and 0.3 ‰ for δD for
218 both series (1σ). All fluid inclusion isotope data are reported on the VSMOW2-SLAP2 scale.

219

220 3.6. Clumped isotopes

221 Clumped isotope analyses were carried out at the AMGC laboratory of Vrije Universiteit Brussel (VUB),
222 Belgium, using a Nu-Perspective-IS-IRMS combined with a Nu-Carb system. The full methodology for the
223 Δ_{47} analyses performed at the AMGC-VUB lab is described in De Vleeschouwer et al. (2022). A total of
224 seven samples (Suppl., fig.7) from core S6 were analysed and replicated 10 to 15 times (Suppl., table 2). For
225 each analysis, a minimum of 500 μg of carbonate was reacted at 70°C with H_3PO_4 acid. Possible
226 contaminations were tracked by scrutinising raw Δ_{49} values of ETH1-4 standards for significant deviations
227 from the mean (Meckler et al., 2014). ETH standards were measured following the recommendations of
228 Kocken et al. (2019) with a sample to standard ratio of 1:1. Analyses were constantly monitored using the
229 Easotope software (John and Bowen, 2016). The raw Δ_{47} values were processed using the IUPAC isotopic
230 parameters (Brand et al., 2010; Daëron et al., 2016; Petersen et al., 2019) within the ClumpyCrunch software
231 (Daëron, 2021) and converted to the I-CDES 90°C scale (Bernasconi et al., 2021). The Δ_{47} values were
232 converted to temperatures using the regression of Anderson et al. (2021). Temperature uncertainties were
233 calculated by propagating the standard error (SE) for each sample and the uncertainties associated with the

234 calibration. The Δ_{47} uncertainties (SE) are based on external Δ_{47} reproducibility of our analytical sessions of
 235 measurements, based on repeated analyses of standards and samples.

237 4 Results

238 4.1. Petrography

239 Petrographic analysis of thin sections (Suppl., fig. 2) shows that core S6 is mainly composed of columnar
 240 calcite, which ranges from poorly-to-well laminated to massive, often with a persistent component of detrital
 241 impurities. Based on clastic content, fabrics, and colour of the calcite, four main lithofacies are identified in
 242 core S6. From base to top (Fig. 2) we observe:

- 244 i) Lithofacies Lf₄, from 392 to 335 mm from top, rich in clastic material and characterised by yellowish
 245 calcite with poorly laminated and more porous structure embedded in a micro-gour structure. Lf₄ reveals
 246 a low but significant content of oxides, rutile and microcalcite fragments (e.g., micrite, sparite) filling the
 247 voids between the calcite crystals.
- 248 ii) Lithofacies Lf₃, from 335 to 98 mm, composed of compact, elongated columnar calcite with highly
 249 laminated structure. Few empty voids are found compared to Lf₄. While the basal part of Lf₃ shows few
 250 impurities and more, well recognizable, fluid inclusions, the upper part shows more impurities and a
 251 higher number of hiatuses.
- 252 iii) Lithofacies Lf₂, from 98 to 50 mm, less clearly laminated but comprises the highest number of impurities,
 253 minerals, bones fragments (microfaunal), and remains of organic material (e.g. amorph material,
 254 unidentified plant remains in the entire record. Oxides, rutiles and pyroclasts are the most common
 255 identified features.
- 256 iv) Lithofacies Lf₁, from 50 mm to the top of the core reveals two kind of microstructures: it is poorly
 257 laminated in its upper part with few impurities whereas its lower part shows a primary calcite formed in
 258 a subaqueous environment.

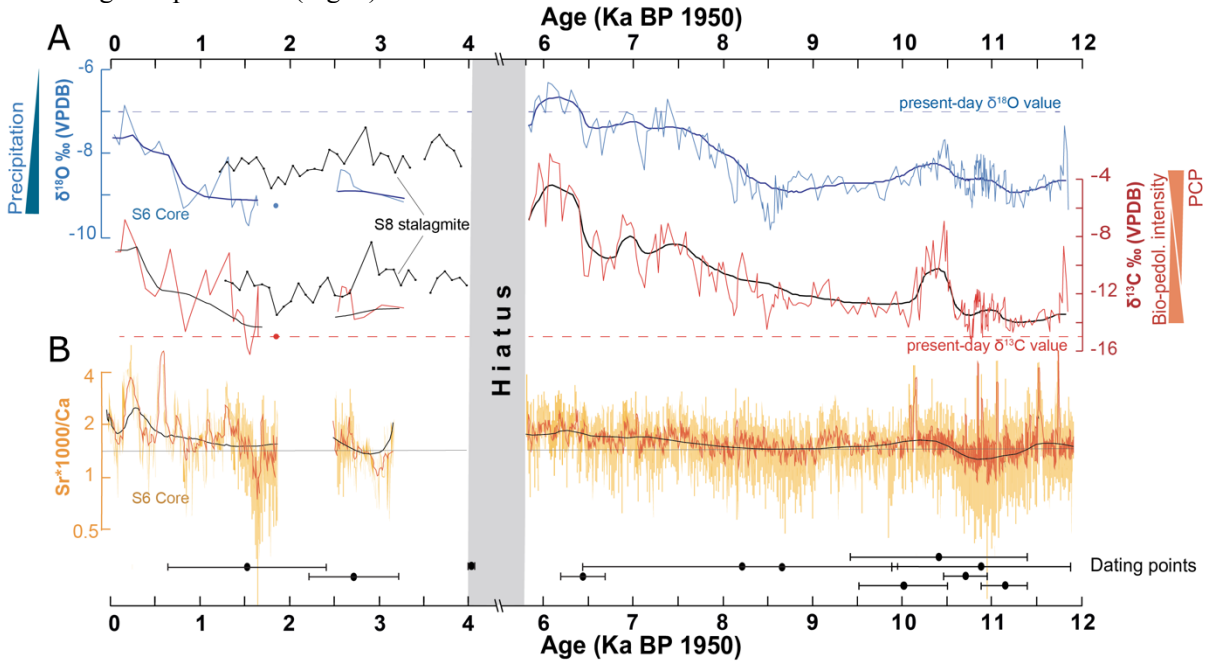
259 4.2. Chronology

260 Ten uranium-thorium (U/Th) ages were obtained from S6, two dates from S6bis cores and five from
 261 stalagmite S8 (Suppl., table 1). Two dates were considered as outliers. Minor age inversions with large
 262 uncertainties coincide petrographically with growth intervals rich in detrital material in the flowstone cores
 263 (Suppl., fig 3, table 1). High initial thorium concentrations are most likely due to organic material, or iron
 264 oxides that are common at depths 390-370 mm and 90-70 mm in core S6. The corresponding dates S6-19-
 265 10 and S6bis-01 are out of sequence and were therefore excluded from the age model (Suppl., table 1). The
 266 COPRA age-depth model indicates that core S6 grew from *ca.* 11.1 ± 1 to 1.5 ± 1.7 ka BP, including several
 267 short discontinuities (Fig. 2B). One noteworthy discontinuity at 80 mm depth in core S6 shows a change in
 268 growth (water flow) direction, followed by precipitation of impurities-rich calcite. This discontinuity covers
 269 the period from ~ 5.8 to ~ 2.7 ka BP. Considering that a thin water film is still running today over the wet
 270 flowstone, the age model was bound with the year of sampling in its youngest part. Core S6bis, sampled 10
 271 cm from core S6, grew between 3.4 ± 2.7 ka and 2.7 ± 0.5 ka BP, with a charcoal encrusted in the calcite
 272 and dated at 4.11 ± 0.04 cal BP. Stalagmite S8 grew from 6.8 ± 0.1 to 1.29 ± 0.22 ka BP (Fig. 2B) and a
 273 hiatus is observed between 5.7 and 4.2 ka BP. A series of dark detrital layers filled with microfaunal bone
 274 fragments mark two major discontinuities, D1 at 18 mm and D2 at 22 mm depth. The latter separates segment
 275 1, dated to between ~ 4.2 and ~ 1.3 ka BP, from segment 2, dated to between ~ 6.7 ka and ~ 5.7 ka BP.

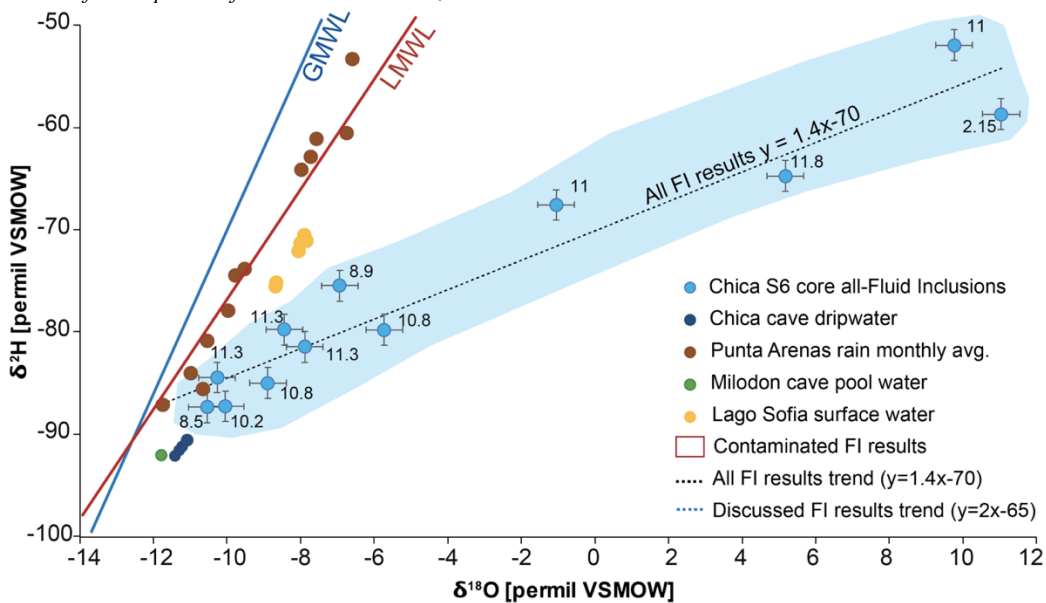
276 4.3. Stable isotopic composition of calcite, modern water and fluid inclusions

277 Stalagmite core S6 $\delta^{18}\text{O}$ (-6.28 to -9.81 ‰) and $\delta^{13}\text{C}$ (-2.14 to -16.26 ‰) values display distinct signatures
 278 (Fig. 3). Between ~ 12 and ~ 8.5 ka BP, the time series shows overall low $\delta^{18}\text{O}$ values, with the lowest
 279 observed at ~ 8.5 ka BP. A distinct positive excursion is noticeable between 10.5 and 10.1 ka BP. From 8.5
 280 to 6 ka BP, we find a general increase in both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values and higher multi-decadal scale variability
 281 (Fig. 3). The highest isotope values (-6.3 ‰ for $\delta^{18}\text{O}$, -2 ‰ for $\delta^{13}\text{C}$) are found at ~ 6 ka BP. From 3.5 to 1.5
 282 ka BP, a shift to lower carbon and oxygen isotope values in core S6 corresponds well with stable isotope
 283 profile from stalagmite S8 (Fig.3). After 1.5 ka BP, the upper part of core S6 shows an increase in both the
 284 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values until the present day. The overall resolution of the isotope record is infra- to multi-
 285 decadal, except for the topmost part (from ~ 3.2 to present), which is lower resolved. At a decadal scale, the
 286 $\delta^{13}\text{C}$ profile co-varies with the $\delta^{18}\text{O}$ record. High growth rate (Fig. 2) generally corresponds with lower $\delta^{18}\text{O}$
 287 and $\delta^{13}\text{C}$ values. After ~ 8.5 ka BP, the growth rate decreases (Fig. 2B) drastically until ~ 5.8 ka BP, coeval
 288 with a slowly increasing trend in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

291 Dripwater stable isotopes collected in the Cueva Chica show an average $\delta^{18}\text{O}$ value of -11.24‰ and a δD
 292 of -91.23‰ . The isotope values cluster very narrowly and fall at the lower end of, and parallel to, the local
 293 meteoric water line (LMWL) (Fig. 4). The corresponding recent calcite deposits that form from the sampled
 294 dripwater show a $\delta^{18}\text{O}$ value of -7.05‰ and $\delta^{13}\text{C}$ of -15.2‰ . Stable isotopes from fluid inclusion exhibit a
 295 wide (20 ‰) range of $\delta^{18}\text{O}_{\text{FI}}$ values, which is much larger than the $\delta^{18}\text{O}$ range in the host calcite (3.5 ‰).
 296
 297 The fluid inclusion deuterium isotope ratios ($\delta\text{D}_{\text{FI}}$) data show a 30 ‰ range. In $\delta\text{D} - \delta^{18}\text{O}$ space the lowest
 298 observed isotope ratios are found close to the GMWL, while values progressively depart from the GMWL
 299 following a slope of ~ 1.4 (Fig. 4).



300
 301 **Fig. 3.** Multiproxy ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and Sr/Ca ratios) records from samples S6 and S8. **A.** The original isotope. **B.** Ratio of μXRF
 302 Sr and Ca intensities (cps) presented as median values (red smoothed line) to highlight the broader trends and variations
 303 from 12 ka BP to present. The moving average (black lines) is shown for both stable isotope and Sr/Ca profiles. No Sr/Ca
 304 data is available for the periods from 5.5 to 3.2 ka BP, and 2.5 to 2 ka BP.



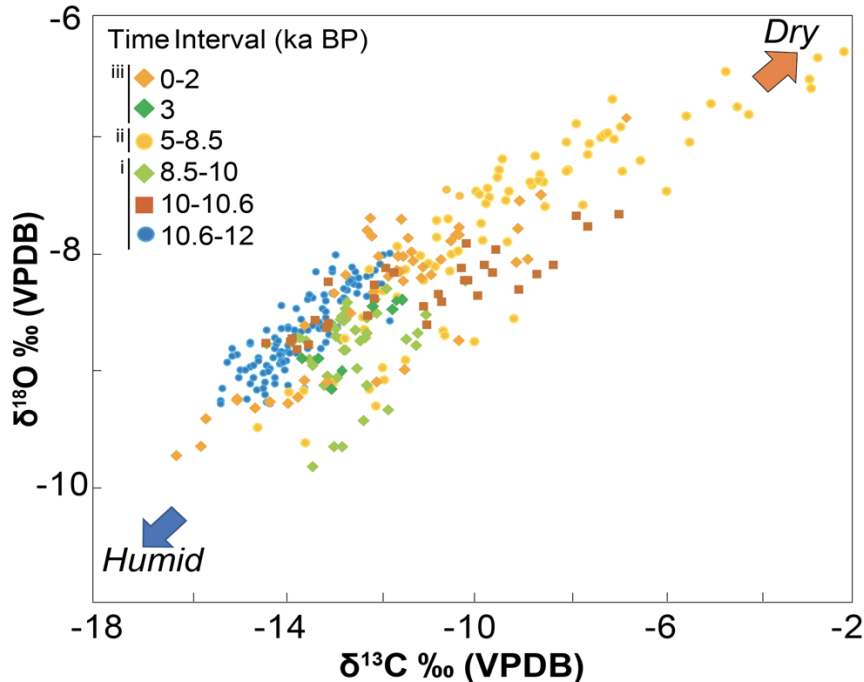
305
 306 **Fig. 4.** Results of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ analyses of 12 fluid inclusion (FI) samples from core S6 along with modern cave pool,
 307 dripwater, river and nearby Lago Sofia (this study) as well as rainwater (IAEA/WMO, 2019) of the nearest monitored station
 308 in Punta Arenas, plotted together with the LMWL and GMWL (Rozanski et al., 1993). The age of each FI sample is indicated
 309 in the graph and the SE for each FI is better than 0.1‰ for $\delta^{18}\text{O}$ and 0.3‰ for $\delta^2\text{H}$ (1σ). The FI distribution shows the most
 310 contaminated/dirty samples on the right while clean (in terms of detrital content) samples to the left.
 311

312 4.4 High resolution Sr/Ca data

313 The Sr/Ca record of core S6 has a temporal resolution of about 0.1 to 12 years (Fig. 3). The 1000*Sr/Ca ratio
 314 varies between 0.2 and 325 cps (avg = 2). The median record (shown in red) varies only little between 1.3
 315 and 2.6 cps but reveals some maxima between prior to 11,5 ka, from ~10,5 and 10 ka BP and from ~1.0 to
 316 present-time ka. Minima occur from ~10.8 to 11.5, ~1.8 and ~3 ka BP. From ~12 to 11.4 ka BP, Sr/Ca
 317 values are higher (median ~1.6) followed by a peak of lower values from ~11.3 to 10.6 ka BP (median
 318 ~1.3). A return to higher values is marked at 10.2 ka and from 9.6 to 9.2 ka BP, followed by a general trend
 319 of increased Sr/Ca values (median ~1.6) from ~8.5 to 5.8 ka BP (median from ~1.5 to 1.9), which coincides
 320 with high stable isotope values. From 3.2 ka BP onwards, the Sr/Ca median values are low (from ~1.4 to
 321 1.5), except for a short excursion at 0.7 and 0.3 ka BP. More rapid changes in the Sr/Ca ratios (median),
 322 showing annual to multi-annual variations, are superimposed on this long-term trend (as shown by the
 323 moving average). The Sr/Ca and stable isotope time series vary simultaneously, with higher Sr/Ca values
 324 corresponding with more positive $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values.

326 4.5. Clumped isotope values

327 The Δ_{47} values of the 8 analyzed samples range from 0.6172 to 0.6601 ‰ (Suppl., table 4). From the bottom
 328 to the top (older to younger part of the speleothems), the Δ_{47} values progressively decrease from 0.6528 to
 329 0.6438 ‰ between 11.3 and 6.5 ka BP. The minimum Δ_{47} value is reached with sample C3 at 6.5 ka BP. At
 330 the top of the speleothem, both samples C2 and C1 record the highest Δ_{47} value, 0.6601 and 0.6565 ‰
 331 respectively. The sample C0, taken at the top of the speleothem, includes the most recent calcite and records
 332 a Δ_{47} value of 0.6415 ‰.



333 **Fig. 5.** Stable oxygen and carbon isotope variability over time, showing their dispersion and interrelationship for different
 334 time intervals. Three phases are identified: i) low $\delta^{13}\text{C}$ (avg: ~ -15 ‰) and $\delta^{18}\text{O}$ values (avg. ~ -9 ‰) during the early-
 335 Holocene wet period (with exception of two excursions at 10.5 ka and 8.5 ka), ii) high $\delta^{13}\text{C}$ (avg. ~ -7 ‰) and $\delta^{18}\text{O}$ values
 336 (avg. ~ -7 ‰) during the mid-Holocene dry period, iii) general wet conditions in the late Holocene with low $\delta^{13}\text{C}$ (avg. ~ -12
 337 ‰) and $\delta^{18}\text{O}$ values (avg. ~ -8 ‰).
 338
 339

340

341 5 Discussion

342

343 5.1 Age growth behaviour of the studied speleothems

344 Building a composite COPRA age-depth models for cores S6a and nearby S6bis required lithofacies
 345 comparison of each core under optical microscope for the same depth interval. Optical observation allowed
 346 to relate several stratigraphical levels along with an additional ^{14}C dating on charcoal in core S6bis (Suppl.,
 347 fig., 3) and consider all the ages of core S6 and S6-bis on the same depth profile. The age model for core S6-
 348 S6bis (Fig. 2) reveals a rather continuous deposition from ~11.9 to ~5.8 ka BP, with an average growth rate
 349 of 90 mm/ka (range 4-300 mm/ka). The basal part from ~12 to ~9 ka BP shows a high growth rate of 170

mm/ka at the beginning of calcite precipitation in the cave. From ~9 ka onwards, the growth rate decreases and remains at/or below 37 mm/ka until ~5.8 ka BP, showing an overall gradual decrease in water availability to build the flowstone. At ~5.8 ka BP, a stratigraphical discontinuity (D) is found, likely resulting from a change in flow direction of the source water. This core section either precipitated at very slow growth rates of 4 mm/ka or contains multiple (unresolved) hiatuses. Charcoal fragments encrusted between clean and muddy calcite layers within D, dates to 4.11 ± 0.04 cal. BP and could be the result of an enhanced incave water flow event after a vegetation fire (with soil surface erosion by overland flow). The speleothem record continues at 2.7 ± 0.5 ka BP and calcite formed continuously but slowly until the present time, resulting of less favourable conditions for intensive calcite precipitation.

Stalagmite S8 grew from ~6.8 to 2 ka BP, with an average growth rate of 68 mm/ka. A period of apparently very slow growth between 5.8 and 4 ka BP is attributed to either significant drying or one or more hiatuses that remain unresolved with the current dating information. This period of limited growth in stalagmite S8 coincides with the timing of the stratigraphical hiatus in core S6. This and less negative $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values at the time (Fig. 4) signify reduced infiltration and less favourable conditions for intensive calcite precipitation in the cave. Local pollen data confirms our interpretation of drier conditions at the time (McCulloch et al., 2021).

5.2 Interpreting the geochemical proxy record from the Cueva Chica

Stable isotope data from speleothem have been intensively studied in moderate temperate regions (Baldini et al. 2021, McDermott, 2004; Mangini et al., 2005; Lachniet, 2009; Fairchild and Baker, 2012), but less in cold temperate areas such as Patagonia (Schimpf et al., 2011).

5.2.1. $\delta^{13}\text{C}$ as proxy for vegetation and soil development

Stalagmite $\delta^{13}\text{C}$ can serve as qualitative proxy for changes in vegetation composition and density, soil microbial activity, and, frequently, for changes in infiltration dynamics (Lechleitner et al., 2018; Fohlmeister et al. 2011). At the Cueva Chica, soil and vegetation activity are highest at the end of spring (Nov-Dec). Together with higher effective infiltration during the winter-spring period, this results in higher CO_2 concentrations with low $\delta^{13}\text{C}$ signature in the soil. Passing through the epikarst, the $\delta^{13}\text{C}$ signal is adjusted to higher values, depending on the residence time of the percolating water and length of water-rock interaction (Lechleitner et al., 2018). In addition, CO_2 degassing can occur i) in the epikarst if air pockets exist in the dry season, leading to prior carbonate precipitation or ii), in a ventilated cave atmosphere (Deininger et al., 2012), in both cases contributing to a shift of $\delta^{13}\text{C}$ towards higher values. The speleothems from the Cueva Chica thus record the combined signal of epikarst and in-cave CO_2 degassing, vegetation composition and density as well as soil microbial activity (McDermott, 2004). The observed very low $\delta^{13}\text{C}$ values in the Chica record, reaching values such as -16 ‰, corroborate the interpretation that multiple processes, including microbial and plant activity, influence the $\delta^{13}\text{C}$ signal. In South Patagonia, the vegetation type is dominantly a C3 type in a broad-leaved forest cover, sometimes mixed with C4 type in open grassland (Cerling and Quade, 1993). Today the Cerro Benitez is covered with *Juncaceae* (rushes) and grasses mixed with sparse forest on thin soils, and several fires have been recorded throughout the Holocene (McCulloch et al., 2021). Thus, the $\delta^{13}\text{C}$ signal of postglacial soils above the cave is expected to be much higher than that of a developed *Nothofagus* forest soil ($\delta^{13}\text{C}$: -26 ‰ in Peri et al., 2012). More open vegetation or thin soils with limited microbial activity and soil CO_2 production would lead to higher $\delta^{13}\text{C}$ (Scholz et al., 2012; Genty et al., 2001, 2003).

5.2.2. Sr/Ca as proxy for infiltration dynamics

Speleothem Sr/Ca constitutes a sensitive proxy for infiltration changes in the epikarst, with higher values being recorded during times of reduced effective moisture availability (Fairchild and Treble, 2009). Other mechanisms can explain Sr mobility within the epikarst and include changes in weathering rates, aerosol input, soil activity (Verheyden et al., 2000; Fairchild et al., 2006; Baker et al., 2021; Sinclair et al., 2012) that can induce high-frequency variability on short timescales (decadal to annual). At the Cueva Chica, speleothem Sr/Ca reacts to local infiltration changes and climate seasonality with i) high precipitation and effective infiltration in austral winter and, ii) low precipitation and high potential evapotranspiration during austral summer (Fig. 1B). Long-term drier climate with shorter wet seasons and low precipitation can lead to open conditions in soil and epikarst, and strong CO_2 degassing from infiltrating water and in the cave atmosphere. These processes can lead to PCP within the epikarst (Verheyden et al., 2000; Fairchild et al.,

2006), which in turn leads to higher Sr concentration in the aqueous solution and ultimately in the speleothem (Verheyden et al., 2000; Fairchild and Treble, 2009).

408

409 5.2.3. Speleothem and dripwater $\delta^{18}\text{O}$

410 Calcite $\delta^{18}\text{O}$ is generally interpreted as related to precipitation amount, atmospheric circulation pattern,
 411 seasonal changes in infiltration (McDermott, 2004; Dreybrodt and Scholtz, 2011). Additionally, local factors
 412 can affect $\delta^{18}\text{O}$ in the epikarst and cave, including evaporation and CO_2 degassing, temperature, and mixing
 413 of older and younger water in the epikarst (Mühlinghaus et al., 2009). Chica cave receives precipitation
 414 mainly from the Pacific Ocean through WSW winds. These waters are characterized by low $\delta^{18}\text{O}$ in
 415 precipitation ($\delta^{18}\text{O}_p$) values (Suppl. S4) and fall on the GMWL, while water collected during spring and
 416 summer show high $\delta^{18}\text{O}_p$. Effective infiltration and epikarst recharge are biased towards the winter season
 417 and possibly early summer. Dripwater collected in Chica and nearby Milodon caves in early summer show
 418 $\delta^{18}\text{O}_{\text{drip}}$ values biased towards autumn-winter $\delta^{18}\text{O}_p$. Unfortunately, no dripwater is available from other
 419 seasons.

420

421 Even though the sampling site is deep inside the cave (60 m from the entrance) with high relative humidity,
 422 we cannot strictly exclude the possibility that evaporation affected dripwater $\delta^{18}\text{O}$. Cave ventilation could
 423 lower relative humidity sufficiently to induce evaporation. The temperature effect on $\delta^{18}\text{O}_{\text{drip}}$ is unlikely a
 424 major factor because cave air temperature does not vary significantly ($T_{\text{cave air}} = 4.6 \pm 0.6^\circ\text{C}$).

425

426 Finally, mixing of infiltration waters of different age in the epikarst likely results in the observed relatively
 427 stable $\delta^{18}\text{O}_{\text{drip}}$ values through a year (Suppl. fig. 6, table 2). The small offset between the two S6 and S8
 428 speleothems suggests some degree of mixing along the individual water pathways to the two sampling sites.
 429 The water residence time in the epikarst is likely on the order of months to years. The epikarst thickness
 430 above Chica cave (40 m) is similar to that observed at Ernesto (Italy), Han sur Lesse (Belgium) or Bunker
 431 (Germany) caves (epikarst thickness between 15 and 30 m, Van Rampelbergh et al., 2014; Kluge et al., 2010;
 432 Miorandi et al., 2010) where lags of several months to years between surface and cave are found. Given the
 433 observed narrow range of $\delta^{18}\text{O}$ in dripwaters (Fig. 4) and infiltration pattern (Suppl., fig. 6) we suggest that
 434 both dripwater and speleothem $\delta^{18}\text{O}$ values represent a buffered (multi-)annual mean signal, likely biased
 435 towards the winter season.

436

437 The Chica $\delta^{18}\text{O}_{\text{cc}}$ record (Fig. 3A) thus integrates a complex signal of precipitation history and corresponding
 438 $\delta^{18}\text{O}_p$. Earlier work on three speleothems combined with intensive cave microclimate monitoring in Marcello
 439 Arévalo (53°S), an open-air cave located further south, showed that $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ were both linked to
 440 effective infiltration (Schimpf et al., 2011). There, the most suitable stalagmites preserving the link between
 441 stable isotope signal and effective infiltration in MA cave were the ones showing high and constant growth
 442 rates ($>67 \mu\text{m/a}$). In our case, the flowstone S6-S6bis in the Chica cave grew mostly continuously, with
 443 limited growth from 5 to 3 ka BP, and growth rates suggest that S6 was regularly submerged. A high growth
 444 rate of $132 \mu\text{m/a}$ is found prior to 9 ka that subsequently slowed to *ca.* $33 \mu\text{m/a}$ over the course of the
 445 Holocene until recent time. Stalagmite S8 grew relatively fast at a rate of $111 \mu\text{m/a}$ from 6 to 5 ka BP, and
 446 slowly at a rate of $11 \mu\text{m/a}$ after 4 ka BP. We suggest then that variations in the Chica $\delta^{18}\text{O}$ signal reflect
 447 mainly changes in amount of precipitation and effective infiltration, and to some extent changing moisture
 448 transport trajectories throughout the Holocene.

449

450 5.3. General climatic conditions during the Holocene in southwest Patagonia

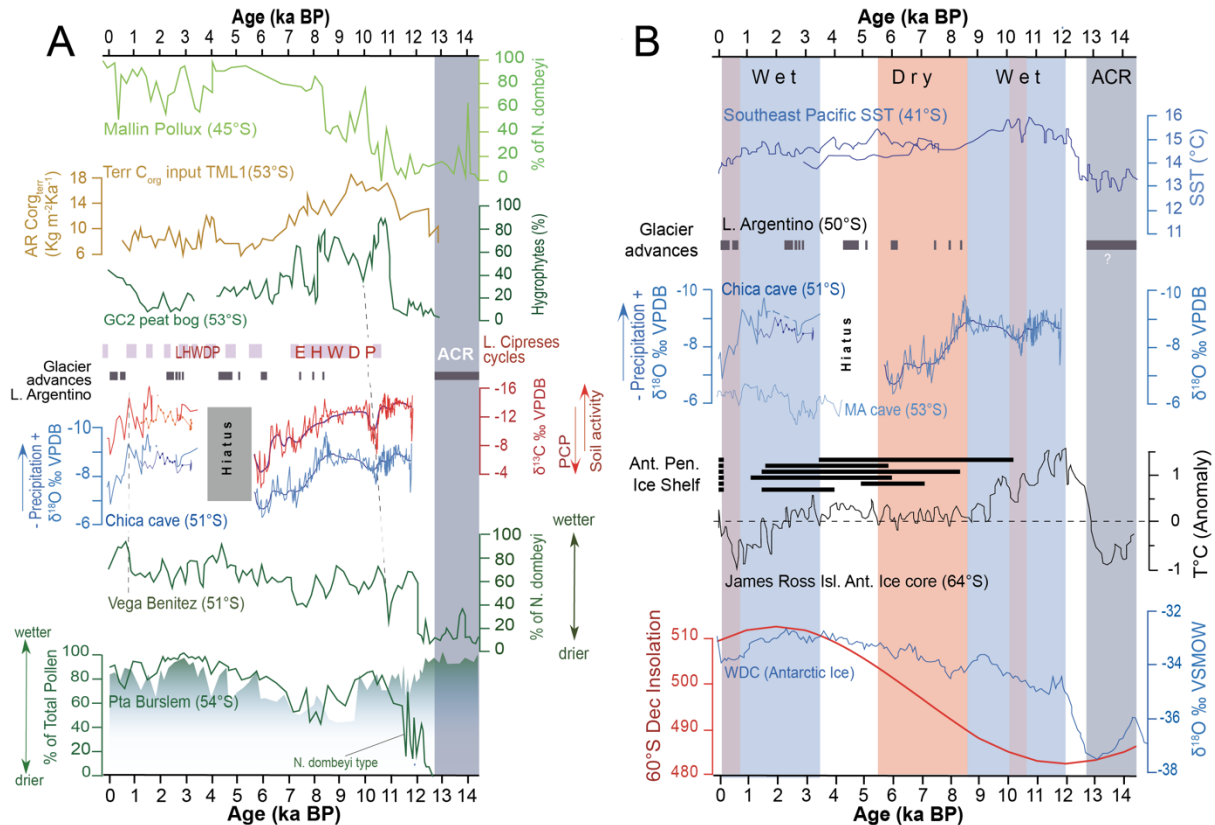
451 The combined stable isotope, growth rate, and Sr/Ca time series suggest three climatic regimes throughout
 452 the last 12 ka in Southern Patagonia: i) an early Holocene wet period (with the exception of two dry
 453 excursions at 10.5 ka and 8.5 ka BP), ii) a mid-Holocene dry period and iii), a return to generally wet
 454 conditions over the late Holocene.

455

456 5.3.1. The early Holocene humid period (12-8.5 ka BP)

457 During the early Holocene period, the $\delta^{18}\text{O}_{\text{cc}}$ values are low until 8.5 ka BP, except for two excursions at
 458 10.5 ka and 8.5 ka BP (Fig. 6). The $\delta^{18}\text{O}$ values are the lowest of the entire Holocene record and vary between
 459 -7 and -9‰ , suggesting relatively wet conditions. This notion is supported by a high growth rate of *ca.* 60 -
 460 $170 \mu\text{m/year}$. The flowstone grew continuously under sufficiently humid conditions to maintain infiltration
 461 in the epikarst zone and limited PCP as indicated by low Sr/Ca and low but highly variable $\delta^{13}\text{C}$ values (Fig.
 462 3A). The study site was likely covered by broad-leaved mixed C3/C4 type vegetation on a developed soil

463 with higher CO₂ production through biogenic activity. The onset of this wet early Holocene period agrees
 464 closely with abundant *N. dombeyi* pollen in the Vega Benitez peat profile (McCulloch et al., 2021),
 465 suggesting a developed woodland cover. A similarly developed *Nothofagus* forest was found in the Punta
 466 Burslem pollen records (McCulloch et al., 2020) further South. Other records from lake Tamar and peat GC2
 467 also suggest a wet period with a peak in humidity around 11 ka BP (Fig. 5A) (Lamy et al., 2010). Both, high
 468 abundance of hygrophytic taxa (a sensitive proxy for extremely humid environments) at peat GC2, and high
 469 terrestrial organic carbon concentrations linked to extreme rainfall events at lake Tamar (Lamy et al., 2010)
 470 are indicative of humid conditions.
 471



472 **Fig. 6.** The δ¹⁸O and δ¹³C records of Cueva Chica compared with global and regional records for the last 14 ka. **A.** (top to
 473 base) Comparison of continental records in Southern Chile, including the % *Nothofagus dombeyi* pollen from Mallin Pollux
 474 (Markgraf et al., 2007), terrestrial organic carbon in lake Tamar (Lamy et al., 2010), % hygrophytes from peat bog GC2
 475 (Lamy et al., 2010), lago Cipreses warm cycles (Moreno et al., 2018), moraine building events in the Lago Argentino basin
 476 (Kaplan et al., 2016) & glacial fluctuations in the Marinelli fjord (Hall et al., 2019), Cueva Chica isotope record (this study),
 477 % of *Nothofagus dombeyi* pollen from Vega Benitez (McCulloch et al., 2021) and % of *Nothofagus dombeyi* pollen vs total
 478 normal pollen from Punta Burslem (McCulloch et al., 2020). ACR timing is indicated by blue shade from Gest et al. (2017).
 479 **B.** (top to base) Comparison of global and regional records in Southern Chile, including sea surface temperatures (SSTs) from
 480 Geob3313-1 (Lamy et al., 2002) and ODP1233 (Lamy et al., 2015), moraine building events from Lago Argentino (Kaplan et
 481 al., 2016), Cueva Chica δ¹⁸O, Marcello Arévalo (MA) speleothem δ¹⁸O (Schimpf et al., 2011), periods of open marine
 482 conditions in Prince Gustav Channel, Antarctic Peninsula & T anomalies from James Ross Island (Mulvaney et al., 2012),
 483 δ¹⁸O from West Dome C (WDC, 2013) and 60°S december insolation (Berger & Loutre, 1991). Red rectangles indicate dry
 484 periods and blue rectangles wet periods defined by the Cueva Chica speleothem record.
 485
 486

487 The early Holocene dry episode begins with a positive excursion of δ¹⁸O and δ¹³C centred at 10.5 ka BP,
 488 concurrent with high Sr/Ca values. Between 10.7 ka and 10.2 ka BP (Fig. 6A), the Cueva Chica isotopic
 489 profile records significant drying with a short-lived low rainfall period impacting the vegetation cover on
 490 the Cerro Benitez. At the Vega Benitez site upstream of the Chica cave watershed, *N. dombeyi* forest declined
 491 from 11.54–10.78 cal ka BP and was replaced by *Poaceae* and *Cyperaceae* taxa. This shift in pollen
 492 assemblages is typical of a continued transition from lake to a peat bog and the increase in *Poaceae* and
 493 *Cyperaceae* likely reflects the spread of grasses and sedges over the emerging bog surface (McCulloch et
 494 al., 2021). This reduction of water surface suggests less water availability and overall reduced local
 495 precipitation, which supports the dry event recorded in the Chica record. The different timing of the dry
 496 event in the Chica samples (10.7-10.2 ka BP) and in nearby lake Vega Benitez (11.5–10.7 ka BP) is likely

497 related to limited chronological control in both archives. It is worth mentioning that comparable drying
 498 events are found in other regional records between 11 and 10 ka BP, including a rapid decline in *N. dombeyi*
 499 in the Punta Burslem peat bog (McCulloch et al., 2020) and a clear reduction in the abundance of hygrophytes
 500 in the GC2 peat bog (Lamy et al., 2010). After 10.2 ka BP, increased moisture availability is noticeable in
 501 the $\delta^{18}\text{O}$ profile with the most negative values of the whole record centred at 8.5 ka BP (Fig. 6A). The $\delta^{13}\text{C}$
 502 values start to become more negative, but not below -12 ‰. This wet phase is less prominent than the early
 503 Holocene period, but still noticeable in the Vega Benitez pollen record, with short-lived peaks in *N. dombeyi*
 504 pollen at 10.7 and 8.2 ka BP (McCulloch et al., 2021).
 505

506 **5.3.2. The mid-Holocene drying trend (8.5- 5.8 ka)**

507 After 8 ka BP, a slow trend towards higher $\delta^{13}\text{C}$ values in the Chica record is mirrored in a persistent rise in
 508 the number of deteriorated pollen and a corresponding decrease of well preserved pollen in most of the
 509 remaining record of Vega Benitez (McCulloch et al., 2021). Overall, the changing $\delta^{13}\text{C}$ trend and pollen
 510 assemblages as well as differential pollen preservation indicate reduced moisture availability. The trend in
 511 the $\delta^{13}\text{C}$ profile and a corresponding two permil shift to higher $\delta^{18}\text{O}$ shows the onset of a robust change in
 512 vegetation cover induced by reduced precipitation and less effective infiltration until 5.8 ka BP. The clear
 513 trend in the Chica record confirms regional drying as interpreted from reduced hygrophytic taxa in the GC2
 514 peat bog and lower terrestrial organic carbon concentrations in lake Tamar. Further South, low abundance
 515 of *N. dombeyi* between 10 ka and 7 ka BP in Punta Burslem (McCulloch et al., 2020) also suggests drier
 516 conditions. The response of all the records indicates a consistent drying trend across 50°S to 55.5°S. This
 517 trend seems to be contrasted by the Mallin Pollux pollen profile further North (Markgraf et al., 2007), which
 518 might mark a contrasting N-S climatic gradient, although additional records would be required to confirm
 519 this interpretation.
 520

521 **5.3.3. Return to wet conditions during late Holocene (4-0.5 ka BP)**

522 From 4 to 0.7 ka BP, decreasing $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values in the Chica record reflect a return of more humid
 523 conditions. This coincides with a relatively humid phase in the Vega Benitez (McCulloch et al., 2021) where
 524 four successive peaks in *N. dombeyi* abundance are observed at 3.3, 2.8, 2, and 0.7 ka BP (Fig. 6A). Low
 525 $\delta^{18}\text{O}$ in Marcello Arévalo (MA) speleothem (53°S) indicates wet conditions from 2.5 to 0.7 ka BP. Higher
 526 *N. dombeyi* concentrations in Burslem peat record confirm this interpretation, but no clear humid phase is
 527 found in Lake Tamar during the late Holocene. After 0.7 ka BP, our stable isotope record increases again
 528 and suggests the return of drier local conditions. This agrees with reduced *N. dombeyi* abundance in the Vega
 529 Benitez record and a similar trend in the MA stable isotope records.
 530

531 **5.4. Global and regional drivers of climate variability in SW Patagonia**

532 The flowstone in the Chica cave started to grow significantly from 12 ka onwards during a period of high
 533 Pacific SST (Lamy et al., 2002; 2015), with > 4°C warming after the Antarctic Cold reversal (ACR) (Fig.
 534 6B). A similar temperature shift is observed at the same time in the Antarctic James Ross Island ice core,
 535 and a trend towards higher $\delta^{18}\text{O}$ is found in Antarctic ice from West Dome C (Weiss divide project members,
 536 2013). Both observations mark the onset of deglaciation in Antarctica and a reduction in sea-ice cover.
 537 Despite minimum summer insolation at 11-12 ka BP in the Southern hemisphere (Berger and Loutre, 1991),
 538 the early Holocene was the warmest period in Southernmost South America, consistent with ice core data
 539 and the rise in SST in the South Pacific. This warming in both polar ice and Southern Ocean surface coincides
 540 with enhancement of the southwesterly winds (SWW) in the centre of the wind belt (Lamy et al., 2010), but
 541 is difficult to explain by orbital forcing. The warming during the early Holocene might be attributable to
 542 changes in global ocean circulation which involves a bipolar seesaw-like surface temperature pattern (Lamy
 543 et al., 2010) and a rise in air temperature in the southern hemisphere (Ganopolski et al., 1998; Hodell et al.,
 544 2001). A response of the SWW to changes in the strength of the Atlantic meridional overturning circulation
 545 (AMOC) has been previously proposed for Termination 1 (Lamy et al., 2007). Model experiments
 546 (Timmermann et al., 2005) suggest a rapid signal transfer between the AMOC and SSW shifts with an
 547 increase of SWW strength south of 50°S and decreased wind strength in the mid-latitudes between 40° to
 548 50°S. The response of terrestrial archives in Southern Patagonia to this atmosphere/ocean feedback is
 549 reflected in a reduction of glacier advances in Lago Argentino (50°S) and higher precipitation at Cerro
 550 Benitez (Cueva Chica) and the Chilean coast, likely in concert with higher temperatures.
 551

552 From 9 to 5 ka BP, a slight decrease in SE Pacific SST went hand in hand with a minor temperature decrease
 553 in on James Ross Island (64°S) (Fig. 6B). The most significant change in regional conditions was increased

554 ice-shelf instability along the Antarctic Peninsula (Mulvaney et al., 2012), that began around 10 ka BP and
 555 expanded gradually after 9 ka BP. The intensification of ice-shelf instability along the Antarctic Peninsula
 556 is concordant with increasing summer insolation in the Southern hemisphere, which intensified a poleward
 557 shift of the SWWs in response to global warming (McCulloch et al., 2020). Over the Patagonian landmass,
 558 this implied a reduction in moisture supply from the Pacific onto the Patagonian shore, lower total
 559 precipitation over the Chica cave and decreased glacier advances (warm/dry conditions) at Lago Argentina
 560 (50°S). After 5 ka BP, a different climatic background arises in Southern Patagonia, with SST and the James
 561 Ross Island ice core records indicating a gradual temperature decline (Fig. 6B). This pattern is consistent
 562 with an equatorward shift of the SWWs in response to a cooling Southern hemisphere, known as the
 563 Neoglacial period. Despite increasing insolation in the Southern hemisphere the Neoglacial cooling began
 564 at least *ca.* 5 ka BP (Glasser et al., 2004). In turn, increased AMOC exported heat from the South Atlantic
 565 to the North Atlantic, triggering sea-ice growth around Antarctica and glacial advances in Southern
 566 Patagonia such as at Lago Argentino (Fig. 6B). One global feedback at the end of the mid-Holocene is the
 567 establishment of modern El Niño–Southern Oscillation (ENSO) frequencies (Clement et al., 1999; Moy et
 568 al., 2002). The SWW storm tracks extended to lower latitudes, inducing a return to wetter conditions along
 569 the Patagonian coast, as found in both Marcello Arévalo speleothem and Chica records after 4 ka BP.
 570 However, moisture patterns were highly variable due to increased insolation seasonality (Markgraf et al.,
 571 2000) and intensified ENSO (Moreno, 2004). This increased variability with rapid climate change (RCC) is
 572 recorded in several lake records (Fletcher and Moreno, 2011; Moreno, 2004; 2018b), but less clear in Cueva
 573 Chica speleothems, although the $\delta^{18}\text{O}$ signal seems to be more variable.

575 5.5. Interpretation of fluid inclusion δD and $\delta^{18}\text{O}$

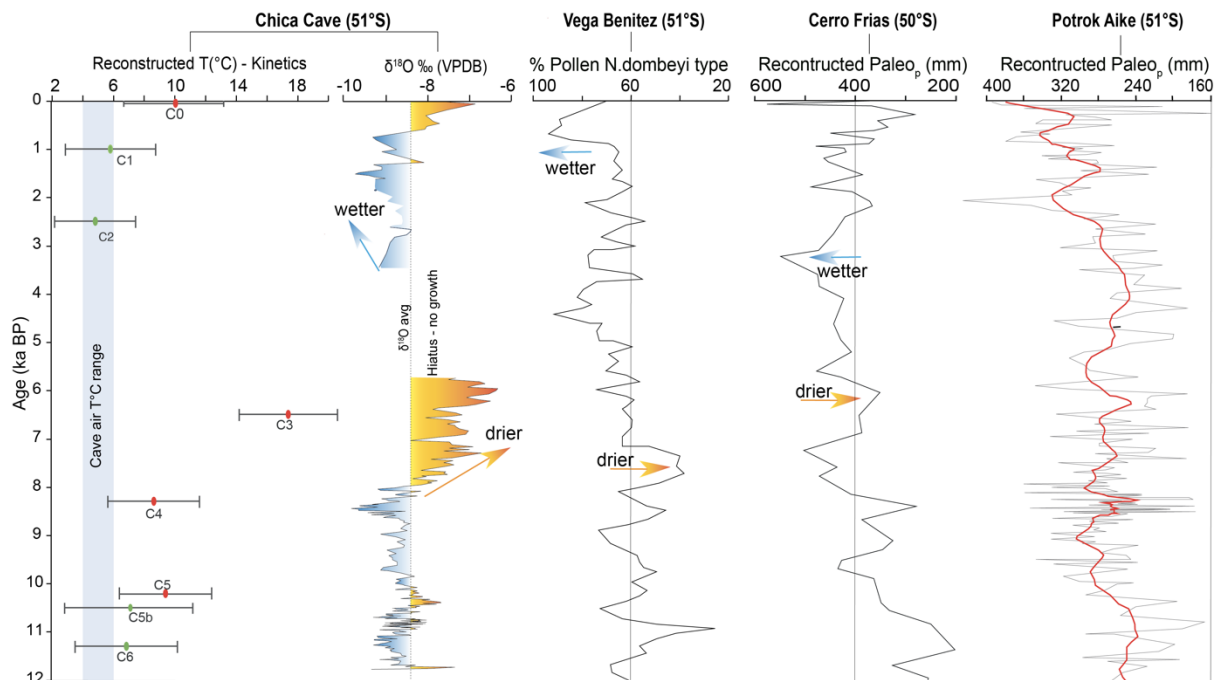
576 Speleothems fluid inclusions (FI) can retain the original isotopic composition on ‘fossil’ dripwater (Affolter
 577 et al., 2014 ; Matthews et al., 2021, van Breukelen et al, 2008; Wassenburg et al., 2021 ; Warken et al. 2022).
 578 When fluid inclusions are well-preserved, FI isotope data ($\delta^{18}\text{O}$; δD) fall onto or near the modern LMWL
 579 while a significantly lower regression slope of FI isotope results would indicate secondary evaporation
 580 (Clark and Fritz, 1999). If the FI_{water} indeed mirrors the original precipitation, the slope would reflect
 581 humidity during and after precipitation. Slope values decrease with humidity (e.g., a slope of 4.5 would
 582 indicate humidity levels of 50%). Fluid inclusion samples from the Chica S6 core reveal some intriguing
 583 complexity. While some samples fall near the LMWL, more than 50% of the samples diverge with a slope
 584 of 1.4 (Fig. 4). This line resembles a secondary evaporation line, which could result from cave ventilation
 585 (Nehme et al. 2020; Warken et al. 2022), but the slope of the line is unusually low for such evaporation
 586 scenario (Clark and Fritz, 1999), which suggests that other processes must be at work, like diagenetic
 587 alteration of fluid inclusion water (Uemura, 2020), or an analytical artefact leading to incomplete water
 588 extraction from the fluid inclusions. (Suppl. S5 table 3, comments).

589
 590 The line of reasoning described here reduces essentially the most trustworthy FI data to samples with the
 591 lowest isotope values in Fig. 4 and closest to LMWL. The trustworthy FI isotope dataset can be therefore
 592 reduced to a single projected dripwater value ($\delta^{18}\text{O}$: -11.8‰ ; δD : -87‰) for the time interval (8.5 to 11.3
 593 ka BP) covered in the data. Since almost all fluid inclusion samples represent a short time interval in the
 594 earliest Holocene, the projected isotope values should be representative of the beginning of the Holocene.
 595 Considering the uncertainties in our projections, we believe values given from FI-3, FI-5 and FI-8b samples
 596 (Suppl., table 3) are not significantly different from the modern-day dripwater values (Fig. 4).

598 5.6. Temperature estimates based on FI and D47

599 Clumped isotope temperatures ($T_{\Delta 47}$) offer the prospect to reconstruct the temperature at which the
 600 speleothem formed, independent of the isotopic composition of the dripwater. Previous work (Daeron et al.
 601 2019; Fiebig et al., 2021) suggests that cave carbonates precipitated near isotopic equilibrium conditions
 602 retain their original $T_{\Delta 47}$ signal. Unfortunately, kinetic isotope fractionation, induced by rapid CO_2 -
 603 degassing or evaporation of dripwater can overprint the original clumped isotope signal and render derived
 604 temperatures unrealistic. This kinetic effect on speleothem is well known and has been studied previously
 605 (Affek et al., 2014; Kluge et al., 2013; Guo and Zhou, 2019, Matthews et al., 2021). Kinetic effects caused
 606 can be evaluated by comparing FI stable isotopes and dripwater isotope data (Fig. 4). Samples close to the
 607 LMWL can be regarded as trustworthy, whereas samples located well below the modern LMWL are likely
 608 to be affected by secondary evaporation. This means that FI-based temperature estimates from samples with
 609 $\delta^{18}\text{O}$ values < -8 permil (Fig. 4) might reflect environmental temperatures, while temperature estimates from
 610 samples with higher $\delta^{18}\text{O}$ values should be discarded.

611 Our samples display $T\Delta_{47}$ values between $4.8\pm 1.3^\circ\text{C}$ and $17.4\pm 1.6^\circ\text{C}$, higher than modern cave air
 612 temperature ($5\pm 1^\circ\text{C}$) (Fig. 7). Four samples (C1, C2, C5b, C6) overlap within error with modern cave air
 613 temperature, but the remaining four (C0, C3, C4, C5) are significantly higher than expected (Fig. 7). Sample
 614 C1 is taken from inside a soda straw that was embedded in the floor after breakage. It is likely that the soda
 615 straw remained filled with water as long as it was actively dripping, therefore constantly replenishing the
 616 DIC pool. At the same time, CO_2 -degassing was likely limited by the barrier of the soda straw walls and a
 617 thick water column inside (Smith, 2021). These conditions support the recording of unaltered clumped
 618 isotope signals and we trust the observed $T\Delta_{47}$ value. Sedimentological evidence suggests that sample C2
 619 precipitated as cave pool carbonate under subaqueous conditions that limit rapid CO_2 -degassing. Under such
 620 conditions the clumped isotope signature is likely to remain unaltered by kinetic fractionation (Kele et al
 621 2015, Breitenbach et al. 2018, Drysdale et al. 2020). These two samples suggest that, starting from at least
 622 2.5 ka, mean temperatures in the Cerro Benitez were close to modern (Fig. 7).
 623



624
 625 **Fig. 7.** The T-Kinetics distribution of the measured clumped isotopes (x axis) and the modern cave air are reported within 2s
 626 error. The values indicated in green are samples with less kinetics and close to the modern cave air, the ones in red are
 627 samples with high kinetics and shifted towards higher temperatures. For comparison, are the Chica cave' $\delta^{18}\text{O}$ signal and the
 628 pollen % from the Vega Benitez peat (McCulloch et al., 2021) as indicator of dryness/wetness trend in the Cerro Benitez. For
 629 regional comparison, the reconstructed paleoprecipitation from pollen record of the Cerro Frias peat (Tonello et al., 2009)
 630 to the North and from pollen record of Potrok Aike Lake in the Argentinian pampa (Shabitz et al., 2013) to the East.
 631

632 Similarly, samples C5b and C6 from the early Holocene were likely deposited from a thicker water film
 633 under wetter conditions and their clumped isotope temperatures are likely valid. While their $T\Delta_{47}$ values
 634 overlap with the modern temperature, the large error margins do not allow us to establish whether the early
 635 Holocene was significantly warmer than today or not. If minimal kinetic fractionation is expected, $T\Delta_{47}$ (max
 636 10°C) might well align with globally higher-than-modern temperatures ($6\pm 3^\circ\text{C}$) between 10.8 and 10.3 ka.
 637 The samples C0, C4 and C5 show temperatures only slightly higher than today and might be realistic,
 638 although we cannot exclude any impact of kinetic fractionation and thus 'to high' reconstructed
 639 temperatures. The remaining sample C3 is unrealistically high and kinetic fractionation due to CO_2 -
 640 degassing and/or evaporation is suspected to affect this sample. A trend towards less negative $\delta^{18}\text{O}$ values
 641 shortly after that sample is noticeable, which might be interpreted as local drying that lead to an increased
 642 kinetic fractionation.

643 To summarize, local temperatures might have been slightly higher in the early Holocene (11-8 ka BP)
 644 compared to today (Fig. 7), although the large error margins do not allow a quantitative estimation of such
 645 warming. Local temperatures were similar to today since at least 2.6 ka BP. A dry interval around 8.5 - 5.5
 646 ka likely affected the cave's hydrology such that kinetic fractionation altered the clumped isotope system,
 647 resulting in too warm apparent reconstructed temperature. The comparison of local reconstructed T($^\circ\text{C}$) from
 648 Chica cave with other semi-quantitative (Pollen) signals from Vega Benitez marsh and reconstructed

649 paleo-precipitation from Cerro Frias (Fig. 7), show a regional similarity in the drying trend during time
 650 interval 8.5 - 5.5 ka and a return to wet conditions from 3.5 to 1 ka. However, the climatic paleo-precipitation
 651 trend shown from Potrok Aiketo the East seems to be out of phase with the Andean records.
 652

653 **6. Conclusions**

654 The Cueva Chica record from the Cerro Benitez (51°S) in one of the rare studied carbonate record located
 655 far South in the Southern Andes. This study provides a first highly-resolved isotopic, elemental and
 656 paleotemperature record spanning the Holocene from the last 12 ka, albeit with a hiatus from 5.8 to 4 ka BP.
 657 The stable isotope and trace elements curves from speleothems show a covariability in the signals. First,
 658 cave monitoring and comparison with regional climatic data suggest that both dripwater and speleothem
 659 $\delta^{18}\text{O}$ values represent a buffered (multi-)annual mean signal, likely biased towards the winter season.
 660 Overall, the oxygen isotope signal reflects changes in amount of precipitation and effective infiltration in the
 661 studied site, and to some extent changing moisture transport trajectories throughout the Holocene. Second,
 662 $\delta^{13}\text{C}$ signal reflects a combined signal of epikarst and in-cave CO_2 degassing, vegetation composition and
 663 density as well as soil activity. Third, the Sr/Ca elemental data shows a moderate sensitiveness to infiltration
 664 changes in the epikarst and behave in concert with the $\delta^{13}\text{C}$ signal, albeit with different intensity to dryness
 665 between both proxies.

666 The multi-proxy analysis of the petrography, growth rate, stable isotopes and elemental data in the Chica
 667 record suggest three main climatic phases throughout the last 12 ka in Southern Patagonia: i) an early
 668 Holocene wet period (with the exception of two dry excursions at 10.5 ka and 8.5 ka BP), ii) a mid-Holocene
 669 dry period and iii), a return to generally wet conditions over the late Holocene. The climate variability from
 670 the Chica record is in phase with other regional peat and lake records located between 50° and 55°S along
 671 the Andes mountain corridor and Western Pacific side, but is out of phase with records located North to 50°
 672 latitude.
 673

674 These tri-phased climate regimes recorded in the Chica cave are the result of global drivers controlling the
 675 strength and shifts of the SWWs throughout the Holocene. The early Holocene was the warmest period in
 676 Southernmost South America, and might be attributable to changes in global ocean circulation which
 677 involves a rise in air temperature and a strength in SWW from 50°S. The response over landmass in Southern
 678 Patagonia is higher precipitation over the Chica cave and the Chilean coast, likely in concert with higher
 679 temperatures. After 9 ka BP, an intensification of ice-shelf instability along the Antarctic Peninsula is
 680 concordant with increasing summer insolation in the Southern hemisphere, which intensified a poleward
 681 shift of the SWWs in response to global warming. This significant change in regional conditions implied a
 682 reduction in moisture supply from the Pacific onto the Patagonian shore and lower total precipitation over
 683 the Chica cave. After 5 ka BP, a different climatic background arises, with a gradual SST decline, consistent
 684 with an equatorward shift of the SWWs in response to a cooling Southern hemisphere, known as the
 685 Neoglacial period. The SWW storm tracks extended to lower latitudes, inducing a return to wetter conditions
 686 with highly variable moisture patterns along the Patagonian coast, as found in the Cueva Chica and other
 687 terrestrial records after 4 ka BP.
 688

689 The reconstructed paleotemperatures based on fluid inclusions and (TA_{47}) from the Cueva Chica
 690 speleothems provide a reasonable estimation of past terrestrial temperature in the Cerro Benitez throughout
 691 the Holocene: local temperatures might have been slightly higher in the early Holocene (11-8 ka BP BP)
 692 compared to today. A dry interval around 8.5 - 5.5 ka BP likely affected the cave's hydrology such that
 693 kinetic fractionation altered the clumped isotope system, resulting in too warm apparent reconstructed
 694 temperature. Since at least 2.6 ka BP, local temperatures were similar to today.
 695

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 704

705 **8. CRediT author statement**

706 **C. Nehme**: conceptualization, writing-original draft preparation, field investigation and sampling, analysis;
 707 **D. Todisco** : conceptualization, field investigation and sampling, reviewing & editing, funding acquisition ;
 708 **S. Breitenbach** : analysis, software validation, writing, review & editing ; **I. Couchoud** : conceptualization,
 709 analysis, writing, review & editing ; **M. Marchegiano** : analysis, validation, writing; **M. Peral** : analysis,
 710 validation, writing ; **H. Vonhof** : analysis, validation, writing ; J. Hellstrom : analysis, **R. Tjallingi** : analysis,
 711 writing ; **Ph. Claeys** : supervision ; **L. Borrero** : review & editing ; **F. Martin** : field investigation, review
 712 & editing, funding acquisition and head of the project.

713

714 9. competing interests statement

715 The authors have no conflict of interests to declare.

716

717 10. References

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